An examination of spatial variation in active-layer freezeback, Illisarvik, Northwest Territories

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Abstract

Investigation into the influence of spatial variations in physiographic variables on active-layer freezeback and 100 cm permafrost temperature was conducted between September 2010 and August 2011 along a hill slope transect near Illisarvik, N.W.T. The results indicate that snow depth, a function of wind and topography, provides the principle control on both the duration of active-layer freezeback and 100 cm temperature. On the hill tops, where snow was thinnest (10 cm ± 10 cm), freezeback was completed in late-November and the mean annual 100 cm temperature was approximately -6.6 °C ± 0.5 °C. At the hill bottom and on the hill slope greater snow depths prolonged freezeback to mid-December and warmed ground temperatures to -4.7 °C ± 0.2 °C. These results are of direct applicability as the active layer can only bear considerable weight, such as that of machinery, following freezeback.
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Section 1  Overview and Objectives

1.1 Introduction

This paper intends to investigate the influence of physiographic characteristics on the duration of active-layer freezeback along a 550 m transect in the western Canadian Arctic. Investigations of the response of active-layer characteristics to climate change are of interest to scientists and engineers as the mechanical properties of permafrost and the active layer are correlated with its physical-state and temperature. Information presented here may therefore have further implications for design and construction of infrastructure in this region.

1.2 Climate and Permafrost

Permafrost is a thermal condition of the ground associated with cold climates (Smith and Riseborough 2002). Commonly defined as ground which remains at or below 0°C for two or more consecutive years, permafrost is endemic in many high latitude and altitude environments (Williams and Smith 1989). The aerial extent to which perennially frozen ground occurs has been used to identify latitudinal zones in which permafrost is spatially continuous, widespread, sporadic, or isolated in occurrence (Smith and Riseborough 2002). Early classifications were made on the basis of air temperature (Brown 1960). However, continuous, widespread, sporadic, and isolated have since come to refer to regions wherein greater than 90%, 50-90%, 10-50%, and less than 10% of the terrestrial surface is underlain by permafrost, respectively.

While climate determines the presence or absence of permafrost at the continental scale, local variations in vegetation, topography, snow cover, and soil characteristics yield ground temperature variations on the order of several degrees over smaller distances (Smith and
Riseborough 2002). The effect of these factors may be identified by comparison of the mean annual air temperature (MAAT), mean annual ground surface temperature (MAGST), and mean annual temperature at the top of permafrost (Hereafter $T_{top}$). Generally, the combined influence of vegetation and snow cover are manifested by a MAGST higher than the MAAT – an observation termed the surface offset (Henry and Smith 2001). Differential thermal conductivities in the active layer between the frozen and thawed states result in a $T_{top}$ lower than the MAGST – a phenomenon termed the thermal offset (Burn and Smith 1988). Smith and Riseborough (2002) suggest that while the effects of snow cover dominate spatial variation in permafrost in the continuous zone, the thermal offset is the principle control on permafrost occurrence in the discontinuous region. The thermal and surface offsets will be discussed in more detail below. Vertically, permafrost may range from decimeters to hectometers thick with depth being a function of local climate, thermal properties, and the duration of conditions conducive to permafrost aggradation.

Energy flow into and out of the ground surface may be simply expressed as:

$$Q_g = Q^* - Q_l - Q_H$$

Where $Q_g$ refers to the ground heat flux, $Q^*$ to the net radiation arriving at the surface, $Q_l$ to the energy absorbed or released by latent heat processes, and $Q_H$ to the energy associated with sensible heat changes (Williams and Smith 1989). Furthermore, $Q^*$ is fundamentally dependent upon the net radiation balance and may therefore be expanded to:

$$Q^* = Q_{S\downarrow} - Q_{S\uparrow} + Q_{L\downarrow} - Q_{L\uparrow}$$

Here, the subscripts S and L refer to shortwave and long wave radiation, respectively, while ↓ and ↑ correspond to the incoming and outgoing components. Indeed, as suggested above, it is
slight variations in the components of these equations which result in spatial and temporal variation in permafrost characteristics.

The active-layer is the upper-most layer of the ground, subject to summer thaw and subsequent winter freezing (Williams and Smith 1989). Hinkel and Nelson (2003; 1) note active-layer thickness is influenced by snow cover, the height and thermal properties of vegetation, summer precipitation, the thickness of the organic layer, and topography. Given the influence of these variables on the energy balance at the ground surface it is likely that they also affect other properties of the active layer including the duration of freezeback and thaw.

1.3 Active-Layer Freezeback and the Zero-Curtain Effect

Carey and Woo (1998) suggest that the one-dimensional energy balance for the active layer may be expressed as:

\[ Q_g = Q_p + Q_s + Q_l \]

where \( Q_g \) refers to the heat available at the ground surface, \( Q_p \) to the heat flux into or out of the permafrost, \( Q_s \) to the sensible heat associated with active-layer temperature, and \( Q_l \) to the latent energy component. Release of latent heat associated with the freezing of pore water results in the maintenance of isothermal temperatures at or just below zero-degrees centigrade within the freezing active-layer – for the duration of freezeback \( Q_g \) may be approximately equal to \( Q_l \). This period of isothermal conditions within the active layer, commonly termed the zero-curtain, is both a physical boundary preventing cooling in the underlying permafrost and a length of time (Outcalt et al. 1990). The zero curtain establishes in the active layer about the phase-equilibrium temperature and decouples the permafrost from the atmosphere for its duration (Osterkamp and Romanovsky 1997). Following the closure of the zero curtain temperature at the top of
permafrost is permitted to decline at a rate governed by the thermal conductivity of the snow pack and the frozen active layer and the prevailing thermal gradient between the atmosphere and permafrost (Burn and Zhang 2009). The closure of the zero curtain, expressed by the decline in temperature at the top of permafrost, denotes the completion of active-layer freezeback and the onset of the cooling period (Burn and Zhang 2009; Karunaratne 2011). Given that the zero curtain is associated with latent heat released from water during soil freezing its development in dry material is not possible (Outcalt et al. 1990).

