A Field Examination of Climate-Permafrost Relations in
Continuous and Discontinuous Permafrost of the Slave Geological Province

by

Kumari Catharine Karunaratne

B.Sc., The University of Western Ontario, 2000
M.Sc., Carleton University, 2003

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ABSTRACT

Climate-permafrost relations were examined across treeline using field data from the Slave Geological Province. The surface and thermal offsets, parameterized in the TTOP model, were used as a framework for the investigation. Air, ground surface, and permafrost temperatures were measured for two years (2004-06) at 24 peatlands in three study areas near Yellowknife, Colomac, and Ekati, Northwest Territories. The Yellowknife and Colomac study areas lie south of treeline in the discontinuous permafrost zone, but Ekati lies north of treeline in the continuous permafrost zone. Surveys of vegetation, snow, and soil were conducted to assess the role of microclimate on the climate-permafrost relation. Air temperatures across the Slave Province were near climate normals (1971-2000) in the first year, and were 4 °C higher on an annual basis in the second year.

South of treeline, the ground thermal regime was similar and did not respond to spatial or temporal changes in air temperature. North of treeline, surface and ground temperatures were considerably lower, and the climate-permafrost relation was stronger. Differences in climate-permafrost relations within the discontinuous permafrost zone, across treeline, and interannually resulted from differences in the duration of active-layer freezeback, when release of latent heat prevents substantial ground cooling. South of treeline, freezeback of the active layer was prolonged by thick snow covers, while north of treeline, where snow covers were thinner, freezeback was shorter and allowed the ground to cool for the majority of the winter. The variability of the surface offset south

of treeline was controlled by total active-layer water, which is not easily estimated remotely, rather than snow.

The thermal offset was controlled locally by changes in the thermal conductivity ratio (rk), and regionally by climate. To evaluate the thermal offset model, values of rk determined from soil samples were compared with values obtained directly from field temperature measurements. The model performed well under normal climate conditions, but only to the north where freezeback was short. The thermal offset model did not calculate ground temperatures accurately for wet active layers under thick snow covers, or for transient conditions regardless of moisture regime or permafrost temperature.

DEDICATION

This thesis is dedicated to my mother,

Elizabeth Karunaratne

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TABLE OF CONTENTS

ABSTRACT	ii
DEDICATION	iv
ACKNOWLEDGEMENTS	v
TABLE OF CONTENTS	viii
LIST OF TABLES	xii
LIST OF FIGURES	xvi
LIST OF ABBREVIATIONS	xxiv
Chapter One:	Overview and Objectives 1
1.1	Introduction..... 1
1.2	Climate-Permafrost Foundations 3
1.3	The TTOP Model 6
1.4	The Slave Geological Province..... 7
1.5	Research Program 9
1.6	Research Hypothesis and Objectives 11
1.7	Thesis Structure..... 11
Chapter Two:	The Relation between Climate and Permafrost..... 12
2.1	The Climate-Permafrost Relation 12
2.2	Climate-Permafrost Investigations..... 12
2.3	The Surface 17
2.3.1	The surface radiation balance 18
2.3.2	The surface energy balance 20
2.3.3	The surface temperature 21
2.4	The Surface Offset 23
2.4.1	Snow 23
2.4.2	Vegetation 26
2.4.3	Organic layer 27
2.4.4	Subsurface conditions.....29
2.4.5	Relation between air and surface temperature 29
2.4.6	Parameterization of the surface offset 32
2.4.7	The n-factor 32
2.4.8	Derivation of the surface offset 36

2.5	The Active Layer	37
2.5.1	Thermal conductivity	37
2.5.2	Heat capacity	39
2.5.3	Thermal diffusivity	39
2.5.4	Unfrozen water content	40
2.5.5	Latent heat	40
2.5.6	Non-conductive heat transfer	43
2.5.7	Active-layer thermal regime	43
2.6	The Thermal Offset	47
2.6.1	Parameterization of the thermal offset	48
2.7	The TTOP Model	49
2.7.1	Limits of discontinuous permafrost	50
2.8	Concluding Remarks	52
Chapter Three: Regional Setting and Study Period Climate		53
3.1	Introduction	53
3.2	The Slave Geological Province	53
3.2.1	Regional Geology of the Slave Province	55
3.2.2	Climate of the Slave Province	56
3.2.3	Permafrost in the Slave Province	58
3.2.4	Vegetation in the Slave Province	58
3.3	Study Period Climate	60
3.3.1	Freezing seasons	63
3.3.2	Thaw seasons	66
3.4	Summary	69
Chapter Four: Physical Characteristics of the Sites		70
4.1	Introduction	70
4.2	Site selection	70
4.3	Vegetation	71
4.3.1	Yellowknife Vegetation	75
4.3.2	Colomac Vegetation	78
4.3.3	Ekati Vegetation	78
4.4	Soil	80
4.4.1	Yellowknife Soils	82
4.4.2	Colomac Soils	85

4.4.3	Ekati Soils.....	86
4.5	Snow.....	86
4.5.1	Yellowknife and Colomac Snow.....	87
4.5.2	Ekati Snow	92
4.6	Active-Layer Thickness	95
4.6.1	Yellowknife Active-Layer Thickness	99
4.6.2	Colomac Active-Layer Thickness	102
4.6.3	Ekati Active-Layer Thickness	102
4.7	Summary	105
Chapter Five:	Air, Surface and Ground Temperatures in the Slave Geological Province	106
5.1	Introduction.....	106
5.2	Instrumentation	107
5.2.1	Instrument Installation.....	108
5.2.2	Missing Data.....	109
5.3	Air Temperature	109
5.3.1	Air temperatures within the study areas	110
5.3.2	Air temperature among the study areas	117
5.4	Ground Thermal Regime	120
5.5	Surface Temperature	127
5.5.1	Freezing surface temperatures.....	129
5.5.2	Thawing surface temperatures.....	132
5.5.3	Annual surface temperatures	133
5.6	Temperature at the Top of Permafrost	134
5.7	Mineral Sites	136
5.8	Summary Points	141
Chapter Six:	The Surface Offset: Relations between air and surface temperatures in the Slave Geological Province	143
6.1	Introduction.....	143
6.2	Relation between air and surface temperature	144
6.2.1	Thermal orbits	146
6.3	Surface Offsets.....	150
6.4	The Freezing Surface Offset	152
6.5	Snow Cover and the Surface Offsets	158
6.6	The Thawing Surface Offset	162

6.7	Vegetation and the Surface Offsets.....	165
6.8	Soil Water and the Surface Offsets	169
6.9	The Annual Surface Offset	174
6.10	Summary Points	177
Chapter Seven:	The Thermal Offset: Relations between Surface and Permafrost Temperatures in the Slave Geological Province	182
7.1	Introduction.....	182
7.2	Active Layer Thermal Regime.....	183
7.2.1	Nonconductive Heat Transfer	185
7.3	The Thermal Offset.....	187
7.4	Thermal Conductivity Ratio (rk).....	192
7.4.1	$rk_{Temperature}$	193
7.5	Field Measurements of rk	195
7.5.1	Thermal conductivity measurements.....	197
7.5.2	rk_{Soil}	199
7.6	Comparison of rk_{Soil} and $rk_{Temperature}$	202
7.7	Latent Heat and the TTOP Model.....	207
7.8	Summary Points	210
Chapter Eight:	Summary and Conclusions.....	213
References	217	
Appendix A:	Vegetation and Soil Surveys	230
Appendix B:	Temperature Data.....	242

LIST OF TABLES

Table 2.1	Albedo and emissivity of natural surface materials (after Oke 1987, Table 1.1, p. 12).	19
Table 2.2	Thermal properties of soil constituents and snow (after Oke 1987, Table 2.1, p. 44; and Williams and Smith 1989, Table 4.1, p. 90).	22
Table 3.1	Annual mean air temperatures (T_{Aa}) at Yellowknife and Ekati Airports for the study period. Annual mean temperature calculated from daily mean temperatures from 1 September to 30 August. Climate data are courtesy of Environment Canada (2010) for the Yellowknife Airport, and BHP Billiton for the Ekati Airport. Mean annual air temperatures (1971-2000) for Yellowknife and Lupin are also included courtesy of Environment Canada (2010).	62
Table 3.2	Descriptive statistics of the study's freezing seasons for Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports including the timing and duration of the seasons, mean winter air temperature temperatures (1 September to 30 April), freezing degree days (DDF_A), and median March snow depth. Snow data for Ekati was collected by Indian and Northern Affairs Canada – Water Resources Division at Daring Lake (48 km west-northwest of Ekati). Statistics for the normal freezing season (1971-2000) at Yellowknife Airport are also included.	64
Table 3.3	Descriptive statistics of the study's thawing seasons for Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports including the timing and duration of the seasons, mean summer air temperature temperatures (1 May to 31 August), thawing degree days (DDT_A), and rainfall. Rainfall data for Ekati were collected by Indian and Northern Affairs Canada – Water Resources Division at Daring Lake (48 km west-northwest of Ekati). Statistics for the normal thawing season (1971-2000) at Yellowknife Airport are also included.	67
Table 4.1	Mann-Whitney U statistics comparing snow depth measured along transects through the study sites at Yellowknife, Colomac, and Ekati in April 2005 and 2006. Significant differences with p values ≤ 0.05 are in bold.....	91
Table 4.2	Timing of active-layer thickness measurements with respect to end of the thaw season, percent of thaw season lapsed, and percent of thawing degree-days ($\sqrt{DDT_A}$) at time of measurement.	

	Measurements of active-layer thickness at Colomac in 2005 were slightly earlier than at Yellowknife and Ekati.....	97
Table 4.3	Mann-Whitney U statistics comparing active-layer thickness measured along transects through the study sites at Yellowknife, Colomac, and Ekati in 2005 and 2006. Significant differences with p values ≤ 0.05 are in bold.	100
Table 5. 1	Results of principal axis analysis comparing daily mean air temperature at the instrumented sites and the Yellowknife (YZF) and Ekati (YOA) airports: (a) coefficient of determination (r^2), (b) slope, (c) intercept of the principal axis.	113
Table 6.1	Median and range of the surface-defined freezing surface offsets (S.O. _f) and n-factors (n_f) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Data from the instrumented sites were not necessarily available for both years.	155
Table 6.2	Mann-Whitney U test comparing freezing surface offsets (S.O. _f) and n-factors (n_f) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Significant differences with p values ≤ 0.05 are in bold.....	156
Table 6.3	Kendall Tau (τ) correlation between n_f and April snow depth for freezing seasons (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values ≤ 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.....	160
Table 6.4	Median and range of the surface-defined thawing surface offsets (S.O. _t) and n-factors (n_t) at Yellowknife, Colomac and Ekati organic sites for 2005 and 2006. Data from the instrumented sites were not necessarily available for both years.....	164
Table 6.5	Mann-Whitney U test comparing n-factors between open canopy sites, shaded sites and Ekati Tundra sites. Significant differences with p values ≤ 0.05 are in bold.	168
Table 6.6	Kendall Tau (τ) correlation between n_f and August 2006 volumetric soil moisture for freezing seasons (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values < 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.	172

Table 6.7	Median and range of the surface-defined annual surface offsets (S.O. _a) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Data from the instrumented sites were not necessarily available for both years.	176
Table 6.8	Kendall Tau (τ) correlation between annual surface offset (S.O. _a) and April snow depth for (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values ≤ 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.....	180
Table 7.1	Results of Mann-Whitney U test to determine statistical difference in rk_{Soil} between study areas in 2005 and 2006. Statistically different pairs are presented in bold for one-tailed test, confidence level of 0.05. Values of rk_{Soil} were higher at Ekati in both years.....	204
Table A.1	Mean height, and diameter at stem base (DSB) of tall shrubs at the instrumented sites.	230
Table A.2	Classes used to estimate vegetation cover (after Goldsmith et al. 1986).....	231
Table A.3	Coverage classes for vegetation and ground cover at the Yellowknife instrumented sites.	232
Table A.4	Coverage classes for vegetation and ground cover at the Colomac sites.	233
Table A.5	Coverage classes for vegetation and ground cover at the Ekati instrumented sites.	234
Table A.6	Soil characteristics at the Yellowknife instrumented sites. Samples collected in September 2006.	235
Table A.7	Soil characteristics at the Colomac instrumented sites. Samples collected in September 2006.	238
Table A.8	Soil characteristics at the Ekati instrumented sites. Samples collected in September 2006.	240
Table B.1	Mean air temperatures for the (a) freezing season (T_{Af} ; 1 September to 30 April), (b) thawing season (T_{At} ; 1 May to 31 August), and (c) annual period (T_{Aa} ; 1 September to 31 August) at the instrumented sites and the Yellowknife and Ekati airports.....	242