1.4 Study Site and Research Program

Permafrost research along the southern Beaufort coast was pioneered by Dr. J. Ross Mackay who, since the 1950s, has completed many of the foundational surveys and geographical studies from the region. Interest in the western Arctic coast was spurred by the discovery of oil and natural gas reserves in the 1960s, 1970s and 1980s.

In August 1978 Mackay initiated the drainage of Lake Illisarvik on Richards Island (Figure 1). The purpose of this field experiment was to observe the aggradation of permafrost into the talik underlying the lake. Active-layer thickness in the lake bottom has been recorded annually since 1979 yielding the longest known record in the western Canadian Arctic. In 1983 a tundra active-layer course was developed to provide a basis for comparison with the measurements from the lake bottom (Mackay and Burn 2002). The duration of measurements along the tundra course made it an ideal location for further study of active-layer phenomena.

In summer 2010, eleven sites were selected and instrumented parallel to the Illisarvik tundra transect with the goal of determining spatial variation in active-layer freezeback. These sites were located within three topographic units. Within each unit sites were additionally selected to
represent the different vegetation zones. While the tundra transect is broadly characterized by earth hummocks and low ground vegetation, willows and alders have established in areas of deep winter snow (Mackay and Burn 2002). The surface characteristics of this transect are representative of other tundra regions in the western Arctic.

Following selection, each site was equipped with temperature sensors below the ground to monitor permafrost temperatures and above the ground to determine snow pack development. During this time soil and permafrost samples were obtained to analyze water content, ice content, organic material, and soil texture. A description of the field and laboratory investigations is provided in Table 1.

1.5 Research Hypotheses and Objectives

This project seeks to develop a better understanding of the behaviour of active-layer freezeback and upper permafrost temperature in response to spatial variation in characteristics of the ground surface and active layer. These properties include snow characteristics, topography, vegetation, soil moisture, ice content, soil texture, and organic content. The hypothesis of this paper is that these characteristics act to control the duration of active-layer freezeback, resulting in spatial variability. The objectives are to:

1. Identify the spatial variability in the temperature at the top of permafrost as well as the physiographic factors mentioned above;
2. Determine the spatial variability in active-layer freezeback along a 550 m transect;
3. Analyse the influence of the physiographic variables on the duration of freezeback.
1.6 Paper Structure

This dissertation is further separated into four sections. Section 2 provides the physiographic background for the study site and briefly introduces the regional setting, geologic and glacio-marine history, and contemporary climate of Illisarvik and Richards Island. Section 3 examines the relation between air temperatures and permafrost conditions through consideration of the surface and thermal offsets. This section additionally presents the results of physiographic investigations along the tundra active-layer course and an analysis of spatial variations in the thermal and physical variables. Section 4 communicates the principles and results of the active-layer freezeback measurements including an analysis of spatial variation. Finally, conclusions from this experiment are highlighted in Section 5.

Section 2 Physiographic Setting

2.1 Regional Setting

The active-layer freezeback transect is located on Richards Island, Northwest Territories, approximately 60 km west of Tuktoyaktuk. This locale lies on the boundary between the Mackenzie delta and Beaufort Sea (Murton 2009). Richards Island is approximately 2200 km² and is bordered to the north by Kugmallit and Mackenzie bays and by the Middle, Harry, and East channels of the Mackenzie River to the east, south, and west (Burn 2002:1282). The region lies within the marine tundra climate zone and is characterized by long cold winters and short cool summers. The mean annual air temperature during the climate normal period (1970-2000) was -10.5 °C at Tuktoyaktuk, the nearest meteorological station.
Positioned within the zone of continuous permafrost, Richards Island consists of ice rich, unconsolidated Quaternary sediments (Hill et al. 1990). Burn and Kokelj (2009) have observed near surface ground temperatures ranging between -6 °C and -9 °C along a south-north gradient on Richards Island. Local permafrost thickness varies considerably due to partial glaciation during the end of the Wisconsin glacial period (Murton 2009). In unglaciated portions of Richards Island perennially frozen ground may reach up to 750 m thick (Allen et al. 1988; Taylor et al. 1996). The study area lies within the northwest limit of the Laurentide Ice Sheet.

2.2 Quaternary Geology and Paleoenvironmental History

Much of the southern extent of Richards Island was glaciated for a short period between approximately 22 and 16 ka ago (Figure 2; Murton et al. 2007). Regions within the ice limit are largely blanketed by moraine deposits (termed the Toker Point Member) (Figure 3; Hill et al. 1990; Murton 2009). These deposits are characteristically till diamictions approximately 8 m thick. Underlying units include the Kittigazuit formation, the Kidluit formation, Hooper clay and Kendall sediments and are described in detail by Murton (2009).

During the previous interglacial period (Sangamonian), the Richards Island area was below sea level. The area was subsequently exposed by marine regression to conditions conducive to permafrost aggradation (Murton 2009). During the Wisconsin glacial period Richards Island was briefly covered by the Mackenzie paleo-ice stream, the northwest margin of the Laurentide Ice Sheet. This advance, culminating in the Toker Point stade, occurred between 22 000 and 16 000 yr BP and has been traced using the distribution of moraine deposits and other ice contact features (Rampton 1988b; Murton 2009). Some uncertainty surrounds this boundary as
suggested in the Tuktoyaktuk phase limit proposed by Mackay et al. (1972) as an alternative late phase Wisconsin boundary.

Deglaciation on Richards Island had likely commenced by 14 300 cal. Yr. BP (Murton et al. 2007). This shift was expressed by meltwater activity resulting in fluvial valley formation and infilling by glaciofluvial sediment (Murton 2009). During deglaciation thermal contraction of the ground occasioned the establishment of ice wedges in the permafrost aggrading north of the receding ice sheet. Concurrent injections of subglacial melt water through the permafrost lead to the formation of massive ice bodies of glacial origin but devoid of ice-contact features (Mackay and Dallimore 1992; Murton 2005).

The early-Holocene period is estimated to be approximately 3 °C warmer than present (Murton 2009). During this period the black spruce forest limit extended to the modern coast. Evidence of this early Holocene climate optimum is recorded in the western Arctic by a thaw unconformity reflecting the thicker active layer associated with warmer summer temperatures. This cryostratigraphic record has been documented by Burn (1997). Burn (1997) suggests that observed cooling following the hypsithermal may be partially attributed to decreasing continentally attributed to marine transgression during this period. Murton (2009) indicates that the sea level rose from 70 m – 90 m below the present level corresponding to a 100 km southward movement in the shoreline (Burn 1997).

2.3 Contemporary climate and the Period of Study

The study area on Richards Island lies within the marine tundra climatological zone (Figure 4; Burns 1973). This region is characterized by long, cold winters and short, cool summers. Approximately 170 mm of precipitation falls annually of which two thirds is deposited as
snowfall during the winter (Figure 5). During periods of sea ice cover the region is dominated by cold arctic air masses. These conditions typically prevail between October and June on the Beaufort Sea (Hill et al. 1990). Conversely, during the ice free season, maritime conditions are permitted to develop (Hill et al. 1990). Summer air temperatures are typically stable compared with the winter temperatures. Figure 5 shows a large standard deviation in the winter months due to fluctuating ice and atmospheric conditions.

The study period under consideration was the hydrological year between 2010 and 2011 (September 1st 2010 thru August 31st 2011). Historically, the onset of the freezing period ranged from September 14th to October 6th between 1974 and 2012. The average date during this time was September 28th. In 2011 the onset of freezing was approximately September 23rd, slightly earlier than average. The mean monthly temperatures for the period of study are provided in Figure 6. These values, compared with the mean monthly temperatures during the climate normal period, suggest that the study year was slightly warmer than average but only deviated significantly in November. Mean daily temperatures at Illisarvik, approximately 500 m distance from the tundra transect, are shown for the study period in Figure 7. Reconstruction of missing data at the end of August was permitted given the strong correlation between Illisarvik and Tuktoyaktuk air temperatures (Figure 8).

**Section 3  Physical Site Characteristics and Ground Temperature**

**3.1 Introduction and Site Selection**

In addition to climatic variables, natural factors which influence the near-surface ground thermal regime include vegetation, snow cover, soil composition, water and ice content, and the
hydrological regime of the active-layer and uppermost permafrost (Romanovsky and Osterkamp 1995). To examine the degree to which these factors influence permafrost temperatures and the duration of active-layer freezeback eleven sites were selected for instrumentation based on local variations in topography and vegetation along the Illisarvik tundra active-layer course. The number of sites allowed for redundancy as some of the sites failed to produce results due to instrument error. The primary criterion for site selection was topographic position as this factor controls regional snow depth (Mackay and Burn 2002). These sites aim to represent three distinct topographic units. The three units identified here were: (1) the hill top, (2) the hill slope; and (3) the hill bottom. Sites 11, 10a, 10b, 4, and 3 belong to the hill top unit; 8a and 8b are part of the hill slope unit; and sites 6a, 6b, 5a, and 5b are representative of the hill bottom unit. The location of these units is presented in Figure 9. Images of a selection of sites from each unit are given in Figure 10.

This section presents: (1) the connection between the atmosphere and permafrost in the context of the surface and thermal offsets; and (2) the results of field measurements made from September 2010 to August 2011. Examination of this data will provide insight into the spatial variation in physiographic characteristics as well as near surface permafrost temperatures. The parameters identified here are discussed with particular reference to active-layer freezeback in Section 4.

3.2 Instrumentation

At each site, data loggers were deployed in conjunction with a temperature probe to a depth of 100 cm. In addition, each site was equipped with a series of miniature temperature loggers at 5, 10, 20, 30, 40, and 50 cm above the ground surface, attached to a one meter length of steel rebar
protruding 50 cm from the ground surface. 50 cm is an appropriate height given the historical snow record provided by Mackay and Burn (2002). The iButtons - oriented northward to reduce the potential for melt pitting - permitted the determination of snow accumulation during the study period (Lewkowicz 2008). Soil samples were collected at continuous 10 cm intervals from the top 110 cm of ground using a CRREL drill. These samples were utilized in analysis of gravimetric moisture content, volumetric ice content, organic content, and soil texture. Additionally, samples were collected from the active layer to determine volumetric water content. Ground temperature measurements were obtained at 100 cm depth every four hours for the period between September 2010 and August 2011. This depth was selected as all temperature sensors would be placed within the top of permafrost and below the depth of influence attributed to the hummock microtopography (after Mackay and Mackay 1974). An example of instrumentation at each site is given in Figure 11.

3.3 Introduction to the Surface and Thermal Offsets

Permafrost is linked to the atmosphere by the active layer, snow cover, and vegetation all of which vary over space and time (Romanovsky and Osterkamp 1995). The effects of snow cover and vegetation combine to produce the surface offset. This is expressed by the difference between air and ground surface temperatures. A thermal offset is also observed between ground surface temperatures and the temperature at the top of permafrost. This phenomenon is attributed to temperature dependent thermal conductivities. Temperature at the top of permafrost is therefore influenced by both the surface and thermal offsets (Smith and Riseborough 2002). A simplified schematic portraying these phenomena is presented in Figure 12.
3.4 The Surface Offset

First proposed by Lachenbruch et al. (1988), the surface offset is the combined effect of snow, vegetation, and the surface organic layer resulting in ameliorated temperature at the ground surface (Luthin and Guymon 1974). Of these factors, Smith and Riseborough (2002) suggest that snow cover is of the greatest influence. As such, they have also termed this phenomenon the nival offset. The low thermal conductivity of snow limits the amount of heat lost from the ground. This is particularly significant as the period of snow cover commonly coincides with the coldest atmospheric temperatures (Smith and Riseborough 2002). Vegetation acts to temper the relationship between air and ground temperature during the summer and winter. During the summer vegetation acts to shade the ground surface while in the winter it traps snow accelerating accumulation during the freezeback period. The result of the surface offset is that air temperatures are more closely related to ground temperatures in the summer than in winter.