Table B.2	Mean surface temperatures for the (a) freezing season (T_{sf} ; 1 September to 30 April), (b) thawing season (T_{st} ; 1 May to 31 August), and (c) annual period (T_{sa} ; 1 September to 31 August) at the instrumented sites.	243
Table B.3	Maximum, minimum and annual (T_{50a} ; 1 September to 31 August) ground temperatures at 50 cm depth (T_{50}) at the instrumented sites.	244
Table B.4	Maximum, minimum and annual (T_{100a} ; 1 September to 31 August) ground temperatures at 100 cm depth (T_{100}) at the instrumented sites.	245
Table B.5	Monthly mean air temperatures at the instrumented sites.	246
Table B.6	Monthly mean surface temperatures at the instrumented sites.	247
Table B.7	Surface defined values of freezing air (T_{Af}) and surface (T_{sf}) temperature and freezing surface offsets ($S.O._f$).	249
Table B.8	Surface defined values of thawing air (T_{At}) and surface (T_{st}) temperature and thawing surface offsets ($S.O._t$).	250
Table B.9	Surface defined values of annual air (T_{Aa}) and surface (T_{sa}) temperature and annual surface offsets ($S.O._a$).	251
Table B.10	Surface defined values of freezing degree-days for the air (DDF_A) and surface (DDF_S) and freezing n-factors (n_f).	252
Table B. 11	Surface defined values of thawing degree-days for the air (DDT_A) and surface (DDT_S) and thawing n-factors (n_t).	253
Table B.12	Surface defined annual mean temperatures at the surface (T_{sa}) and 100 cm depth (T_{100a}), and the thermal offset ($T.O.$).	254
Table B.13	Surface defined thermal offset ($T.O.$), thawing degree-days for the surface (DDT_S), annual period (P), and thermal conductivity ratio determined from $T.O.$, DDT_S and P	255

LIST OF FIGURES

Figure 1.1	Southern limits of continuous and discontinuous permafrost (modified from Heginbottom <i>et al.</i> 1995).....	2
Figure 1.2	Locations (dots) for measurement to characterize the climate-permafrost relation: air temperature (T_A) from a radiation shield, surface temperature (T_S), and temperature at the top of permafrost (T_{TOP}) (after Lachenbruch <i>et al.</i> 1988, Fig. 17).....	4
Figure 1.3	The Slave Geologic Province (Geological Survey of Canada 1997) with study areas at Yellowknife and Colomac south of treeline (Brown <i>et al.</i> 1998), and the Ekati study area north of treeline.	8
Figure 1. 4	Schematic of instrumented and study sites with Yellowknife, Colomac, and Ekati study regions of the Slave Province. Mineral sites with minimal organic cover are indicated with (*).	10
Figure 2.1	Mean annual temperature profile through the lower atmosphere, active-layer, and permafrost. Modified from Smith and Riseborough (2002, Fig. 3).	15
Figure 2.2	The climate-permafrost thermal relation including the buffer zone components that control the surface offset. Modified from Luthin and Guymon (1974, Fig 4).	24
Figure 2.3	Southern limit of continuous permafrost (Heginbottom <i>et al.</i> 1995) and treeline (Brown <i>et al.</i> 2001).....	31
Figure 2.4	The unfrozen water content at temperatures below 0 °C for various soils. After Williams and Smith (1989, Fig. 1.4).	41
Figure 2.5	Unfrozen water content, thermal conductivity, apparent heat capacity, and thermal diffusivity as a function of temperature for a silty clay (dashed) and a sandy loam (solid). After Williams and Smith (1989, Figs 4.1 and 4.2).	42
Figure 2.6	Ground temperature envelope through active layer and permafrost. Modified from Duchesne <i>et al.</i> (2008).....	44
Figure 3.1	Study areas Yellowknife, Colomac and Ekati within the Slave Geological Province, including Lupin Mine where climate Normals were recorded, and Daring Lake, where daily mean snow depths	

	were recorded by Indian and Northern Affairs Canada – Water Resources Division.....	54
Figure 3.2	Mean monthly air temperatures (lines), and mean monthly precipitation (bars) (1971-2000) for the Yellowknife and Lupin Airports (Environment Canada 2010).	57
Figure 3.3	Daily mean air temperatures (T_{Ad}) for the study period and the preceding year at the Yellowknife Airport (Environment Canada 2010) and Ekati Airport (BHP Billiton 2007). The daily mean temperatures have been smoothed with a 10-day running mean and symbols have been placed every 15 days.	61
Figure 3.4	Daily mean snow depth at Yellowknife Airport (Environment Canada 2010) and Daring Lake for the study period freezing seasons (2004-05 and 2005-06) and the preceding winter (2003-04). Daring Lake is located 48 km west-northwest of Ekati. Daring Lake snow data were collected by Indian and Northern Affairs Canada – Water Resources Division.....	65
Figure 3.5	Monthly total rainfall at Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports for the study period thawing seasons (2005 and 2006) and the preceding summer (2004).	68
Figure 4.1	The Yellowknife study area, showing locations of the five study sites. The Yellowknife Airport is located within 5 km of city-centre.	72
Figure 4.2	The Colomac study area, showing locations of the three study sites.	73
Figure 4.3	The Ekati study area, showing locations of the five study sites. The Ekati Airport is located between the Ekati camp and E-2.....	74
Figure 4.4	Photograph examples of Yellowknife instrumented sites with (a) open canopy, Y-4a; and (b) closed canopy, Y-3b.	77
Figure 4.5	Photograph examples of Colomac instrumented sites with (a) open canopy, C-1a; and (b) closed canopy, C-3c.	79
Figure 4.6	Photograph examples of Ekati instrumented sites at (a) high centred polygon, E-1a; and (b) atop esker, E-5a.	81
Figure 4.7	Examples of soil bulk density values for the Surface, Middle and Base layers of the active layers at Yellowknife, Colomac, and Ekati. Soil samples for the (a) Surface Layer were taken at 0 to 5	

	cm depth; (b) Middle Layer were taken at 10 to 40 cm depth; (c) Base Layer were taken at base of the active layer.....	83
Figure 4.8	Examples of soil volumetric water content for the Surface, Middle and Base layers of the active layers at Yellowknife, Colomac, and Ekati. Soil samples for the (a) Surface Layer were taken at 0 to 5 cm depth; (b) Middle Layer were taken at 10 to 40 cm depth; (c) Base Layer were taken at base of the active layer.....	84
Figure 4.9	Photograph examples of snow conditions south of treeline at (a) Colomac study site C-2, and north of treeline at (b) Ekati study site E-2. Photos taken April 2005.	88
Figure 4.10	Snow density through the snow pack at Yellowknife (Y-3a), Colomac (C-3a) and Ekati (E-2a) in (a) April 2005, (b) January 2006, and (c) April 2006. Snow density at Ekati was only measured in April 2006.	89
Figure 4.11	Snow depths from transects through the study sites at Yellowknife, Colomac and Ekati. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths. Sample size ranged from 60 to 120. Snow surveys were not completed at Ekati in January 2006, and measurements at Ekati esker site E-5 in April 2006 were excluded because of their unusually low values.	90
Figure 4.12	Snow depths at Yellowknife instrumented sites in (a) April 2005; (b) January 2006; and (c) April 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and n=9.	93
Figure 4.13	Snow depths at Colomac instrumented sites in (a) April 2005; (b) January 2006; and (c) April 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and n=9.	94
Figure 4.14	Snow depths at Ekati instrumented sites in (a) April 2005; and (b) April 2006;. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and n=9.	96
Figure 4.15	Active-layer thicknesses at the study sites for summers 2005 (dark grey) and 2006 (light grey). Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and n=24.....	98

Figure 4.16	Active-layer thicknesses at Yellowknife instrumented sites for summers 2005 (dark grey) and 2006 (light grey). Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and n=9.....	101
Figure 4.17	Active-layer thicknesses at Colomac instrumented sites for summers 2005, and 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and n=9.....	103
Figure 4.18	Active-layer thicknesses at Ekati instrumented sites for summers 2005, and 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and n=9.....	104
Figure 5.1	Daily mean air temperatures at sites Y-1a, C-1a and E-1a, from September 2004 to September 2006. The daily mean air temperature was smoothed with a five-day running mean for clarity.....	111
Figure 5.2	Principal axes of daily mean air temperatures within the (a) Yellowknife, (b) Colomac, and (c) Ekati study areas.	112
Figure 5.3	Mean air temperatures for the (a) freezing season (1 September to 30 April), (b) thawing season (1 May to 31 August), and (c) annual period (1 September to 31 August) at the instrumented sites and the Yellowknife and Ekati airports.	114
Figure 5.4	Principal axes of daily mean air temperatures at (a) Yellowknife Airport (YZF) and site Y-1a, and (b) Ekati Airport (YOA) and site E-1a.	116
Figure 5.5	Principal axes of daily mean air temperatures between the study areas at (a) Y-1a and C-1a, (b) C-1a and E-1a, and (c) Y-1a and E-1a.	119
Figure 5.6	Daily mean temperatures at the surface (T_{sd}), and 20 cm, 50 cm, and 100 cm depth for sites E-3b, C-2a, and Y-2c from September 2004 to September 2006.....	122
Figure 5.7	Active-layer freeze-back duration as a percent of the freezing season.	125
Figure 5.8	Daily mean surface temperatures from September 2004 to September 2006 at Yellowknife, Colomac, and Ekati sites with (a) cold surfaces, and (b) warm surfaces.	128

Figure 5.9	Mean surface temperatures for the (a) freezing season (1 September to 30 April), (b) thawing season (1 May to 31 August), and (c) annual period (1 September to 31 August) at the organic and mineral instrumented sites.....	130
Figure 5.10	Daily mean ground temperatures at 100 cm depth from September 2004 to September 2006 at selected Yellowknife, Colomac, and Ekati instrumented sites with (a) cold ground, and (b) warm ground.....	135
Figure 5.11	Annual (1 September to 31 August) mean ground temperatures at top of permafrost (100 cm depth) at the organic and mineral instrumented sites.....	137
Figure 5.12	Daily mean surface temperatures from September 2004 to September 2006 at an organic and mineral site at (a) Ekati and (b) Yellowknife.....	138
Figure 5.13	Daily mean ground temperatures (150 cm depth) from September 2004 to September 2006 at and organic and mineral site at (a) Ekati and (b) Yellowknife.	140
Figure 6.1	Daily mean air and surface temperature for 1 September 2004 to 31 September 2006 at (a) Ekati (E-1a) and (b) Yellowknife (Y-1a). Daily means were calculated from 12 observations. Air temperatures were measured in a radiation shield 1.5 m above the ground, and surface temperatures were measured 5 cm below the radiative surface at the base of the live moss.	145
Figure 6.2	Plots of monthly mean surface vs. monthly mean air temperature for 2004-05. Plots (a), (c) and (e) are examples of sites with low mean annual surface temperatures (i.e. cold), while plots (b), (d) and (f) are sites with high mean annual surface temperatures (i.e. warm). September 2004 (SEP), February 2005 (FEB), and July 2005 (JUL) are labelled on each plot.	147
Figure 6.3	Plots of monthly mean surface vs. monthly mean air temperature for 2005-06 at sites with exposed mineral soil at the surface at (a) Yellowknife Y-5a, and (b) Ekati E-5a. September 2005 (SEP), February 2006 (FEB), and July 2006 (JUL) are labelled on each plot.....	149
Figure 6.4	Surface defined values of freezing (a) surface offsets and (b) n-factors at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are	

	represented by (+). Data from the instrumented sites were not necessarily available for both years.	154
Figure 6.5	Relation between freezing n-factor (n_f) and April snow depth (cm) at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.	159
Figure 6.6	Surface defined values of thawing (a) surface offsets and (b) n-factors at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are represented by (+). Data from the instrumented sites were not necessarily available for both years.	163
Figure 6.7	Comparison of (a) freezing n-factors (n_f), and (b) thawing n-factors (n_t) between open and shaded sites at Yellowknife and Colomac and the Ekati tundra sites. Data from the instrumented sites were not necessarily available for both years.	167
Figure 6.8	Relation between freezing n-factor (n_f) and active-layer water at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.....	171
Figure 6.9	Surface defined values of the annual surface offsets at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are represented by (+). Data from the instrumented sites were not necessarily available for both years.....	175
Figure 6.10	Relation between annual surface offset ($S.O._a$) and April snow depth (cm) at the Yellowknife, Colomac and Ekati study sites for (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.....	178
Figure 6.11	Relation between annual surface offset ($S.O._a$) and active-layer soil moisture at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years. There was a significant correlation between $S.O._a$ and active-layer water in 2004-05 ($\tau=0.47$; $p=0.005$), but not in 2005-06.	179
Figure 7.1	Daily mean temperature at the surface and top of permafrost from 1 September 2004 to 31 September 2006 at (a) Ekati (E-1a) and (b) Yellowknife (Y-1a). Daily means were calculated from 12 observations. Surface temperatures were measured 5 cm below the radiative surface at the base of the live moss, and temperatures at the top of permafrost were measured at 100 cm.	184

Figure 7.2	Change to the thermal gradient between 20 and 50 cm every 2 hours from 1 September 2004 to 31 August 2006 at (a) Ekati (E-4b), (b) Colomac (C-3b), and (c) Yellowknife (Y-2c).....	186
Figure 7.3	Annual thermal offset (T.O.) at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Annual period was defined by surface temperature. Values for the two years were not necessarily for the same sites.....	189
Figure 7.4	Relation between annual thermal offset (T.O.) and thawing degree-days at the surface (DDT _s) at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=-0.73$, $p<0.000$) and (b) 2005-06 ($\tau=-0.66$, $p=0.005$).....	191
Figure 7.5	Thermal conductivity ratio based on temperatures ($rk_{Temperature}$) at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. The annual thermal offset (T.O.), thawing degree-days at the surface (DDT _s), and annual period (P) were used to calculate $rk_{Temperature}$. Surface temperature was used to define P, T.O., and DDT _s . Values for the two years were not necessarily for the same sites.	194
Figure 7.6	Relation between thermal conductivity ratio based on temperatures ($rk_{Temperature}$) and total active-layer soil water at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=-0.62$, $p<0.000$), and (b) 2005-06 ($\tau=-0.62$, $p=0.008$).	196
Figure 7.7	Measurements of thermal conductivity obtained using Hukseflux TPSY02 Thermal Conductivity Measurement System compared with estimates of thermal conductivity obtained from soil samples using Johansen's Model (1975).....	198
Figure 7.8	Thermal conductivity ratio (rk) values for the Surface, Middle and Base Layers of the active layers at Yellowknife, Colomac and Ekati. Soil samples were collected in September 2005 and 2006.....	200
Figure 7.9	Thermal conductivity ratio based on field measurements of soil (rk_{Soil}) at Yellowknife, Colomac and Ekati. Soil samples were collected throughout the active layers in September 2005 and 2006. Values of k_t and k_f were obtained using Johansen (1975), and the weighted harmonic mean was used to summarize k_t and k_f for the active layer.	203
Figure 7.10	Relation between thermal conductivity ratio based on temperatures ($rk_{Temperature}$) and thermal conductivity ratio based on soil samples (rk_{Soil}) at the Yellowknife, Colomac, and Ekati instrumented sites	

	for (a) 2004-05 ($\tau=0.52$, $p=0.003$), and (b) 2005-06 ($\tau=0.46$, $p=0.050$)	205
Figure 7.11	Relation between the difference between thermal conductivity ratio based on temperatures and those based on soil samples ($rk_{Soil} - rk_{Temperature}$) versus total active-layer water at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=0.55$, $p=0.002$), and (b) 2005-06 ($\tau=0.22$, $p=0.349$).....	208
Figure 7.12	Relation between the difference between thermal conductivity ratio based on temperatures and those based on soil samples ($rk_{Soil} - rk_{Temperature}$) versus the active-layer freeze-back period (FBP) as a percentage of the freezing season duration at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=0.47$, $p=0.007$), and (b) 2005-06 ($\tau=0.30$, $p=0.217$).....	209

LIST OF ABBREVIATIONS

C_g	Heat capacity of the ground ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
C_m	Heat capacity of soil minerals ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
C_o	Heat capacity of organic material ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
C_w	Heat capacity of soil water ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
C_a	Apparent heat capacity ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
C_s	Heat capacity of frozen soil ($\text{J kg}^{-1} \text{ }^\circ\text{C}^{-1}$)
DDF_A	Freezing degree-days for the air ($^\circ\text{C-d}$)
DDT_A	Thawing degree-days for the air ($^\circ\text{C-d}$)
DDF_S	Freezing degree-days for the surface ($^\circ\text{C-d}$)
DDT_S	Thawing degree-days for the surface ($^\circ\text{C-d}$)
k_f	Thermal conductivity of frozen ground ($\text{Wm}^{-1}\text{K}^{-1}$)
k_t	Thermal conductivity of thawed ground ($\text{Wm}^{-1}\text{K}^{-1}$)
k_{sat}	Thermal conductivity of saturated soils ($\text{Wm}^{-1}\text{K}^{-1}$)
k_{dry}	Thermal conductivity of dry soils ($\text{Wm}^{-1}\text{K}^{-1}$)
k_s	Thermal conductivity of soil solids ($\text{Wm}^{-1}\text{K}^{-1}$)
k_w	Thermal conductivity of soil water ($\text{Wm}^{-1}\text{K}^{-1}$)
k_t	Thermal conductivity of dry soils ($\text{Wm}^{-1}\text{K}^{-1}$)
$K_\downarrow$	Incoming short-wave radiation (Wm^{-2})
$K_\uparrow$	Reflected short-wave radiation (Wm^{-2})
$L_\downarrow$	Incoming long-wave radiation (Wm^{-2})
$L_\uparrow$	Outgoing long-wave radiation (Wm^{-2})
L_f	Latent heat of fusion (J m^{-3})
MAAT	Mean annual air temperature ($^\circ\text{C}$)
MAST	Mean annual surface temperature ($^\circ\text{C}$)
MAGT	Mean annual ground temperature ($^\circ\text{C}$)
n	porosity
n_f	Freezing n-factor
n_t	Thawing n-factor
P	Annual period (d)
P_f	Freezing period (d)
P_t	Thawing period (d)
Q^*	Net radiation (Wm^{-2})
Q_H	Sensible heat flux (Wm^{-2})
Q_E	Latent heat flux (Wm^{-2})
Q_G	Ground heat flux (Wm^{-2})
ΔQ_S	Net energy stored (Wm^{-2})
rk	Thermal conductivity ratio (k_t/k_f)
$rk_{Temperature}$	rk from thermal offset model (See Equation 7.2)
rk_{Soil}	rk from soil samples
S_R	Soil saturation ratio
S.O.	Surface offset ($^\circ\text{C}$)
S.O. _a	Annual surface offset ($^\circ\text{C}$)

$S.O._f$	Freezing surface offset (°C)
$S.O._t$	Thawing surface offset (°C)
T_A	Air temperature (°C)
T_{Aa}	Annual mean air temperature (°C)
T_{Ad}	Daily mean air temperature (°C)
T_{Af}	Freezing season air temperature (°C)
T_{At}	Thawing season air temperature (°C)
T_S	Surface temperature (°C)
T_{Sa}	Annual surface temperature (°C)
T_{Sd}	Daily mean surface temperature (°C)
T_{Sf}	Freezing season surface temperature (°C)
T_{St}	Thawing season surface temperature (°C)
T_{100}	Ground temperature at 100 cm (°C)
T_{TOP}	Temperature at the top of permafrost (°C)
T_{TOPa}	Annual mean temperature at the top of permafrost (°C)
$T.O.$	Thermal offset (°C)
X_m	Volumetric fraction of soil minerals
X_o	Volumetric fraction of organic material
X_w	Volumetric fraction of soil water
z	Depth of the ground (m)
θ_u	Unfrozen water content

Chapter One: Overview and Objectives

1.1 Introduction

This thesis examines climate-permafrost relations in the Slave Geological Province of the Canadian Shield. Permafrost is a geologic manifestation of climate, and is defined as ground that remains below 0 °C for at least two consecutive years (ACGR 1988). An active layer that freezes and thaws annually lies above permafrost. Permafrost forms in areas of high latitude and altitude where the intensity and duration of the freezing season allow frozen ground to persist through the following thawing season. Permafrost is integral to northern environmental systems because it both affects and is affected by vegetation, hydrology, geomorphology, and biogeochemical cycling. It also impacts infrastructure and development in cold regions.

The permafrost region of Canada has been divided into continuous and discontinuous permafrost zones (Figure 1.1) (Heginbottom *et al.* 1995). Permafrost in the continuous zone is both spatially and temporally continuous (Brown 1967). In the discontinuous permafrost zone, permafrost exists only where surface and ground characteristics combine to maintain ground temperatures below 0 °C. Since permafrost is primarily controlled by climate, it is directly affected by climate change through rising air temperatures. However, permafrost is also indirectly affected through climate related changes to other components of the northern environmental system such as vegetation. Climate change is expected to cause ground temperatures to increase,

Figure 1.1 Southern limits of continuous and discontinuous permafrost (modified from Heginbottom *et al.* 1995).

resulting in a northward displacement of the boundaries for the continuous and discontinuous permafrost zones (Anisimov and Nelson 1997; Stendel and Christensen 2002). Observations of recent ground warming in permafrost regions have been reported for Siberia (Romanovsky *et al.* 2007; Romanovsky *et al.* 2010), Alaska (Osterkamp and Romanovsky 1999; Osterkamp 2005), the Canadian Arctic (Smith *et al.* 2005; Smith *et al.* 2010), and Europe (Harris *et al.* 2003).

The widespread degradation of permafrost will have wide-ranging impacts on northern environmental systems (Turetsky *et al.* 2002; Yoshikawa and Hinzman 2003; Epstein *et al.* 2004; Lewkowicz 2007). A field-based understanding of the climate-permafrost relation in the continuous and discontinuous permafrost zones is required to model present and future permafrost conditions, which will, in turn, assist in the prediction of the effects of permafrost degradation.

1.2 Climate-Permafrost Foundations

Lachenbruch *et al.* (1988) separated the climate-permafrost relation into the difference between the annual mean temperatures of (1) the air and the ground surface; and (2) the ground surface and the top of permafrost. These have come to be known as the surface and thermal offsets respectively (Figure 1.2). The surface offset is controlled by the surface energy balance, while the thermal offset is dependent on the ground's frozen and thawed thermal conductivities; however both the surface and thermal offsets are also controlled by the intensity of the freezing and thawing seasons.

Figure 1.2 Locations (dots) for measurement to characterize the climate-permafrost relation: air temperature (T_A) from a radiation shield, surface temperature (T_S), and temperature at the top of permafrost (T_{TOP}) (after Lachenbruch *et al.* 1988, Fig. 17).

The challenge in defining the climate-permafrost relation lies in characterizing the factors that control the offsets in the freezing and thawing seasons. In summer, vegetation, moisture availability, and the proximity of the thawing front cause surface and air temperatures to differ only slightly (Eaton *et al.* 2001; Klene *et al.* 2001; Karunaratne and Burn 2004). In winter, snow insulates the surface from the colder air above resulting in large temperature differences between the two (Karunaratne and Burn 2004). Latent heat released at depth during ground freezing also contributes to the deviation of surface temperature from that of the air (Goodrich 1978). On an annual basis, the surface offset results in surface temperature exceeding air temperature.

The thermal offset is affected by seasonal changes in the ground's thermal conductivity. Ice has a thermal conductivity that is four times that of water (Williams and Smith 1989, p. 87). Thus, when soil water freezes, the ground's thermal conductivity increases. The largest changes to soil thermal conductivity upon freezing occur in peat due to the low thermal conductivity of organic matter, and the characteristically large seasonal change in moisture content. The thermal conductivity of bedrock, which lacks water, changes little throughout the year.

These seasonal changes in surface and ground characteristics require that the examination of climate-permafrost relations be divided into the differences between: (1) air and surface temperature in summer; (2) air and surface temperature in winter; (3) surface and permafrost temperature in summer; and (4) surface and permafrost temperature in winter. The TTOP (Temperature at the Top Of Permafrost) model describes how the surface and thermal offsets relate climate (air temperature) to permafrost (ground temperature) throughout the year (Smith and Riseborough 1996).

1.3 The TTOP Model

The TTOP model uses indices to parameterize the surface and thermal offsets and thereby relate climate (air temperature) to permafrost (ground temperature) (Smith and Riseborough 1996). Within the TTOP model, the surface offset is summarized through n -factors, which are ratios of ground surface to air freezing or thawing indices (Lunardini 1978; Smith and Riseborough 1996). The thermal offset is defined by way of the soil-conductivity ratio; a ratio of thawed to frozen ground thermal conductivity (Romanovsky and Osterkamp 1995). The TTOP model is:

$$T_{TOPa} = \frac{k_t n_t DDT_A - k_f n_f DDF_A}{k_f P} \quad [1.1]$$

where T_{TOPa} is the annual mean temperature at the top of permafrost, DDF_A and DDT_A are the air freezing and thawing indices (degree-days, °C-d), n_f and n_t are the freezing and thawing n -factors, and k_f and k_t are the frozen and thawed ground thermal conductivities ($\text{Wm}^{-1}\text{K}^{-1}$), and P is the annual period (365 days).

The TTOP model has been used to predict regional and continental permafrost conditions (Henry and Smith 2001; Wright *et al.* 2003), and to examine the behaviour of the climate-permafrost relation through numerical modelling (Smith and Riseborough 2002; Riseborough 2002, 2003, 2007; Juliussen and Humlum 2007). However, there are several aspects of the TTOP model that have not been examined. First, the TTOP model requires surface temperature to be estimated from air temperature since records of surface temperature are limited. Surface temperature is controlled by surface and subsurface characteristics, and it is thought that surface temperature can be estimated remotely

knowing the terrain, snow, vegetation and soil conditions (Jorgenson and Kreig 1988; Henry and Smith 2001; Juliussen and Humlum 2007). However, these site-specific characteristics, particularly snow cover, vary interannually, while subsurface conditions are difficult to determine remotely and are integrated with the conditions that TTOP seeks to predict. The estimation of surface temperature is a significant challenge to the TTOP model. Second, the TTOP model does not explicitly consider the latent heat associated with soil freezing and thawing. Third, the TTOP model uses single values for k_f and k_t . This does not address the slow transition between the thawed and frozen state of the ground which is common in moist soils, nor the presence of both organic and mineral soil layers within the active layer. Finally, TTOP is an equilibrium model, and it is unknown if it can be used under the transient conditions of degrading or aggrading permafrost. This thesis examines the TTOP model through field measurements of air, surface and ground temperatures, and associated surface and ground characteristics in the Slave Geological Province.

1.4 The Slave Geological Province

The Slave Geological Province lies between Great Slave Lake and Coronation Gulf, and extends across the boundary between the continuous and discontinuous permafrost zones (Figure 1.3). The area contains some of the oldest rocks on earth and is well known for its long history of mining and the recent discovery of diamondiferous kimberlite. Knowledge of climate-permafrost relations in the Slave Province is of great interest to those involved in the mining industry. However, the lack of permanent roads throughout the Slave Province has restricted access and hindered regional permafrost investigations. The permafrost literature for the Slave Province focuses on ground-ice

Figure 1.3 The Slave Geologic Province (Geological Survey of Canada 1997) with study areas at Yellowknife and Colomac south of treeline (Brown *et al.* 1998), and the Ekati study area north of treeline.

conditions north of treeline (Wolfe *et al.* 1997; Dredge *et al.* 1999; Hu *et al.* 2003).

Permafrost conditions near Yellowknife have been reported for natural terrain by Brown (1973), and have been described with respect to engineering problems around the city (Apsler 1979; Seto *et al.* 2004; Hoeve *et al.* 2004), and at the Giant Mine (Bateman 1949; Boyle 1951).