3.4.1 Snow Regime

The winter snow regime is often cited as the principal source of difference between air and ground temperatures (Smith and Riseborough 2002). While thin layers may result in cooling due to augmentation of the surface albedo, the net effect of snow is to warm the ground (Gold 1967; Zhang 2005). This warming is attributable to the poor thermal conductivity of the snow pack which isolates the ground from the atmosphere. This results in a weak correlation between air temperature and ground surface temperatures during the snow period (Zhang 2005). Mackay and Mackay (1974) have observed that the effectiveness of the snowpack at ameliorating ground temperatures undergoes logarithmic decrease with increasing depth. This suggests that a critical depth exists after which additional snow has a negligible impact on the correlation between air and ground temperatures. Harris (1981) has identified this depth as 50 cm.
The efficiency of the snow pack as an insulator depends on density, depth, structural characteristics, and its duration (Zhang 2005). Goodrich (1982) notes early, thick snow covers are the most efficient at retarding latent heat loss and are thereby correlated with elongated freezeback periods. Locally, snow regimes are impacted by local topography and vegetation as they relate to wind strength and direction. Walker et al. (1999) have demonstrated that augmentation to the annual snowpack using snow fences resulted in higher ground and surface temperatures and decreased spatial and temporal temperature variation. Goodrich (1982) notes that ground temperatures share a close correlation with maximum snow depth as well as snow conditions early in the freezing season.

### 3.4.2 Snow Characteristics

Snow accumulation was monitored continuously during the 2010-2011 freezing period using miniature temperature loggers (Thermochron iButtons – Model DS1921G, Dallas Semiconductor Corporation, USA). The application of these thermistors was initially proposed by Lewkowicz (2008). In this instance, the methodologies outlined by Lewkowicz (2008) were modified to reflect the lesser snowfall in the western arctic, compared with mountainous sites in northwestern Canada. This method was selected as a simple, inexpensive technique to remotely monitor snowpack evolution and decay. In addition to the iButton array, direct measurements were made in April 2011 and are presented along with the conclusions from the data logger array in Table 2. While site-by-site comparison suggests that these methods yielded different results both procedures indicate a similar pattern regarding the distribution of snow across the transect.

Snow depth observations varied considerably along the length of the tundra active-layer course. As indicated in Figure 13, Figure 14, and Figure 15, hill top sites generally had a much thinner maximum snow depth than either the hill slope or hill bottom zones. Maximum values
ranged between 5 cm and 20 cm at the tops of hills rising to 30 cm to 50 cm in the hill bottoms and slopes. While snow depth was closely correlated with $T_{\text{top}}$ it shared a poor correlation with vegetation highlighting the importance of topography in this region (Figure 16). This is consistent with the observation by Mackay and Mackay (1974) that the spatial distribution of snow across the landscape is fundamentally controlled by topography. Mackay and Mackay (1974) further note the disproportionate accumulation of snow in depressions on Garry Island, N.W.T., a tundra site 50 km to the west. In 2011 the snow pack generally disappeared by the 17th of March on the hill tops but lingered until the 20th – 26th on the hill slopes and at the hill bottom. Given that no site had a snow sensor at 0 cm the duration of the snow pack may be an underestimate.

### 3.4.3 Vegetation

The influence of vegetation on ground temperature depends on the season under consideration. During the snow free period vegetation reduces the amount of energy reaching the ground surface due to shading and the obstruction of solar radiation (Dingman and Kouts 1974). Vegetation additionally modifies evaporation regimes by changing the active surface and apparent surface roughness. During the winter, surface vegetation augments snow accumulation due to trapping and modification to local wind patterns. Furthermore, leaf and debris accumulation augment the surface organic layer which influences the surface offset through seasonal differences in thermal conductivity and specific heat capacity. Studies from the continuous permafrost zone have shown that the removal of the surface organic layer results in an increase in temperature at the ground surface (Smith 1975; Nicholas and Hinkel 1996). Generally, the most significant effect of vegetation is its role in snow trapping.
3.4.4 Vegetation Characteristics

Surveys were conducted at each site to determine the characteristics of surface vegetation. Of particular interest was the effect of vegetation on snow depth and the radiation balance. These surveys aimed to specifically identify the species composition and vegetation height at each site. The results of these surveys are presented in Table 3 and Table 4. While trends in vegetation type are not clearly discernible across the transect vegetation height was greater at the hill bottom and hill slope sites than at the hill tops. Mean vegetation heights ranged from 15 to 57 cm and generally shared a poor correlation with other measured parameters. Interestingly, the presence of alders at site 8b resulted in a disproportionately thinner active layer and the warmest ground temperatures observed along the transect. This effect was not observed at sites with tall willows.

3.5 The Thermal Offset

The thermal offset refers to differences between $T_{\text{top}}$ and MAGST and is often attributed to seasonal variation in the thermal properties of the active layer due to differential thermal conductivities in frozen and unfrozen ground. Frozen ground typically has a higher thermal conductivity than unfrozen material owing to the fact that ice is approximately four-times more conductive than water (Williams and Smith 1989). While the thermal offset is most pronounced in moist, organic rich soils it is also a characteristic of block fields and talus material (Juliussen and Humlum 2008). The thermal offset commonly manifests as a negative thermal anomaly defined by a mean $T_{\text{top}}$ which is cooler than the mean ground surface temperature (Burn and Smith 1988; Smith and Riseborough 2002). Romanovsky and Osterkamp (1995) have observed that the magnitude of this thermal offset is dependent upon the thermal properties of the active
layer. Goodrich (1982: 430) has observed thermal offsets ranging up to 4 °C in soils with temperature-dependent thermal conductivities. Carey and Woo (1998) further note that the thermal conductivity and heat capacity of the active layer can change within the thawed material as water is replaced by air due to evaporation and drainage loss.