1.5 Research Program

The purpose of this research is to conduct a field investigation of the climate-permafrost system in the subarctic taiga and low arctic tundra of the Slave Province. The parameterization of the surface and thermal offsets in the TTOP model was used as the framework for the investigation. Three study areas were chosen in the Slave Province: Yellowknife, Colomac Mine, and Ekati Diamond Mine (Figure 1.3). Yellowknife (62° 28' N 114° 27' W) is located in the subarctic taiga and the discontinuous permafrost zone. Colomac (64° 12' N 116° 01') is also located in the subarctic taiga and the discontinuous permafrost zone, but is further north and closer to treeline than Yellowknife. Ekati (64° 42' N 110° 37' W) is north of treeline in the continuous permafrost of the low arctic tundra. Study sites containing instrumented sites were established in each of the three study areas to measure air, surface, active layer, and top of permafrost temperatures between 2004 and 2006 (Figure 1.4). Site characteristics such as snow and active-layer depths, vegetation, and soil properties were surveyed at each of the instrumented sites. The relation between air and permafrost temperatures was examined with respect to the physical characteristics of the surface and the ground.

Figure 1. 4 Schematic of instrumented and study sites with Yellowknife, Colomac, and Ekati study regions of the Slave Province. Mineral sites with minimal organic cover are indicated with (*).

1.6 Research Hypothesis and Objectives

The hypothesis of this thesis is that latent heat, released during freezeback of the active layer, affects the spatial and temporal variation of the climate-permafrost relation, and may compromise the utility of the TTOP Model. The objectives are:

- (1) To determine the variability of the surface and thermal offsets within and among the study areas in the Slave Province.
- (2) To identify the physical characteristics that control the surface and thermal offsets in the Slave Province.
- (3) To evaluate the TTOP model.

1.7 Thesis Structure

This thesis consists of eight chapters. In Chapter Two, the relation between climate and permafrost will be examined both north and south of treeline, and the TTOP model will be described in detail. Chapter Three describes the Slave Geological Province and the climate of the study period. The physical characteristics of the instrumented sites are reported in Chapter Four, as well as the field and laboratory methods used to quantify the site characteristics. The results are presented in three separate chapters. Chapter Five examines variations in air, surface and ground temperatures within and among the study areas. Chapter Six considers the factors controlling the surface offset. Chapter Seven examines the thermal offset. The conclusions of the thesis are summarized in Chapter Eight.

Chapter Two: The Relation between Climate and Permafrost

2.1   The Climate-Permafrost Relation

Permafrost is related to climate. The zones of permafrost prevalence marked on continental-scale maps correlate generally with mean annual air temperature (Brown 1967), and recently permafrost temperatures have increased in response to climate warming (Osterkamp 2005; Smith *et al.* 2005; Romanovsky *et al.* 2007; Burn and Zhang 2009; Smith *et al.* 2010). The relation between climate and permafrost is obscured by surface and lithologic properties – most notably snow and soil thermal characteristics – which control the heat flow in and out of the ground (Luthin and Guymon 1974; Smith 1975; Smith and Riseborough 1983). Variations in these properties complicate the climate-permafrost relation and cause it to vary both spatially and temporally, so that differences in surface and lithologic properties can explain the distribution of permafrost in the discontinuous permafrost zone (Brown and Péwé 1973; Smith and Riseborough 1983; Williams and Burn 1996; Shur and Jorgenson 2007). An understanding of the climate-permafrost relation is of utmost importance for estimates of permafrost distribution under past, present, and future climates. This chapter examines the relation between climate and continental permafrost using Lachenbruch *et al.*'s (1988) framework of the surface and thermal offsets. Specifically, differences in climate-permafrost relations north and south of treeline in summer and winter will be addressed.

2.2   Climate-Permafrost Investigations

Climate-permafrost investigations have been closely associated with the development of permafrost maps. Ideally, maps of permafrost distribution should be

based on ground temperature measurements, but over a country the size of Canada these measurements are relatively sparse and were even more so in the past. Therefore the first permafrost maps used mean annual air temperatures (MAAT) and an understanding of climate-permafrost relations to plot the southern limits of permafrost (Heginbottom 1984, 2002). The first map of permafrost distribution was published in 1882 and depicted the southern limit of continuous permafrost in Russia aligning with the -2°C MAAT isotherm (Heginbottom 1984). Nikiforoff (1928) used the same association between MAAT and mean annual ground temperature (MAGT) to produce the initial map of continuous permafrost in Canada. Over the next few decades, as interest in permafrost grew, Bratsev (1939) and Muller (1945) produced continental-scale permafrost maps to refine Nikiforoff's map. The first map of permafrost in Canada based on field observations, from a questionnaire sent to northern settlements, was produced by Jenness (1949). This map differentiated between continuous and discontinuous permafrost and included the -5 and -3.3°C MAAT isotherms.

R.J.E Brown and others at the National Research Council of Canada began collecting temperature measurements and field observations of permafrost in 1953 and continued for the next several decades (Brown 1963, Heginbottom 1984). Brown (1967) produced the first Canadian map of permafrost based on ground temperatures. This map divided the permafrost region into continuous and discontinuous permafrost based on the Russian convention that the southern limit of continuous permafrost has a MAGT of -5°C . Using this map, Brown concluded that MAGT was on average 3.3°C higher than MAAT, and that the southern limits of the continuous and discontinuous permafrost zones were associated with the -8.3°C and -1.1°C MAAT isotherms respectively.

Although these boundaries were interpolated using field data, in areas where little information was available the boundary was drawn according to the MAAT isotherm. Brown's 1967 map has been refined as more field information has become available (Brown 1978, Heginbottom *et al.* 1995).

Only small-scale maps of permafrost distribution can be constructed based solely on MAAT. Local variations in permafrost presence, thickness and temperature are dependent on terrain characteristics such as relief, vegetation, drainage, snow, and soil and rock characteristics, which vary spatially as well as temporally (Brown 1967). This is especially true for the discontinuous permafrost zone (Shur and Jorgenson 2007). The ground surface is an important component of the climate-permafrost system because its temperature integrates the thermal influences of the various terrain factors (Ferrians and Hobson 1973).

Lachenbruch *et al.*'s (1988) conceptual model defined the climate-permafrost system in terms of the mean annual temperatures of the air, ground surface, and top of permafrost (Figure 1.2). This model divides the energy exchanges into those occurring within the buffer or surface layer, the layer above the ground that contains vegetation, snow, and organic layers; and those occurring within the active layer (Williams and Smith 1989). The mean annual temperature at the ground surface (MAST) is higher than MAAT primarily because of snow cover (Figure 2.1). In winter the insulating snow prevents heat loss and keeps the surface considerably warmer than the overlying air, while vegetation and evaporative fluxes generally cause the surface to be slightly cooler than the air in summer. On an annual basis, surfaces that receive a snow cover are warmer than the air (Nicholson and Granberg 1973). The difference between MAST and

Figure 2.1 Mean annual temperature profile through the lower atmosphere, active-layer, and permafrost. Modified from Smith and Riseborough (2002, Fig. 3).

MAAT ($MAST - MAAT$) is known as the surface offset. The surface offset is normally a positive quantity.

Under equilibrium conditions, MAST is also higher than the MAGT or the mean annual temperature at the top of permafrost (TTOP) due to seasonal changes in the thermal properties of the active layer (Goodrich 1982; Burn and Smith 1988; Romanovsky and Osterkamp 1995). Ice conducts heat approximately four times as effectively as water (Williams and Smith 1989), and thus, to maintain an equilibrium annual heat balance, temperature gradients are considerably steeper in summer than in winter when the active layer is frozen. The discrepancy between the seasonal temperature gradients causes MAGT to decrease with depth through the active layer. The difference between TTOP and MAST ($TTOP - MAST$) is usually negative and is known as the thermal offset. The climate-permafrost relation is a composite of these two offsets, and the TTOP at any given location is the sum total of MAAT, the surface offset, and the thermal offset (Riseborough and Smith 1998).

Parameterizations of the surface offset (Carlson 1952; Lunardini 1978) and the thermal offset (Kudryavtsev 1977; Romanovsky and Osterkamp 1995) were operationalized in the TTOP model by Smith and Riseborough (1996). TTOP is a numerical equilibrium model that predicts temperature at the top of permafrost (T_{TOP}) based on MAAT and estimates of the surface and thermal offsets. The TTOP model has been used to successfully replicate Brown's (1967) map of permafrost in Canada (Henry and Smith 2001), and to explore the climate-permafrost relation (Smith and Riseborough 2002).

2.3 The Surface

In micrometeorology, the surface is defined as the principal plane where (1) drag on airflow is exerted, (2) precipitation is intercepted, (3) radiation is absorbed, reflected and emitted, and (4) the main transformations of energy and mass occur (Oke 1987, p. 33). This plane is termed the *active surface*, and generally lies at the canopy, which could be tens of metres above the ground depending on the height of the vegetation. Within the climate-permafrost system, the surface is the boundary beneath which conductive heat transfer dominates (Smith and Riseborough 2002). While this simple definition is well suited to numerical modeling, it is problematic for field measurement. For natural surfaces, the transition from convective to conductive transfer occurs over a layer rather than at a plane. The thickness of this surface layer depends on the presence and density of the ground cover vegetation, and the porosity of the soil's organic horizon. The layer is thin on exposed mineral soils, thick on low-density feather mosses, and is temporally variable with changes in saturation and between the thawed and frozen states.

For field examinations of the climate-permafrost system, the temperature of the surface is often measured at 2 to 5 cm depth in mineral soils and at the base of the live mosses in organic soils (Romanovsky and Osterkamp 1995; Klene *et al.* 2001; Riseborough 2003; Karunaratne and Burn 2004; Jorgenson *et al.* 2010). This method is attractive because it shields the sensor from radiation, and positions the sensor close to the relatively warm ground surface during the summer, and is appropriate for measuring winter surface temperature, near the peat-snow interface. Such data can also be used to compare organic and exposed mineral surfaces. In this thesis, the surface refers to the ground surface rather than the active surface.

The temperature of the surface is a function of the surface energy balance, which is driven by the net radiation (Q^*). The radiation balance quantifies fluxes of radiation at the surface to give Q^* , while the surface energy balance describes how Q^* drives other energy fluxes away from and toward the surface. The surface temperature at any given location and time is the result of a specific combination of the surface energy fluxes (Outcalt 1972), and generally follows Q^* (Oke 1987, p. 23). Thus the surface radiation and energy balances and their controls are integral to surface temperature.

2.3.1 The surface radiation balance

At the surface, the radiation balance is:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) \quad [2.1]$$

where $K\downarrow$ and $K\uparrow$ are surface incoming and outgoing short-wave solar (0.15 to 3.0 μm) radiation respectively. $L\downarrow$ and $L\uparrow$ are surface incoming and outgoing long-wave terrestrial (3.0 to 100 μm) radiation respectively. Atmospheric constituents intercept $K\downarrow$ by absorbing, scattering, and reflecting incoming radiation. Surface objects reduce $K\downarrow$ by intercepting incoming radiation, resulting in lower $K\downarrow$ in the shade as opposed to full sunlight. $K\downarrow$ varies throughout the day, and seasonally at high latitudes. $K\uparrow$ is short-wave radiation reflected off the surface. The magnitude of $K\uparrow$ is dependent on the magnitude of $K\downarrow$ and the reflectivity of the surface, or albedo (Table 2.1; Oke 1987, p. 12).

Long-wave radiation is a function of a body's temperature and emissivity - the ability of a body to emit energy (Oke 1987, p. 10). $L\downarrow$ from the atmosphere is dependent on the emissivity, temperature, and concentration of the atmospheric constituents, especially water vapour, carbon dioxide, and ozone (Oke 1987, p. 13).

Table 2.1 Albedo and emissivity of natural surface materials (after Oke 1987, Table 1.1, p. 12).

Surface	Remarks	Albedo	Emissivity
Soils	Dark, wet to Light, dry	0.05 to 0.40	0.98 to 0.90
Grass	Long (1 m) to Short (0.02 .m)	0.16 to 0.26	0.90 to 0.95
Forest	Deciduous	0.15 to 0.20	0.97 to 0.98
	Coniferous	0.05 to 0.15	0.97 to 0.99
Water	Small zenith angle	0.03 to 0.10	0.92 to 0.97
	Large zenith angle	0.10 to 1.00	0.92 to 0.97
Snow	Old to Fresh	0.40 to 0.95	0.82 to 0.99

Contributions of long-wave radiation from nearby objects are controlled by the emissivity and temperature of the object, as well as the size and proximity of the object. Natural surface materials have a narrow range of emissivities (Table 2.1).

On an annual mean basis, Q^* at the surface decreases with latitude from approximately $2.9 \text{ MJ m}^{-2} \text{ d}^{-1}$ near Yellowknife to $1.2 \text{ MJ m}^{-2} \text{ d}^{-1}$ directly north along the Arctic coast (Hare 1997). At any given time and surface, Q^* is either positive if incoming exceeds outgoing radiation, or negative if outgoing exceeds incoming radiation. In general, Q^* is positive during the day and in summer, and negative at night and in winter. The value of Q^* determines the direction and magnitude of the thermal fluxes that make up the surface energy balance.

2.3.2 The surface energy balance

The surface energy balance describes fluxes of thermal heat at the surface and is driven by Q^* :

$$Q^* = Q_H + Q_E + Q_G + \Delta Q_S \quad [2.2]$$

where Q_H , Q_E , and Q_G are the sensible, latent, and ground heat fluxes respectively, and ΔQ_S is the change in stored energy (Oke 1987, p. 34). The relative partitioning of Q_H , Q_E , Q_G , and ΔQ_S is controlled by terrain characteristics (Oke 1987, p.25).

At the surface, Q_E is heat absorbed during evaporation and melting, or heat released during condensation and freezing. The magnitude of Q_E is primarily controlled by the availability of surface water; being negligible for a dry surface. The magnitude of Q_G is controlled by the temperature gradient between the surface and the sub-surface, and the thermal conductivity of the soil. Thermal conductivity (k , $\text{W m}^{-1} \text{ }^\circ\text{C}^{-1}$) is the soil's ability to conduct heat, and is a function of the conductivity of the soil constituents (Table

2.2). Ice has a thermal conductivity four times that of water, thus a soil's thermal conductivity regularly increases substantially upon freezing (Williams and Smith 1989, p 87). The partitioning of Q^* between Q_G and the convective heat transfers at the surface is controlled by the thermal admittance of the surface and atmosphere. Thermal admittance describes the ease with which a body releases or takes up heat.

The steady-state ground heat flux through a homogeneous soil can be estimated using Fourier's heat conduction equation:

$$Q_G = -k \frac{\partial T}{\partial z} \quad [2.3]$$

where T is temperature and z is depth in the ground (Williams and Smith 1989, p.84).

Under equilibrium conditions Q_G away from the surface in summer equals Q_G toward the surface in winter. If the thermal conductivity of the ground changes upon freezing, the temperature gradient must change to maintain an equilibrium heat flux over the year.

2.3.3 The surface temperature

Fluxes of thermal energy at the surface are driven by temperature gradients between the surface and overlying air and between the surface and the underlying ground. When the incoming flux exceeds the outgoing flux, the surface warms as heat energy is stored. When the outgoing flux exceeds the incoming flux, the surface cools as stored heat is released (Oke 1987, p. 7).

The surface is a critical boundary in the climate-permafrost relation. Surface temperature is common to both the surface and thermal offsets within the Lachenbruch model, and therefore must be obtained to relate climate to permafrost conditions. Unfortunately, records of surface temperature exist for only a few locations, and usually for limited time periods. Air temperature records, on the other hand,

Table 2.2 Thermal properties of soil constituents and snow (after Oke 1987, Table 2.1, p. 44; and Williams and Smith 1989, Table 4.1, p. 90).

Material	Heat Capacity (J kg ⁻¹ °C ⁻¹)	Thermal Conductivity (W m ⁻¹ °C ⁻¹)	Thermal Diffusivity (x10 ⁻⁶ m ² s ⁻¹)	Thermal Admittance (J m ² s ^{-1/2} °C ⁻¹)
Soil Constituents				
Quartz	800	8.8	4.14	-
Clay minerals	900	2.92	1.22	-
Organic matter	1920	0.25	0.10	-
Water (0°C)	4180	0.56	0.13	1545
Ice (0°C)	2100	2.24	1.16	2080
Air (still)	1010	0.025	20.63	5
Snow				
Fresh snow	2090	0.08	0.10	130
Old snow	2090	0.42	0.40	595

are available for many communities throughout the permafrost region. Since surface and air temperature are interdependent and positively correlated because they control the thermal gradient for Q_H , surface temperature can be estimated from air temperature by understanding the terrain components that control the surface offset.

2.4 The Surface Offset

Although the surface offset concept was proposed by Lachenbruch *et al.* (1988), the term was coined by Henry and Smith (2001), and has also been referred to as the nival offset because of the dominant role of snow (Smith and Riseborough 2002). The surface offset can range widely in an area with the same MAAT, due to variation in site-specific characteristics that control the surface energy balance and thereby change MAST (Henry and Smith 2001). Luthin and Guymon (1974) described the thermal regime between the ground and atmosphere as the *buffer zone*, consisting of vegetation, snow, and surface organic layers (Figure 2.2). These layers are interrelated and determine the surface offset by controlling the surface energy balance.

2.4.1 Snow

Within the climate-permafrost literature, snow cover is a surface characteristic, like vegetation, rather than part of the ground. At high latitudes, the net effect of snow cover is to increase the MAST above MAAT (Gold 1963) (Figure 2.1). The low thermal conductivity of snow inhibits heat loss from the ground during winter (Sturm *et al.* 1997), and isolates the ground from fluctuations in air temperature causing a weak correlation between air and ground surface temperatures during winter (Zhang *et al.* 1997; Karunaratne and Burn 2003). The influence of snow on the surface temperature is

Figure 2.2 The climate-permafrost thermal relation including the buffer zone components that control the surface offset. Modified from Luthin and Guymon (1974, Fig 4).

dependent on the snow cover's depth, density, and structure (Goodrich 1982; Sturm and Johnson 1992; Zhang 2005).

In general, the insulating ability of snow increases with its thickness, so relatively warm surfaces and large surface offsets are associated with thick snow covers. Sturm and Holmgren (1994) found that within a horizontal distance of 1.5 m, snow depths varied between 10 and 35 cm and were associated with snow-ground interface temperatures between -15 and -5°C. Field measurements show that ground surface temperatures increase with snow depth up to 50 cm, but thicker snow covers have little further effect (Nicholson and Granberg 1973; Smith 1975). Iijima *et al.* (2010) found surface temperatures at sites in the central Lena River basin responded to air temperatures in the early freezing season until snow depths reached 10 cm.

The insulating capability of the snow is inversely related to its thermal conductivity. Numerical modeling has shown that decreasing the thermal conductivity of snow from $0.7 \text{ Wm}^{-1}\text{K}^{-1}$ to $0.1 \text{ Wm}^{-1} \text{ K}^{-1}$ results in a 5.5 °C increase in mean annual ground surface temperature (Goodrich 1982). The thermal conductivity of snow is positively correlated with its density (Sturm *et al.* 1997). Snow density and thermal conductivity can range from 0.135 g cm^{-3} and $0.070 \text{ Wm}^{-1}\text{K}^{-1}$ for new snow, to 0.444 g cm^{-3} and $0.359 \text{ Wm}^{-1}\text{K}^{-1}$ for hard drift snow (Sturm *et al.* 1997).

Sturm and Johnson (1992) reviewed measurements of snow thermal conductivity and density, and found that the structure of snow particles controlled the relation between thermal conductivity and density of snow. Depth hoar, a layer of coarse crystals that commonly develops near the base of snow packs under a steep temperature gradient, has a considerable spread of thermal conductivities for a given density (Sturm

and Johnson 1992). In the subarctic the depth-hoar layer frequently represents between 50 to 80% of the snow pack (Sturm and Johnson 1991), and is especially thick on the tundra in areas with shrubby vegetation cover (Sturm *et al.* 2001).

The arrival of snow cover and its timing with respect to the beginning of the freezing season can greatly influence the MAST (Goodrich 1982). If a thick snow cover arrives at the beginning of the freezing season, it can retard freezing of the ground by inhibiting the evacuation of latent heat. At Barrow, Alaska, delaying the arrival of snow cover by 10 days was predicted to decrease MAST by 0.7 °C (Ling and Zhang 2003). A thin snow cover can act to cool the surface and thereby promote freezing by increasing the albedo and emissivity (Zhang 2005).

2.4.2 Vegetation

Vegetation's effects on the surface offset are multifaceted, and difficult to isolate because they are interwoven with climate and terrain factors (Brown 1963). The most apparent influence of vegetation is the reduction of Q^* at the ground surface through obstruction of incoming solar radiation, or shading (Brown 1963; Dingman and Koutz 1974; Luthin and Guymon 1974; Smith 1975; Heggem *et al.* 2006).

Vegetation can also affect the surface temperature indirectly in summer. First, the reduction in Q^* through shading causes lower evaporation rates, and thus greater water availability and cooler surfaces. Second, leaf and needle litter, along with the availability of water, promote the development of an organic layer and the growth of mosses. Third, vegetation also indirectly affects surface temperatures through its influence on surface wind velocities (Brown 1963).

The greatest impact that vegetation has on the surface offset is through its ability to affect snow-cover thickness (Smith 1975). On the tundra, and in other unsheltered environments, wind erodes, transports, and redeposits snow and thereby controls the spatial distribution of snow-cover thicknesses (Pomeroy *et al.* 1993; Liston and Sturm 1998). The ability of the wind to erode and transport snow increases with speed, while a reduction of wind speed results in snow deposition. Vegetation lowers the mean wind speed by exerting a frictional drag on the lower atmosphere. Thus vegetation elements such as protruding shrubs, trap transported snow and are associated with thick snow covers in tundra environments.

In forests, falling snow is intercepted by trees and retained in the canopy for extended periods before it either falls or melts to the ground, or sublimates (Hedstrom and Pomeroy 1998). This leads to lower snow accumulation in forests than in nearby clearings (Gray and Male 1981). The interception efficiency of forests increases with leaf area and canopy cover, and is highest for coniferous forests (Pomeroy and Goodison 1997). Interception of snow and the resulting thin snow cover have been shown to cause low ground surface temperatures and to influence the ground thermal regime (Smith 1975; Karunaratne and Burn 2004).

2.4.3 Organic layer

The organic layer is situated at the ground surface with predominantly convective energy exchange above and conductive energy exchange below. Although ground-covering vascular plants are considered part of the organic layer, the mosses, lichens, and peat within this layer have a much greater influence on the surface offset through their insulating ability than vascular plants. In summer, the surface organic layer

has an extremely low thermal conductivity due to the low bulk density and moisture content of the mosses and lichens. The organic layer also affects surface moisture. Mosses, especially *Sphagnum*, are hygroscopic and have a high water holding capacity, but are unable to draw water from depth and transpire (Brown 1963). However, the top of the mosses dry quickly, and water is wicked to the surface through capillary action and evaporated. This can act to enhance Q_E because moisture is available for evaporation, but the thermal conductivity of the organic layer remains low because it is unsaturated. Also, water vapour can easily migrate through large pores of organic material. Numerous studies have shown that the removal of the organic layer results in active-layer thickening and increases of sub-surface summer temperatures (Linell 1973; Smith 1975; Nicholas and Hinkel 1996).

Karunaratne and Burn (2004) suggested that surfaces with thick organic layers remained cool in spring because the thaw front remained within the depth of diurnal temperature fluctuation for an extended period and inhibited surface warming. Surface and air temperatures converge once the frost table retreats below the depth of diurnal temperature fluctuation. Since near-surface thermal diffusivity controls the rate of thaw penetration, and consequently the duration of the period of suppressed surface temperature, surfaces with high near-surface thermal diffusivity will be warmer over the entire thaw season than surfaces with low near-surface thermal diffusivity. This observation was based on surface temperatures measured at 2-5 cm depth, rather than the radiative surface.

Precipitation in autumn causes saturation of the organic layer and a dramatic increase in its thermal conductivity because water conducts heat better than air. When the

organic layer freezes in winter, the thermal conductivity increases substantially again, thus promoting heat loss in winter (Brown 1963; Nicholas and Hinkel 1996). Thick organic layers are associated with cold ground temperatures because they promote ground cooling in winter and inhibit ground warming in summer.

2.4.4 Subsurface conditions

Luthin and Guymon's (1974) model of the buffer zone is important for summer conditions, when heat flow is into the ground. In contrast, Karunaratne and Burn (2004) have shown that subsurface conditions affect surface temperatures, especially in winter. As the active layer freezes, latent heat is released and inhibits ground cooling until the active layer is completely frozen. The influence of subsurface conditions on the surface offset is dependent on the water content of the active layer, as well as air temperature and snow conditions that drive the ground temperature gradient (Goodrich 1982; Smith and Riseborough 2002). The effect of subsurface conditions on the surface temperature during winter is the theoretical basis for estimating the presence or absence of permafrost from the temperature at the base of the snowpack (Haeberli 1973; Ednie and Lewkowicz 2004; Bonnaventure and Lewkowicz 2008).

2.4.5 Relation between air and surface temperature

The difference between air and surface temperature is smaller in summer than in winter. In summer, the differences between air and surface temperatures are smallest where Q_H is high. Thus the largest surface offsets in summer are thought to occur over moist surfaces where Q_E dominates, and for initially thawing surfaces where a steep thermal gradient causes Q_G to be high. The difference between air and surface temperature is large in winter due to the presence of snow, which insulates the ground

surface from fluctuations in air temperature. In addition, cooling of the surface is inhibited by latent heat released at depth as soil water freezes. Thus the largest surface offsets in winter are thought to occur where the snow cover is thick and active-layer moisture is high.

There are systematic differences in the terrain characteristics that control the surface offset in the discontinuous and continuous permafrost zones. The gradation between discontinuous and continuous permafrost lies within the Forest-Tundra ecotone; a latitudinal region that extends across central Canada from the outer Mackenzie Delta southeast towards Churchill and along the southwest shores of Hudson Bay (Hare and Ritchie 1972). The region encompasses the transition from closed-canopy boreal forest in the south to northern tundra. Within the Forest-Tundra, the southern limit of continuous permafrost parallels the northern limit of tree growth, or treeline (Figure 2.3). The presence of frozen ground does not inhibit tree growth (Gill 1975), therefore treeline is not dependent on the discontinuous-continuous permafrost boundary, suggesting instead that treeline is the controlling variable. Since vegetation exerts a considerable control on the surface energy balance both directly through shading and transpiration and indirectly through its control on snow distribution, moisture/evaporation, and organic layer development, differences in surface offset across treeline are expected.

The presence of trees in the discontinuous permafrost zone reduces the equivalence between air and surface temperature in three ways. First, the trees inhibit redistribution of snow by wind, allowing a uniform, thick, low-density snow pack to develop as opposed to a thin wind-packed snow cover present on the tundra.