### 3.5.1 Water Content and Near Surface Ground Ice

Soil moisture influences the ground surface energy regime by modifying the evaporation regime during the summer period. Heat exchange with the surface is further altered by the influence that soil water has on thermal conductivity and specific heat capacity – both of which impact the thermal diffusivity (Williams and Smith 1989). During the freezeback period, soil moisture provides a significant source of latent heat.

Volumetric water content (g/cm³) was measured throughout the active layer at 15 cm intervals in July 2010. From this, the total moisture content of the active layer was calculated by taking the average volumetric measurement and multiplying it by the depth of the active layer. The results of these calculations are presented in Table 5. In addition, gravimetric water content and volumetric (excess) ice content was determined at each site at 10 cm intervals down to 110 cm (Figure 17). High values of volumetric ice typically occur below the active layer. This is consistent with observations of the transient layer proposed by Shur et al. (2005). Spatial patterns in these parameters are difficult to discern and as such no conclusions are made in this regard. In some instance excess ice is reported within the active layer due to the incomplete development at the time of sampling.
3.5.2 Organic Content

The significance of soil organic material in active-layer freezeback is principally derived from its impact on soil moisture. Organic content was determined at each site by loss-on-ignition (Figure 17). A single subsample was selected from a homogenized mixture of soil at 10 cm intervals. The depth of sampling was intended to extend to 70 cm to fully account for the active layer at each site. At some sites sampling was less extensive due to insufficient material. Sites were generally characterized by two peaks in organic content: one at the ground surface and the other at the bottom of the active-layer. Given the local hummocky topography, the observation of organic material at the bottom of the active layer is attributed to cryoturbation (Mackay 1980; Bockheim and Tarnocai 1998). Site 6b is located in peat material while the other sites are predominantly mineral in soil composition. Field observations suggest that the surface organic layer thickens towards the bottom of the hill and is generally greatest in areas of dense vegetation.

Given that only one sample was taken at each depth, a short analysis of intersample variability was conducted to determine the representativeness a single subsample. 11 subsamples were taken from 3 samples of differing organic content resulting in 33 samples for investigation. The results, presented in Table 6, indicate standard deviations ranging between 0.3 % and 0.7 % with greater deviations occurring at progressively higher organic content. These results permit the conclusion that a subsample from a homogenized sample may be taken as representative of the whole.

3.5.3 Mineral Soil

 Soil samples obtained at each site from various depths within the active layer were analyzed using the pipette method in order to determine the textural composition of the mineral
These investigations suggested that along the tundra transect the mineral soil was ubiquitously sandy. In all cases, samples were classified as either a loamy sand, sandy loam, or sandy clay loam under the USDA classification framework. No spatial trends were distinguished with regards to soil texture.

### 3.6 The Active Layer

The active-layer is defined as the near surface region of soil or organic material subject to annual freezing and thawing. This region is influenced by differential thermal conductivity, water content and state, and meteorological events (Romanovsky and Osterkamp 1995). Generally, active-layer thickness varies in response to air temperature, snow cover, summer rainfall, soil characteristics, and vegetation (Nelson et al. 1998). The duration of active-layer development is fundamentally constrained by air temperatures and the timing of snow melt. Provided temperatures above zero degrees, thawing is generally assumed to commence following the disappearance of snow from the ground surface.

Active-layer depth was monitored along the tundra transect using a graduated steel rod inserted into the soil until the point of refusal. The hill top and hill slope active-layer is characterized by a mineral section overlain by a thin layer of organic material. However, at the hill bottom (site 6b) the active-layer is almost entirely organic in composition. Active-layer thickness ranged between 23 cm and 54 cm in 2010 and 31 cm and 62 cm in 2011 (Table 7). In both years the thinnest active-layer was observed at site 8b and attributed to the shading effect of alders in local occurrence. The greatest thaw depths were observed at site 10a in the hill top region and credited to a thin organic layer and positioning atop a hummock. No clear pattern in active-layer thickness was observed between the three spatial units or with respect to vegetation.
characteristics. Deeper active-layer depths observed in 2011 were attributed to slightly warmer summer temperatures. However, a slight active-layer thinning was observed at site 6b in 2011, perhaps due to peat soil which is less affected by changes in air temperature (Karunaratne 2011).

3.7 Temperature at the Top of Permafrost

Ground temperature was measured at the top of permafrost ($T_{\text{top}}$) at all sites (Table 8). 100 cm was used uniformly to represent $T_{\text{top}}$. In some cases this depth was considerably deeper than the active-layer thaw and consequently reported values may not truly reflect conditions at the top of permafrost. Earlier studies from the region have indicated that a depth of 100 cm should place the temperature sensor below the depth of influence from microscale topography. Two paired sites (10a and 10b), located less than 50 cm from each other, indicate that the observations are not below the effect of the hummocky topography. Site 10a was installed in the top of a hummock while 10b was located in the adjacent trough. 10a, with significantly less snow cover than 10b, recorded a minimum of $-15.8\,^{\circ}\text{C}$ compared with $-14.7\,^{\circ}\text{C}$ at 10b.

At all sites $T_{\text{top}}$ varied during the year, typically reaching a minimum in early to mid-March. Minima were recorded earlier at hill top sites than on the hill slope or at the hill bottom. Mean annual temperatures were lowest at the hill top (ranging between $-6.1\,^{\circ}\text{C}$ and $-7.1\,^{\circ}\text{C}$). Similar mean annual temperatures were observed at hill slope and hill bottom sites with all $T_{\text{top}}$ values ranging between $-4.5\,^{\circ}\text{C}$ and $-4.9\,^{\circ}\text{C}$. As indicated by Walker et al. (1999) spatial variation in ground temperature is reduced in areas of thick snow cover. The lowest minimum temperatures were observed at the hill top sites likely due to limited snow cover. Hill top minima averaged 5 $^{\circ}\text{C}$ cooler than the hill slope or hill bottom sites. This observation is consistent with those presented by Mackay and Burn (2002) who have continuously monitored minimum temperatures.
from this transect since 1983. These results are also congruous with the relative location of thermal minima on other hill slope transects in the western Arctic (Burn and Zhang 2009). In these instances ground cooling was delayed and restricted by snow cover (Mackay and Burn 2002; Burn and Zhang 2009). Measured temperature profiles decreased most rapidly following freezeback. The rate of temperature change decreased following freezeback largely due to the impact of an ameliorated thermal gradient and snow pack development.