Figure 2.3 Southern limit of continuous permafrost (Heginbottom *et al.* 1995) and treeline (Brown *et al.* 2001).

Second, the vegetation shades the surface in summer, and third, reduces the wind speed at the surface. In summer, vegetation characteristics have been shown to have less influence on the surface offset than the surface thermal properties (Karunaratne and Burn 2004).

2.4.6 Parameterization of the surface offset

Surface temperature can be estimated from air temperature using models of the surface offset. Index models that summarize the surface energy balance are favoured over physically-based ones, because they require far fewer input parameters. In the TTOP Model, the surface offset is parameterized using n-factors, indices that relate air to surface temperature through their ratio integrated over the freezing and thawing seasons (Lunardini 1978; Jorgenson and Kreig 1988; Taylor 1995; Klene *et al.* 2001; Karunaratne and Burn 2004).

2.4.7 The n-factor

The n-factor summarizes the surface energy balance under given microclimatic conditions and is calculated from measurements of air and surface temperatures for entire freezing and thawing seasons. Detailed descriptions of the microclimatic controlling characteristics such as vegetation, moisture conditions, and snow cover accompany calculated n-factor values. Once determined, an n-factor is applied to a surface with similar microclimatic controlling characteristics to estimate surface temperature from the local air temperature.

Initially known as “correction factors”, n-factors were the product of a permafrost engineering study undertaken by the U.S. Army Corps of Engineers at Fairbanks, Alaska in the late 1940s (U.S. Army 1950; Carlson 1952; Cysewski and Shur

2008). Over the next thirty years, n-factors were determined for a variety of engineered surfaces to model frost penetration. Lunardini (1978) produced the first comprehensive review of n-factors, which included an examination of the relations between n-factors and heat transfers associated with engineered surfaces, and tables of published n-factors.

Jorgenson and Kreig (1988) first applied n-factors to natural surfaces to predict regional permafrost distribution. Since then, the use of n-factors to estimate the surface temperature in natural permafrost environments has greatly increased (Buteau *et al.* 2004; Juliussen and Humlum 2007; Duchesne *et al.* 2008), and several studies have examined the behaviour of n-factors in undisturbed natural environments (Taylor 1995; Klene *et al.* 2001, 2003; Klene 2008; Karunaratne and Burn 2003, 2004; Karunaratne *et al.* 2008; Juliussen and Humlum 2007; Hinkel *et al.* 2008).

The n-factor is the time-temperature integral for the surface over that for the air, and is calculated for the freezing (n_f) and the thawing (n_t) seasons. Freezing and thawing degree-days are commonly used to integrate temperature over the season so that:

$$n_f = \frac{DDF_S}{DDF_A} \quad [2.4a]$$

and,

$$n_t = \frac{DDT_S}{DDT_A} \quad [2.4b]$$

where DDF_S , DDT_S and DDF_A , DDT_A are the freezing and thawing degree-days for the surface and air respectively (Klene *et al.* 2001; Karunaratne and Burn 2004). Degree-days are straightforward to calculate and are therefore used as temperature indices in this research. In the spring and autumn, daily mean air temperatures commonly fluctuate across 0 °C making determination of seasonal end points difficult. Therefore, surface temperature, which fluctuates only moderately, is used to define the freezing and thawing

seasons (Taylor 1995; Karunaratne and Burn 2004; Juliussen and Humlum 2007).

Positive daily mean temperatures are not included in n_f calculations, and negative values are not used for n_t calculations. Typically, surface temperatures measured at 2-5 cm are used to calculate n-factors, however some of the early explorations of n-factor behaviour measured surface temperature at 20 cm (Taylor 1995; Burn 1998; Karunaratne and Burn 2003; Burn and Smith 1988).

The behaviour of n-factors is primarily a function of the surface temperature, so the surface and subsurface terrain characteristics that control surface temperature and surface offsets also control n-factors. Low n-factors indicate large differences between air and surface temperatures: low n_f signify warm winter surfaces and low n_t signify cool summer surfaces. Generally, variation in n_f is attributed to snow cover and active-layer moisture conditions. Freezing n-factors as low as 0.10 were reported for moist peatlands near Yellowknife with more than 50 cm snow cover (Karunaratne *et al.* 2008).

Theoretically, n_f could be as high as 1.0 for exposed bedrock free of snow where the surface and air are at the same temperature, but such values have not been published.

Freezing n-factors reported from tundra environments are generally above 0.50 (Smith *et al.* 1998; Klene *et al.* 2008) and considerably lower south of treeline (Taylor 1995; Karunaratne and Burn 2004; Karunaratne *et al.* 2008). Variation in n_t is attributed to moisture and shading conditions. Thawing n-factors are less variable than n_f and considerably higher, even exceeding 1.0. Typical n_t values range between 0.7 to 0.9 but extreme values of 0.5 for a heavily shaded organic site and 1.5 for an exposed dry site have been reported (Taylor 1995; Klene *et al.* 2001; Karunaratne and Burn 2004).

Although variation in n-factors is largely attributed to surface and subsurface terrain conditions that control surface temperature, influences of air temperature have also been reported. Locally, extreme n_i values in maritime environments can result from cool advected air masses that are not in equilibrium with the radiatively-heated surfaces (Kade *et al.* 2006). On a continental scale, n_f is thought to be inversely correlated with MAAT. Sensitivity analysis with a finite element model of the ground thermal regime (TONE, Goodrich 1982) has suggested that n_f will decrease with increasing MAAT particularly at snow depths greater than 30 cm (Riseborough and Smith 1998). The influence of MAAT on n_f may be surrogate for the more direct influence of latent heat released from the active layer during freezing. As the MAAT increases, the duration of active-layer freezeback does as well thereby raising mean freezing season temperatures at the surface. However, it may be impossible to test the influence of MAAT on n_f in the field due to insufficient control on snow cover and subsurface moisture conditions over an area large enough to capture significant variation in MAAT.

Seasonal surface offsets are different from n-factors despite sharing the same controlling surface and subsurface characteristics. Surface offsets are the absolute difference between air and surface temperature while n-factors represent the relative difference and are therefore independent of the magnitude of the absolute difference. For a given value of n_f , as the seasonal index for the air increases the surface offset also increases. Therefore, n-factors rather than seasonal surface offsets should be used in comparing the relation between air and surface temperatures interannually or between areas with different MAAT.

2.4.8 Derivation of the surface offset

The surface offset is derived using n-factors and degree-days to integrate air temperature for the freezing and thawing seasons. According to Smith and Riseborough (2002):

$$MAAT = \frac{DDT_A - DDF_A}{P} \quad [2.5]$$

where P is the annual period of 365 days, and:

$$MAST = \frac{DDT_A \times n_t - DDF_A \times n_f}{P} \quad [2.6]$$

Calculating mean annual temperatures through degree-day totals and period duration negates the need to define the seasons. The annual surface offset ($S.O._a$) is:

$$S.O._a = MAST - MAAT \quad [2.7a]$$

$$S.O._a = \frac{DDT_A(n_t - 1) + DDF_A(1 - n_f)}{P} \quad [2.7b]$$

If n_t is set to 1.0 (Henry and Smith 2001):

$$S.O._a = \frac{DDF_A(1 - n_f)}{P} \quad [2.8]$$

Offsets for the freezing ($S.O._f$) and thawing ($S.O._t$) seasons are:

$$S.O._f = \frac{DDF_A(n_f - 1)}{P_f} \quad [2.9a]$$

$$S.O._t = \frac{DDT_A(n_t - 1)}{P_t} \quad [2.9b]$$

where P_f and P_t are the periods of the freezing and thawing seasons in days respectively.

High values of n_f and n_t imply small differences between air and surface temperature and thus low surface offsets. Conversely, surface offsets increase with air index values (DDT_A and DDF_A). Thus for a given n-factor, $S.O._f$ will increase with

latitude as DDF_A increases, and $S.O.$, will decrease with latitude as DDT_A decreases. The freezing and thawing surface offsets will decrease with longer seasons.

2.5 The Active Layer

The thermal offset is a function of the ground thermal properties of the active layer that govern its thermal regime. The presence of soil water in the active layer results in seasonal changes in ground thermal properties, which drive the thermal offset. Water also affects the active layer through the release and uptake of latent heat and facilitates nonconductive heat transfers.

2.5.1 Thermal conductivity

The thermal regime of the ground is largely dependent on the ground's thermal conductivity. Based on the thermal conductivities of the soil constituents outlined in Table 2.2, the thermal conductivity of the ground increases with the mineral fraction, particularly that of quartz, and as water content increases at the expense of air. Increasing the soil moisture increases the soil's thermal conductivity until saturation (Oke 1987, p. 43). Since the thermal conductivity of ice is four times that of water, the thermal conductivity of a soil dramatically increases as pore-water freezes (Williams and Smith 1989, p. 87). Thus if water is present, a soil's frozen thermal conductivity (k_f) is higher than its thawed thermal conductivity (k_t) (Williams and Smith 1989, p. 89).

Thermal conductivity can be estimated from soil samples using Johansen's method (1975). This model requires volumetric estimates of the soil constituents and calculates k by interpolating between dry and saturated conditions based on the saturation ratio (S_R). Values of S_R range from 0 to 1.0. The thermal conductivity for saturated soils

(k_{sat}) is determined using the geometric mean of the thermal conductivities and volumetric fractions of the soil constituents, such that:

$$k_{sat} = k_s^{1-n} \cdot k_w^n \quad [2.10]$$

where, k_s and k_w are the thermal conductivities of the soil solids and soil water respectively, and n is the porosity. The same model is used to determine the thermal conductivity of saturated frozen soils except that the conductivity of ice rather than liquid water is used. In his authoritative monograph, Farouki (1981) estimated that thermal conductivities of dry soils (k_{dry}) were constant at $0.058 \text{ Wm}^{-1}\text{°C}^{-1}$ for organic soils, and may be obtained empirically for natural mineral soils using:

$$k_{dry} = \frac{0.135\gamma_d + 64.7}{2700 - 0.947\gamma_d} \pm 20\% \text{ Wm}^{-1}\text{°C}^{-1} \quad [2.11]$$

where, γ_d is the dry bulk density in kg m^{-3} .

For organic soils with intermediate moisture conditions, k_t is obtained using:

$$k_t = k_{dry} + (k_{sat} - k_{dry})S_R^2 \quad [2.12]$$

and k_f is obtained using:

$$k_f = k_{dry} \left(\frac{k_{sat}}{k_{dry}} \right)^{S_R} \quad [2.13]$$

For fine-grained natural soils with intermediate moisture conditions, k_t is obtained using:

$$k_t = k_{dry} + (k_{sat} - k_{dry})(\log S_R + 1) \quad [2.14]$$

and k_f is obtained using:

$$k_f = k_{dry} + (k_{sat} - k_{dry})S_R \quad [2.15]$$

For coarse-grained mineral material, k_{dry} and k_t are determined differently from fine-grained soils. For coarse-grained material, k_{dry} is determined using:

$$k_{dry} = 0.039n^{-2.2} \pm 25\% \text{ Wm}^{-1}\text{°C}^{-1} \quad [2.16]$$

and k_t is obtained from:

$$k_t = k_{dry} + (k_{sat} - k_{dry})(0.7 \log S_R + 1) \quad [2.17]$$

2.5.2 Heat capacity

The thermal regime of the ground is also controlled by the ground's heat capacity (C). The volumetric heat capacity is the heat required to raise the temperature of a unit volume of a substance by 1 °C, and can be estimated for soil using:

$$C_g = X_m C_m + X_o C_o + X_w C_w \quad [2.18]$$

where C_g , C_m , C_o and C_w are the heat capacities of the ground, soil minerals, organic material and water respectively (Table 2.2), and X_m , X_o , and X_w are the volume fractions of the soil minerals, organic material and water respectively (Williams and Smith 1989, p. 91). The heat capacity of a soil increases linearly with soil moisture content due to the high heat capacity of water (Oke 1987, p. 43).

2.5.3 Thermal diffusivity

The rate of temperature change in soil is governed by the thermal diffusivity, which is the ratio of soil's thermal conductivity to its heat capacity (Oke 1987, p. 85). As soil moisture increases, the thermal diffusivity rises due to the increase in thermal conductivity; however as the soil approaches 20% soil moisture content, the thermal diffusivity decreases because the thermal conductivity changes little, while the heat capacity continues to rise (Oke 1987, p. 43). The thermal diffusivity is limited at low soil moisture contents by the low thermal conductivity and at high moisture contents by the high heat capacity. When soil water freezes, the thermal diffusivity increases because the thermal conductivity of ice is higher than that of water, and the heat capacity is lower (Table 2.2).

2.5.4 Unfrozen water content

When the ground temperature falls below 0 °C soil water begins to freeze, causing the thermal properties of the ground to change because the thermal properties of water and ice differ. The unfrozen water content of a soil decreases with decreasing temperature below 0 °C as more pore water freezes, and the rate of decrease is dependent on soil texture (Koopmans and Miller 1966). In silty and clayey soils, the unfrozen-water content is considerable at temperatures slightly below 0 °C because the chemistry and particle-water energetics associated with small grain-size depress the freezing point (Williams and Smith 1989, p. 6). As the temperature drops, an increasing amount of pore water freezes and the unfrozen water content declines (Figure 2.4). Coarse-grained soils (sand-size and larger) have a low unfrozen-water content because almost all the pore-water freezes at temperatures just below 0 °C. Since the thermal properties of a soil change as pore-water freezes, the unfrozen water content and the soil thermal properties below 0°C are temperature dependent, especially for fine-grained soils at temperatures between 0 and -2 °C (Williams and Smith 1989, p. 89) (Figure 2.5).