Section 4  Active-Layer Freezeback: Regional Relationships

4.1 Features of the freezing process within the active layer

Outcalt et al. (1990) note that before the freezing front can penetrate into the ground ice nuclei must form at the surface. This necessitates the super cooling of water below the frost point. Following ice nucleation, crystal development proceeds drawing moisture towards the freezing front due to cryosuction (Williams and Smith 1989). In regions with two-sided active-layer freezeback a second freezing front penetrates upwards from the top of permafrost. Two-sided freezeback results in ice segregation at the bottom and top of the active layer as well as desiccation and consolidation of the closing unfrozen region (Mackay 1980; Outcalt et al. 1990). Outcalt et al. (1990: 1514) concluded that upward freezing was due to evaporative cooling in response to a vapour pressure gradient directed across the zero-curtain and towards the ground surface.
4.2 Principles, Methodology, and Sources of Error

$T_{\text{top}}$ will remain stationary during the freezeback period due to the release of latent heat of fusion associated with the freezing active layer. This phenomenon, observable in annual temperature records from moist soils, allows for the estimation of the timing of zero curtain closure. Dates for the onset of the freezing period were taken to be the time at which the temperature in the atmosphere was consistently below 0 °C. The completion of freezeback was determined by the date at which the $T_{\text{top}}$ decreased following the passing of the zero curtain. This method has been used successfully by Burn and Zhang (2009) to ascertain the duration of active-layer freezeback along a 750-m hill slope transect at Qikiqtaruq, Yukon Territory, Canada. Romanovsky and Osterkamp (1995) suggest that in Northern Alaska freezeback dates depend upon air temperature, active-layer thickness, soil water content, and the timing and variation in snow cover.

Osterkamp and Romanovsky (1997) observed upward freezing of the active layer beginning approximately two weeks prior to the onset of the freezing season on the coastal plain of Alaska. Several days following the establishment of the downward freezing front, Osterkamp and Romanovsky (1997) observed the establishment of the ‘zero curtain’. This suggests a time lag between freezing conditions and isothermal conditions in the active layer. These observations are significant as they suggest that active-layer freezeback may start earlier and be of shorter duration than inferred from $T_{\text{top}}$ and air temperatures. Therefore, the method utilized here to determine the length of the freezeback period may be a slight underestimate.
4.3 Freezeback Observations

This report is primarily concerned with the freezeback period, which extends between from the onset of continuous freezing temperatures at the ground surface in late September until early December. Air temperature data indicate that the 2010-2011 freezing period began on the 23rd of September. Following the onset of freezing conditions, the zero-curtain period is easily identified by the relatively constant temperatures shown in Figure 18. The completion of active-layer freezeback is signified by the decline in temperature at the top of permafrost.

The mean date for the completion of active layer freezeback at hill top sites was between 22 November and 6 December (Table 9). These sites were followed by those at the hill slope and hill bottom which completed freezeback between 13 and 16 December and 13 and 15 December, respectively. The average duration of the freezeback period was 75 days. Freezeback duration ranged between 65 and 84 days at sites 4 and 8b, respectively. Additionally, the ratio of the duration of active-layer freezeback to the total length of the freezing season averaged 32%. This proportion is significant as it quantifies the amount of the freezing season occupied by the zero curtain. The values presented here suggest a spatial variation in the duration of active layer freeze back along the 550 m tundra transect of approximately 19 days in the 2010/2011 winter period.

Freezeback periods were approximately two-weeks longer in areas of deep snow than in those with thin cover. Correlation analysis was conducted to compare active-layer freezeback with the maximum annual snow depth, maximum snow depth during the freeze back period, and average snow depth during the freeze back period (Figure 19). In these cases comparison yielded a positive relationship between freezeback duration and the independent variable. The coefficient of determination was strongest between active-layer freezeback and maximum snow
depth. Comparison with other parameters including active-layer depth, mean vegetation height, and active-layer moisture yielded inconclusive results. A strong correlation between freezeback duration and $T_{\text{top}}$ is not surprising given that $T_{\text{top}}$ was the parameter used to measure the duration of freezing.

The conclusion that snow depth is the predominant control on active-layer freezeback along the tundra transect is consistent with conclusions drawn from elsewhere in the western arctic. Burn and Zhang (2009) note on Qikiqtaruk (Herschel Island), sites where snow accumulates in early winter are consistently the last to freezeback. Romanovsky and Osterkamp (1995) have additionally identified that a very thin snow cover often results in an accelerated active layer freezeback due to the effects of albedo on the radiation balance.

### 4.4 Significance

The results presented here are of direct applicability as geotechnical instability associated with the unfrozen active layer limits the period during which heavy machinery may be operated safely on the western Arctic tundra. This is of particular importance in the Mackenzie Valley region due to current interest in natural gas development. Furthermore, the Western Arctic has experienced accelerated climate warming over the past four decades; this warming has been particularly evident during the winter months (Sturm et al. 2001; Burn and Zhang 2009). Active-layer freezeback shares an identified correlation with increasing ground temperatures and snow depths, both of which are anticipated to increase under climate change scenarios. This provides problematic implications as an extended freezeback duration would shorten the available work period in winter.
Section 5 Conclusion and Summary

This paper has examined the spatial variation in the duration of active-layer freezeback and temperature at the top of permafrost along a 550 m tundra transect near Illisarvik, NWT. Temperature and physiographic data, collected from eleven instrumented sites, permitted additional investigations into the spatial variability of snow depth, vegetation, moisture content, ground ice content, organic content, and active-layer thickness. The results of these investigations, presented in section 4, indicate that while snow depth shares a close correlation with topography and $T_{\text{top}}$ the recognition of spatial trends within other variables is more problematic. In Section 5, the significant influence of the snow pack in retarding heat loss from the freezing active layer was identified in the strong association between the duration of active-layer freezeback and snow depth. While snow is likely not the only variable influencing the period of freezeback it appears to be the most significant across small spatial scales.

The following conclusions can be extracted from these investigations of active-layer freezeback:

1. Spatial variation in snow depth is principally influenced by topography along tundra transect at Illisarvik. The wind redistributed snow pack is thickest at the hill bottom and on the hill slope and thinnest on the hill top. Remotely measured snow depths ranged from 5 cm to 50 cm.

2. Lateral variations in other physical parameters including vegetation, moisture content, ground ice content, organic content, and active-layer thickness were not clearly identified.

3. Snow depth is the dominant control on $T_{\text{top}}$ and the duration of active-layer freezeback along the transect.
4. The range between mean annual temperatures at a depth of 100 cm ($T_{\text{top}}$) was between -4.5 °C and -7.1 °C. Ground temperatures were on average 1.5 °C cooler at the hill top sites than either the hill bottom or hill slope sites due to the thinner snow pack. The influence of snow is further highlighted by the temperature minima which were approximately 5 °C cooler at the hill top sites.

5. The duration of active-layer freezeback varied spatially by 19 days. At hill the wind-blown hill top sites, freezeback was typically completed by late-November. In areas where snow drifts accumulated freezing was often delayed until mid-December.
Section 6  Acknowledgements

This research has been supported by the Royal Canadian Geographic Society, the Northern Scientific Training Program (Indian and Northern Affairs Canada), the Natural Sciences and Engineering Research Council of Canada, the Polar Continental Shelf Project, Carleton University, and the Aurora Research Institute. I am indebted to Dr. C.R. Burn who has provided me with a matchless introduction to the conduct of permafrost research and to the western Canadian Arctic region. In addition, Dr. Burn has been the source much advice and example throughout this project both in the field and during writing. Many thanks go to my friends who, all together, have provided a constant source of joy, inspiration, and excuses during my long-tenured undergraduate program. Special thanks to Peter Morse, Brendan O’neill, and Pascale Roy-Léveillé for laboratory and field assistance as well as constructive discussion.
Section 7 Literature


Karunaratne KC. 2011. A Field Examination of Climate-Permafrost Relations in Continuous and Discontinuous Permafrost of the Slave Geological Province. *PHd Dissertation*.


Section 8  Tables

Table 1. Summary of field investigations, 2010-2011.

<table>
<thead>
<tr>
<th>Year</th>
<th>Activity Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>2010</td>
<td>1. Installation of $T_{top}$ sensor and hobo data logger, June-July</td>
</tr>
<tr>
<td></td>
<td>2. Installation of iButton array, June-July</td>
</tr>
<tr>
<td></td>
<td>3. Removal of drill core for ice and water content analysis, June-July</td>
</tr>
<tr>
<td></td>
<td>4. Vegetation surveys to determine local species diversity and height, June-July</td>
</tr>
<tr>
<td></td>
<td>5. Soil pit excavation for determination of active-layer bulk density and volumetric moisture content, July</td>
</tr>
<tr>
<td></td>
<td>6. Active-layer measurements, mid-August</td>
</tr>
<tr>
<td></td>
<td>7. Determination of volumetric ice content and gravimetric moisture content, June-July</td>
</tr>
<tr>
<td>2011</td>
<td>8. Snow survey to determine snow depth, April</td>
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<tr>
<td></td>
<td>9. Removal of iButton array, July</td>
</tr>
<tr>
<td></td>
<td>10. Removal of hobo temperature loggers, August</td>
</tr>
<tr>
<td></td>
<td>11. Active-layer measurements, August 21st</td>
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<tr>
<td></td>
<td>12. Determination of organic content, November</td>
</tr>
<tr>
<td></td>
<td>13. Soil Textural Analysis, December</td>
</tr>
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Table 2. Snow characteristics during the 2010-2011 snow season. † denotes depths determined using iButton array. * suggests that the snow depth exceeded the stake length so it could not be found during the period of manual measurements.

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<th>10b</th>
<th>8a</th>
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<td>13/10</td>
<td>12/10</td>
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<td>24/10</td>
<td>9/10</td>
<td>12/10</td>
<td>12/10</td>
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<td>Departure (2011)†</td>
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<td>17/05</td>
<td>17/05</td>
<td>24/05</td>
<td>20/05</td>
<td>26/05</td>
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<td>9</td>
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<td>24</td>
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<td>16</td>
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<td>Snow Depth (cm; April 2011)</td>
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Table 3. Estimation of vegetation abundance in %. x indicates that a few individuals were present while i suggests only one individual was observed.

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Table 4. Average of 10 vegetation height measurements from each site.

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Table 5. Total volumetric water content within the active-layer.

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<th>10b</th>
<th>8a</th>
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<th>5a</th>
<th>5b</th>
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</thead>
<tbody>
<tr>
<td>Water Content (g/cm³)</td>
<td>39.3</td>
<td>25.5</td>
<td>16.1</td>
<td>30.1</td>
<td>16.8</td>
<td>37.4</td>
<td>16.7</td>
<td>18.6</td>
<td>34.5</td>
<td>21.7</td>
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Table 6. Results of intrasample variability in organic content using loss on ignition method.