2.5.5 Latent heat

Latent heat of fusion ($3.33 \times 10^8 \text{ J m}^{-3}$) is released as soil-water freezes, absorbed as it thaws, and is distributed over a range of temperatures below 0 °C, because of the temperature dependency of the unfrozen water content. The effects of latent heat cause the ground to display an apparent heat capacity in which the volumetric heat capacity of the soil is dwarfed by the temperature dependent release or absorption of latent heat (Williams and Smith 1989, p. 92). The apparent heat capacity (C_a) causes the

Figure 2.4 The unfrozen water content at temperatures below 0 °C for various soils. After Williams and Smith (1989, Fig. 1.4).

Figure 2.5 Unfrozen water content, thermal conductivity, apparent heat capacity, and thermal diffusivity as a function of temperature for a silty clay (dashed) and a sandy loam (solid). After Williams and Smith (1989, Figs 4.1 and 4.2).

thermal diffusivity of frozen soils to be highly temperature dependent. C_a is given by:

$$C_a(T) = C_s(T) + L_f(\partial\theta_u/\partial T)_T \quad [2.19]$$

where C_s is the heat capacity of the frozen soil and L_f is the latent heat of fusion and $\partial\theta_u/\partial T$ is the slope of the unfrozen water content *versus* temperature.

2.5.6 Non-conductive heat transfer

Several studies have examined non-conductive heat transfers in the active layer (Nelson *et al.* 1985; Outcalt *et al.* 1990; Hinkel and Outcalt 1994; Kane *et al.* 2001). Many of these processes involve the transfer of sensible and latent heat associated with the movement of water. These coupled heat-mass transfers occur under a variety of gradients such as: gravitational, osmotic, density, and pore-water and vapour pressure.

Kane *et al.* (2001) suggested that non-conductive heat transfers were seasonal and particularly important during periods of phase change, such as the infiltration of melt water in spring. Internal distillation – the evaporation, migration, and condensation in unsaturated soils – is also thought to be prevalent during soil freezing in early winter (Outcalt *et al.* 1990; Hinkel and Outcalt 1994). Various lines of evidence, including closure of a heat budget using conductive analysis only (e.g. Romanovsky and Osterkamp 1995), suggest that convective effects are of second-order importance overall, although for short periods and in particular locations they may be important components of the ground thermal regime.

2.5.7 Active-layer thermal regime

As the annual surface temperature wave propagates into the ground the amplitude of the wave diminishes and the wave is increasingly lagged. The ground thermal properties dictate how the temperature wave changes with increasing depth.

Figure 2.6 Ground temperature envelope through active layer and permafrost. Modified from Duchesne *et al.* (2008).

The amplitude and attenuation of the wave will decrease rapidly with depth in ground with low thermal diffusivity. The ground temperature envelope, above the depth of zero annual amplitude, is bounded by the annual minimum and maximum temperatures which converge with depth (Figure 2.6). The direction of heat flow does not change seasonally below the depth of zero annual amplitude. At sites that do not experience seasonal changes in the active layer's thermal properties due to the freezing of soil water, the mean annual ground temperature increases with depth along the geothermal gradient (approximately $1^{\circ}\text{C}/40\text{ m}$). This gradient is sufficiently low that mean annual surface temperatures are similar to those at the top of permafrost and at the depth of zero annual amplitude.

The thermal regime is complicated at sites that experience seasonal ground thawing and freezing of the active layer. Outcalt and Hinkel (1996) divided the thermal regime of the active layer into four seasons: (1) thaw season (spring and summer following snowmelt); (2) active-layer freezeback season (fall and early winter); (3) frozen season (mid to late winter); and (4) snow ablation season (early spring). Thawing of the active layer begins in the spring after snowmelt when surface temperatures are above 0°C . The rate of thaw is controlled by the thermal properties of the active layer and the temperature gradient dictated by the surface temperature and depth of the thaw front. As the thaw season progresses, the rate of active layer thaw decreases exponentially – as approximated by the Stefan Solution – as the temperature gradient decreases with the migration of the frost line away from the surface (Andersland and Ladanyi 2004, p.61). The rate of active-layer thaw also declines towards the end of the thaw season because the base of the active layer is often ice-rich (Mackay 1980, 1983) and therefore

progression of the frost line is limited by the absorption of latent heat as ground ice melts. The active layer reaches its maximum depth in late August or early September before surface temperatures drop below 0 °C.

Freezing of the active layer differs from thawing in that freezing proceeds from both the surface downwards and from top of permafrost upwards if a temperature gradient is present. Upward freezing from the top of permafrost begins slowly when the mean daily surface temperature declines causing isothermal conditions throughout the active layer. As soon as the daily mean surface temperature drops below 0 °C, downward freezing of the active layer begins from the surface.

During the active-layer freezeback season, latent heat is released as soil water freezes, and temperatures near the middle and base of the active layer remain isothermal at or slightly below 0 °C (Outcalt and Hinkel 1996). The persistence of temperatures near 0 °C due to latent heat is referred to as the zero-curtain. Sumgin *et al.* (1940), as quoted by Muller (1947, p. 17), are credited for naming the zero-curtain. However this term is poorly defined and has been used to describe a condition or effect, a soil layer, a freezing boundary, a time lag, and a length of time (Outcalt *et al.* 1990). During the freezeback season, the surface thermal signal cannot propagate through the active layer to the underlying permafrost. Active-layer temperatures will remain close to 0 °C throughout the entire winter where freezing of the active layer continues the entire winter. In contrast, active-layer temperatures will readily decline in response to the air temperature if there is little soil water, as in bedrock.

Temperatures at the ground surface, within the active layer and at the top of permafrost decrease, often abruptly, at the start of the frozen season, which follows

freezeback of the active layer. At this time sensible heat rather than latent heat is removed from the ground. The frozen season often ends abruptly when the active layer warms rapidly due to infiltration and freezing of snow-melt water. During the ablation season, surface temperature rises, but does not rise above 0 °C on a daily basis until the snow has disappeared.

2.6 The Thermal Offset

The thermal offset is an active-layer phenomenon caused by seasonal differences in thermal conductivity. Thermal equilibrium conditions for the ground are maintained when heat leaving the ground over the long winter is equal to heat entering the ground during the short summer. Where the frozen thermal conductivity (k_f) exceeds the thawed thermal conductivity (k_t), the equilibrium is retained through a change in the temperature gradient. In summer, the temperature gradient is steep with temperatures similar to air temperatures at the surface, rapidly declining with depth as the low thermal conductivity of the ground is unable to dissipate the surface energy. In winter, the temperature gradient through the active layer is less steep, and surface temperatures are closer to TTOP than to air temperatures due to the presence of snow. The annual expression of these differences in thermal conductivity and subsequent differences in temperature gradient throughout the active layer cause curvature of the mean annual ground temperature profile, such that mean annual ground temperature decreases with depth to the top of permafrost. In permafrost beneath the active layer, seasonal changes in ground thermal properties are minimal, and the mean annual ground temperature increases with depth along the geothermal gradient. Thus the thermal offset shifts the minimum mean annual ground temperature from the surface to the top of permafrost.

The thermal offset increases with the difference between the frozen and thawed thermal conductivity. Bedrock and dry soils do not support thermal offsets because only a change in the state of water will change the thermal conductivity of the soil. Peat soils, on the other hand, are associated with large thermal offsets because they have a high porosity and are often saturated. When peat is dry in summer and saturates prior to freezing in winter, the difference between k_t and k_f is substantial. For example, a peat soil with bulk density of 0.1 g cm^{-3} would have a k_t value of $0.05 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$ when dry and a k_f value of $1.89 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$ when saturated (Johansen 1975).

2.6.1 Parameterization of the thermal offset

Parameterization of the thermal offset is obtained through the ratio of thawed and frozen thermal conductivity (rk) of the active layer, such that:

$$rk = k_t/k_f \quad [2.20]$$

This ratio summarizes the different thermal conductivities and associated thermal gradients in the active layer during the summer and winter. Similar to the n -factor, rk is used as a multiplier to represent changes in conductivity in the different seasons. Since k_t is less than k_f , rk for the permafrost condition is commonly less than 1.0 and equal to 1.0 in bedrock and dry material where there is no soil water. Through sensitivity analyses, Riseborough and Smith (1996) determined that typical rk values range from 0.6 to 0.9 for mineral soils and from 0.3 to 0.6 for organic soils. The rk does not refer to absolute values of k , so a dry peat with no water may have a rk similar to bedrock despite large differences in thermal conductivities.

The mean annual temperature at the top of permafrost is:

$$\text{TTOP} = \frac{(rk \times \text{DDT}_S) - (\text{DDF}_S)}{P} \quad [2.21]$$

(equivalent to Romanovsky and Osterkamp 1995, eq. 11) and given equation 2.7, the thermal offset (T.O.) for the permafrost condition is:

$$T.O. = T_{TOP} - MAST \quad [2.22a]$$

$$T.O. = \frac{DDT_s (rk-1)}{P} \quad [2.22b]$$

(equivalent to Romanovsky and Osterkamp 1995, eq. 12) The parameterization of the thermal offset was initially presented by Kudryavtsev (1977) as cited by Romanovsky and Osterkamp (1995) who provided a detailed derivation and comparison with field measurements. Independently and shortly after Romanovsky and Osterkamp (1995), Smith and Riseborough (1996) presented the same relation, determined through sensitivity analysis using a one-dimensional discontinuous finite element model, TONE (Goodrich 1978). Using TONE, a series of curves were fit through equilibrium values of the thermal offset and DDF_s , DDT_s , k_t and k_f were identified as the critical factors (Riseborough and Smith 1998). The parameterization of the thermal offset by way of the rk ratio is exact for equilibrium conditions.

Since rk is always less than or equal to 1.0, T.O. is zero when rk is 1.0 or negative when rk is below unity. A small rk value represents large differences between k_t and k_f , and leads to a large thermal offset. Conversely, T.O. is directly related to the surface temperature index for the thawing season (DDT_s). Irrespective of the rk value, increases in DDT_s cause increases in T.O.

2.7 The TTOP Model

The TTOP model combines parameterization of the surface and thermal offsets through n_f , n_t , and rk to relate T_A to T_{TOP} on an annual basis (Smith and Riseborough 1996). The TTOP model is:

$$T_{TOP} = MAAT + S.O. + T.O. \quad [2.23a]$$

$$T_{TOP} = \frac{(rk \times n_t \times DDT_A) - (n_f \times DDF_A)}{P} \quad [2.23b]$$

The TTOP model assumes the soil thermal regime is at equilibrium (Romanovsky and Osterkamp 1995). Based on the TTOP model, several climatic controls on T_{TOP} are implicit. First, permafrost is present (i.e. $T_{TOP} < 0$ °C) when the freezing potential exceeds the thawing potential (i.e. $rk \times n_t \times DDT_A < n_f \times DDF_A$). Second, T_{TOP} decreases with increasing latitude as DDF_A increases and DDT_A decreases. Third, T_{TOP} is likely to be negative if DDF_A greatly exceeds DDT_A regardless of n_f , n_t , or rk values (Smith and Riseborough 2002).

Localized control of T_{TOP} is represented by n_f , n_t , and rk . The sensitivity of T_{TOP} to n_t and rk increases with DDT_A , and to n_f increases with DDF_A . Smith and Riseborough (2002) proposed that the local distribution of ground temperatures is more sensitive to rk and n_f than to n_t because of the difference in their natural range. The influence of rk on T_{TOP} is on the order of three to four times – ranging from 0.3 to 1.0 (Riseborough and Smith 1998). The influence of n_f is on the order of ten times – ranging from 0.1 to 1.0 (Riseborough and Smith 1998). The influence of n_t on T_{TOP} is considerably less at only two times – generally ranging from 0.6 to 1.2 (Jorgenson and Kreig 1988; Klene *et al.* 2001). For this reason n_t is sometimes omitted from the TTOP Model (Henry and Smith 2001).

2.7.1 Limits of discontinuous permafrost

Smith and Riseborough (2002) examined the limits of the discontinuous permafrost zone using the TTOP model, values of MAAT, DDF_A , and DDT_A determined

from Canadian air temperature records, and n_f values estimated from snowfall data. The southern limit of permafrost is marked by a minimum T_{TOP} of 0 °C and is given by:

$$rk \times n_t \times DDT_A = n_f \times DDF_A \quad [2.24]$$

Thus for any given combination of climate and surface characteristics the minimum value of rk required to maintain permafrost, rk_{crit} can be calculated as:

$$rk_{crit} = \frac{n_f \times DDF_A}{n_t \times DDT_A} \quad [2.25]$$

Values of rk_{crit} near the observed southern extent of permafrost were close to the reasonable minimum value of rk for peat (0.30). Thus rk was deduced as the critical factor controlling the southern limit of permafrost (Smith and Riseborough 2002).

Smith and Riseborough (2002) marked the northern limit of discontinuous permafrost by a maximum T_{TOP} of 0 °C, and recognized that unfrozen ground is maintained along this boundary in bedrock, where rk equals 1.0 by:

$$0^\circ\text{C} = \frac{(n_t \times DDT_A) - (n_f \times DDF_A)}{P} \quad [2.26]$$

If values of n_t are assumed to be 1.0, the critical n_f required to maintain unfrozen ground in bedrock ($rk=1.0$) is:

$$n_{fcrit} = \frac{DDT_A}{DDF_A} \quad [2.27]$$

Based on the relation between n_f , MAAT and snow depth, n_{fcrit} was converted into a critical snow depth above which unfrozen ground was supported (Smith and Riseborough 2002). The snow depth required to maintain unfrozen ground increased with latitude as MAAT declined, especially at MAAT between -6 and -8 °C. This range in MAAT corresponds with the northern limit of discontinuous permafrost, treeline, and an abrupt decrease in snow-cover thickness. Thus n_f and the role of snow was deduced as the

critical factor controlling the southern limit of continuous permafrost (Smith and Riseborough 2002).