<table>
<thead>
<tr>
<th>Organic Content</th>
<th>Mean (%)</th>
<th>Minimum (%)</th>
<th>Maximum (%)</th>
<th>Range (%)</th>
<th>Standard Deviation (%)</th>
<th>n</th>
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<tr>
<td>Low</td>
<td>6.1</td>
<td>5.5</td>
<td>6.4</td>
<td>0.8</td>
<td>0.3</td>
<td>11</td>
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<tr>
<td>Medium</td>
<td>38.0</td>
<td>37.6</td>
<td>39.2</td>
<td>1.6</td>
<td>0.4</td>
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<tr>
<td>High</td>
<td>80.4</td>
<td>79.5</td>
<td>81.7</td>
<td>2.2</td>
<td>0.7</td>
<td>11</td>
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Table 7. Active-layer depth statistics, measurements made in mid-august 2010 and 2011. * n refers to the number of measurements taken at each site.

<table>
<thead>
<tr>
<th></th>
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<th>10a</th>
<th>10b</th>
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<th>8b</th>
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<th>6b</th>
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<tr>
<td><strong>2010</strong>&lt;br&gt;(cm)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td>54</td>
<td>54</td>
<td>51</td>
<td>23</td>
<td>52</td>
<td>43</td>
<td>49</td>
<td>45</td>
<td>41</td>
<td>40</td>
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</tr>
<tr>
<td>Median</td>
<td>53</td>
<td>55</td>
<td>53</td>
<td>23</td>
<td>51</td>
<td>44</td>
<td>49</td>
<td>45</td>
<td>40</td>
<td>40</td>
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</tr>
<tr>
<td>Minimum</td>
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<td>50</td>
<td>40</td>
<td>48</td>
<td>43</td>
<td>39</td>
<td>38</td>
<td></td>
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<tr>
<td>Maximum</td>
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<td>58</td>
<td>55</td>
<td>25</td>
<td>55</td>
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<td>4</td>
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<td></td>
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<tr>
<td><strong>Average</strong></td>
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<td>62</td>
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<td>62</td>
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<td>53</td>
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<td>41</td>
<td>60</td>
<td>48</td>
<td>47</td>
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<td>41</td>
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Table 8. $T_{\text{top}}$ characteristics during the September 2010 – April 2011 Period. Temperature measurements from 100 cm below the ground surface.

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<th>10b</th>
<th>8a</th>
<th>8b</th>
<th>6a</th>
<th>6b</th>
<th>5a</th>
<th>5b</th>
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<tbody>
<tr>
<td>Average</td>
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<td>-6.6</td>
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<td>-1.1</td>
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</table>
Table 9. Active-layer freeze-back measurements. Freeze-back was said to begin on 23/09/2010 when the air temperatures fell and remained below 0 °C.

<table>
<thead>
<tr>
<th></th>
<th>11</th>
<th>10a</th>
<th>10b</th>
<th>8a</th>
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<tbody>
<tr>
<td>Duration (Days)</td>
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<td>67</td>
<td>69</td>
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<td>84</td>
<td>82</td>
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<td>81</td>
<td>65</td>
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<tr>
<td>% of Freezing Period</td>
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<td>28</td>
<td>29</td>
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<td>35</td>
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<td>28</td>
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<tr>
<td>Date of Completion</td>
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<td>29/11</td>
<td>01/12</td>
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<td>16/12</td>
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<td>15/12</td>
<td>13/12</td>
<td>27/11</td>
<td>22/11</td>
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</tbody>
</table>
Figure 1. Outer Mackenzie delta and Tuktoyaktuk peninsula with study site at Illisarvik (From Mackay and Burn 2002).
Figure 2. Glacial limits and ice flow lines on the Beaufort Continental Shelf (From Murton 2009: 109)
Figure 3. Surficial Geology of the Beaufort Sea Coastal Region (Hill et al. 1990). Illisarvik indicated by arrow.
Figure 4. Climatic zones of the Mackenzie Delta and surrounding regions (Burns 1973). Illisarvik indicated by arrow.
Figure 5. Climate data for Tuktoyaktuk 1971-2000, 60 km to the east of Illisarvik (Environment Canada 2012).
Figure 6. Climate normal data from Tuktoyaktuk compared with mean monthly temperatures from Illisarvik during the study period (Environment Canada 2012).
Figure 7. Smoothed air temperature from Illisarvik during the study period (September 2010 – August 2011).
Figure 8. Comparison of air temperatures at Illisarvik and Tuktoyaktuk during the period of overlap between September 2010 and August 2011.
Figure 9. Overview of the Illisarvik tundra transect. The inset profile was adapted from Mackay and Burn (2002) and shows the freezeback site identities and locations above the original numbering.
Figure 10. Images of fully instrumented sites taken summer 2010. (A) Site 3; (B) Sites 10a (right) and 10b (left); (C) Site 8a; (D) Site 8b; (E) Site 5b; (F) Site 4.
Figure 11. Temperature monitoring equipment characteristic of each site. iButton photo from http://www.maxim-ic.com/; image of hobo logger and thermistor cable from http://www.onsetcomp.com/.
Figure 12. Mean annual temperature profile through the surface boundary layer, showing the relationship between permafrost and air temperature (From Smith and Riseborough 2002:5)
Figure 13. Daily mean temperatures at the hill top sites. Snow depth determined using iButtons is indicated by the shaded region.
Figure 14. Daily mean temperatures at the hill slope sites. Snow depth determined using iButtons is indicated by the shaded region.
Figure 15. Daily mean temperatures at the hill bottom sites. Snow depth determined using iButtons is indicated by the shaded region.
Figure 16. Comparison of Maximum Snow Depth with (A) mean annual temperature at the top of permafrost; and (B) mean vegetation height.
Figure 17. Depth profiles of (A) Gravimetric Water Content; (B) Volumetric Ice Content; and (C) Soil Organic Content from all sites. Samples were obtained from sediment coring and soil pits during July 2010.
Figure 18. Daily mean temperatures at the top of permafrost from September 2010 to August 2011.
Figure 19. Comparison of Freezeback Duration with: (A) mean annual temperature at the top of permafrost; (B) maximum snow depth; (C) maximum snow depth during the freezeback period; (D) mean snow depth during the freezeback period; (E) 2010 active-layer depth; (F) 2011 active-layer depth; (G) active-layer moisture; and (H) mean vegetation height.