2.8 Concluding Remarks

The TTOP model elegantly summarizes the surface and thermal offsets to explore the equilibrium climate-permafrost relation at continental scale (Smith and Riseborough 2002). Since Henry and Smith (2001) used TTOP to reproduce the permafrost map of Canada, confidence in the model has increased, and interest is growing to use it in remotely mapping regional permafrost distribution (Juliussen and Humlum 2007). However, this chapter outlined three items that may jeopardize the operationalization of the TTOP model. First, the surface offset is thought to be controlled primarily by snow depth, which can be estimated from precipitation records and snow redistribution models. Here I have suggested that surface temperatures in winter can be greatly affected by active-layer conditions (Goodrich 1982; Smith and Riseborough 2002; Karunaratne and Burn 2004), which are both difficult to estimate remotely, and require *a priori* knowledge of permafrost presence to do so. Second, the parameterization of the thermal offset via the rk ratio has not been validated in the field under a variety of conditions. This validation is needed especially for the discontinuous permafrost zone, where rk has been proposed as the most important factor controlling permafrost distribution. Third, the TTOP model is intended to describe equilibrium conditions (Romanovsky and Osterkamp 1995; Smith and Riseborough 1996), and it may not be able to predict permafrost conditions under a transient climate. Finally, the TTOP model has been validated at small map scale, but site effects and varying moisture regimes may inhibit downscaling TTOP results to precise field conditions.

Chapter Three: Regional Setting and Study Period Climate

3.1 Introduction

Comparison of the climate-permafrost relation between the continuous and discontinuous permafrost zones requires a study area that spans these zones, and is not affected by mesoscale phenomena, such as coastal weather or orographic precipitation, and has consistent geology and geomorphic history, resulting in similar terrain characteristics. The Slave Geological Province, located in the northwest corner of the Canadian Shield, fulfills these criteria and has mines in construction, operation, or reclamation that provide access to remote sites. This chapter describes the regional characteristics of the Slave Province, and summarizes the climate during the study period.

3.2 The Slave Geological Province

The Slave Geological Province of the Canadian Shield covers about 230 000 km², and extends in a northeast direction from Great Slave Lake to Coronation Gulf (Figure 3.1). The climate-permafrost relation was examined at three locations in the Slave Province: Yellowknife, Colomac Mine, and the Ekati Diamond Mine. All three study areas are in the Northwest Territories. The City of Yellowknife (62° 28' N 114° 27' W), lies on the southern boundary of the Slave Province. The Colomac Mine is located 210 km north-northwest of Yellowknife. The Colomac gold mine (64° 12' N 116° 1' W) operated from 1989 to 1997, but is now being reclaimed by Indian and Northern Affairs Canada – Contaminated Sites Division. Ekati (64° 44' N 110° 36' W), Canada's first

Figure 3.1 Study areas Yellowknife, Colomac and Ekati within the Slave Geological Province, including Lupin Mine where climate Normals were recorded, and Daring Lake, where daily mean snow depths were recorded by Indian and Northern Affairs Canada – Water Resources Division.

diamond mine, is in the centre of the Slave Province 325 km north-northeast of Yellowknife. Ekati opened in 1998, and is operated by BHP Billiton.

3.2.1 Regional Geology of the Slave Province

The Slave Province is an Archean craton composed of three domains: the Anton microcontinent to the west, the younger Hackett River Arc to the east, and the Contwoyto central accretionary prism. Two thirds of the Province is comprised of granitoid rocks, while metaturbidites and metavolcanics make up 25% and 10% respectively (Padgham and Fyson 1992). Diamondiferous kimberlite pipes intruded into the central Slave Province between 45 and 75 million years ago (Nowicki *et al.* 2004).

The Slave Province has experienced multiple glaciations, but the current glacial features are attributed to the Late Wisconsinan Laurentide ice sheet, which retreated from the region approximately 9000 years ago (Kerr *et al.* 1997). Glaciation has resulted in gently undulating to moderately rugged topography with bedrock outcrops (Dredge *et al.* 1999). Glacial features create relief that is approximately 10-30 m, while river terraces are generally less than 10 m. Till, the dominant unconsolidated sediment is found in thin veneers, blanket deposits (including drumlins) up to 10 m thick, and hummocky deposits up to 30 m thick (Dredge *et al.* 1999). All tills are stony diamictons, but till derived from granitic rocks has a silty, sandy matrix while that from metasedimentary rocks has higher clay content (Dredge *et al.* 1999). Glaciofluvial deposits are commonly found in eskers, kames, and outwash plains throughout the area. Evidence of large-scale subglacial meltwater flow is abundant in the Slave Province and includes gravel bars, boulder lags, plunge pools, and corridors where till has been eroded (Rampton 2000).

The Slave Province is covered with lakes of various sizes. Small lakes occur due to glacial scouring, kettles, and poor drainage, while large lakes form in eroded faults and dykes. Great Slave Lake at the southern perimeter of the province covers 28 400 km² and is a remnant of post-glacial Lake McConnell. The Yellowknife study area lies along the Cameron River east of Yellowknife. Colomac lies within the Snare River watershed that also flows into Great Slave Lake. Ekati lies just north of Lac de Gras – the headwaters of the Coppermine River that flows north to the Arctic Ocean.

3.2.2 Climate of the Slave Province

The study areas in the Slave Province have an arctic continental climate due to their latitude and distance from moderating water bodies (Kendrew and Currie 1955). However, there are differences in mean annual air temperature (MAAT), and precipitation across the region. Lupin, a gold mine in Nunavut approximately 100 km northwest of Ekati (Figure 3.1), has a MAAT of -11.1 °C, and daily mean temperatures above 0 °C from June to September (Figure 3.2) (Environment Canada 2010). The mean monthly air temperature at Lupin is 11.5 °C for July, the warmest month, and -30.4 °C for January, the coldest month. Total annual precipitation at Lupin is approximately 300 mm, of which 45% arrives as snow. In contrast, Yellowknife is considerably warmer, with a MAAT of -4.5 °C, and daily mean temperatures above 0 °C from May to September (Figure 3.2) (Environment Canada 2010). The warmest and coldest months at Yellowknife are also July and January but have mean monthly air temperatures of 16.8 °C, and -26.8 °C respectively. Yellowknife receives slightly less precipitation than Lupin, with a mean annual total of 280 mm, and 40% falling as snow.

Figure 3.2 Mean monthly air temperatures (lines), and mean monthly precipitation (bars) (1971-2000) for the Yellowknife and Lupin Airports (Environment Canada 2010).

3.2.3 Permafrost in the Slave Province

Gradations in vegetation and climate accompany changes to permafrost conditions across the Slave Province. In the south, permafrost at Yellowknife is widespread but discontinuous (Heginbottom *et al.* 1995), and previously a mean annual ground temperature of -1.1 °C at 2.5 m depth was recorded (Brown 1960). Permafrost can reach depths of over 50 m in peatlands and spruce forest but is absent where bedrock is exposed; active-layer thickness ranges from 0.3 m to 1.3 m depending on organic cover (Brown 1973). Colomac also lies within the widespread discontinuous permafrost zone.

In contrast, permafrost is continuous north of treeline in the Slave Province. At Diavik Diamond Mine, 25 km south of Ekati, permafrost reaches depths of 250 m and the mean annual ground temperature at 20 m is between -3 °C and -6 °C (Hu *et al.* 2003). The active layer is thickest in bedrock, 5 m, and may be less than 15 cm thick in organic soil (Dredge *et al.* 1999). Most permafrost research in the Slave Province has focused on ground ice rather than ground temperatures (e.g. Wolfe *et al.* 1997; Dredge *et al.* 1999; Hu *et al.* 2003).

3.2.4 Vegetation in the Slave Province

Treeline crosses the Slave Province in a northwest-southeast direction, dividing it into the Western Taiga Shield and the Southern Arctic Ecozones (Wiken *et al.* 1993). The Western Taiga Shield lies south of treeline in the northern boreal forest which extends in NW-SE direction across Canada. Both Yellowknife and Colomac lie in the Western Taiga Shield Ecozone. Unlike the boreal forest, the subarctic taiga is characterised by open-canopy forests and shrublands. Upland areas of this ecozone consist of exposed bedrock and till veneers, and are colonized by a variety of lichens,

kinnikinnick (*Arctostaphylos uva-ursi*), and jack pine (*Pinus banksiana*). Low-lying depressions are often occupied by peatlands with black and white spruce (*Picea mariana* and *P. glauca*), tamarack (*Larix laricina*), Labrador tea (*Ledum groenlandicum*), and blueberry (*Vaccinium spp.*) (Wiken 1996). Although several different types of wetlands occur within the Western Taiga Shield, basin fens and bogs are most abundant (National Wetlands Working Group 1997). Areas disturbed by forest fires, which are common in this ecozone, are colonized by white birch (*Betula papyrifera*), trembling aspen (*Populus tremulus*), and balsam poplar (*Populus balsamifera*).

At the northern extent of the taiga lies the Forest-Tundra where trees are found in colonies which become increasingly smaller and more dispersed with latitude, and become surrounded by shrubland tundra (Timoney *et al.* 1992). The transition between continuous and discontinuous permafrost lies in the Forest-Tundra, as does the mean position of the Arctic Front in July (Bryson 1966). The precise relation between the northern extent of trees and discontinuous permafrost, and the Arctic Front is unknown, but the three phenomena are considered to be interrelated (Hare and Ritchie 1972).

The northern Slave Province lies in the Southern Arctic Ecozone which is characterised by continuous shrub tundra. Northern Labrador tea (*Ledum decumbens*) and dwarf birch (*Betula pumila*) are the common shrubs in this region but willows (*Salix spp.*) are found along streams and in wet areas. Bog cranberry (*Vaccinium vitis-idaea*) and dwarf bog rosemary (*Andromeda polifolia*) are the dominant short shrubs. Wetlands are colonized by *Sphagnum spp.* mosses, cotton-grass (*Eriophorum russeolum*), and other sedge species exclusively. *Aulacomnium turgidum* and *Rhacomitrium lanuginosum* mosses are also common in the area. Numerous species of lichens are found in the

Southern Arctic, but *Cladina Spp.* and *Cladonia Spp.* are the most abundant. The thickest deposits of peat are associated with ice-wedge polygons and sedge meadows, found in depressions underlain by fine-grained sediments that impede drainage.

3.3 Study Period Climate

The study period extended from September 2004 to September 2006. Climate conditions are presented for the study period and the previous year to acknowledge antecedent conditions. Air temperatures and rainfall totals measured at the Yellowknife and Ekati airports represent climate patterns in the Slave Province during the period of study. Snow depth was measured at the Yellowknife Airport but not at Ekati. However, snow depth was recorded by Indian and Northern Affairs Canada – Water Resources Division at Daring Lake. Located in the Southern Arctic Ecozone approximately 48 km west-northwest of Ekati, Daring Lake is a research facility operated by the Government of Northwest Territories.

Throughout the study period, daily mean air temperatures (T_{Ad}) at Ekati were consistently cooler than at Yellowknife, and T_{Ad} at the two locations generally covaried suggesting a common climate system (Figure 3.3). Annual mean air temperatures (T_{Aa}) at the Yellowknife and Ekati airports were calculated from T_{Ad} between 1 September and 31 August. At Yellowknife, T_{Aa} were slightly lower than normal in 2003-04 and 2004-05, at about -5 °C, but considerably higher than normal in 2005-06, at -0.9 °C (Table 3.1). Likewise at Ekati, T_{Aa} was similar for the first two years of the study, at -10.8 °C, and several degrees higher in 2005-06 at -6.5 °C (Table 3.1). The 2005-06 warm year represents a climatic anomaly analogous to the 2071-2100 climate warming scenario for the region (Canadian Climate Change Scenarios Network 2010).

Figure 3.3 Daily mean air temperatures (T_{Ad}) for the study period and the preceding year at the Yellowknife Airport (Environment Canada 2010) and Ekati Airport (BHP Billiton 2007). The daily mean temperatures have been smoothed with a 10-day running mean and symbols have been placed every 15 days.

Table 3.1 Annual mean air temperatures (T_{Aa}) at Yellowknife and Ekati Airports for the study period. Annual mean temperature calculated from daily mean temperatures from 1 September to 30 August. Climate data are courtesy of Environment Canada (2010) for the Yellowknife Airport, and BHP Billiton for the Ekati Airport. Mean annual air temperatures (1971-2000) for Yellowknife and Lupin are also included courtesy of Environment Canada (2010).

Ekati	T_{Aa} (°C)
Normal (Lupin) 1971-2000	-11.1
2003-04	-10.7
2004-05	-10.8
2005-06	-6.5
Yellowknife	
Normal 1971-2000	-4.5
2003-04	-4.9
2004-05	-5.1
2005-06	-0.9

The study period was comprised of two summer thaw seasons and two winter freezing seasons. Commencement of the freezing seasons and the termination of the thawing seasons were defined as the first day in autumn that daily mean air temperatures (T_{Ad}) were consistently below 0 °C. The end of the freezing seasons and the beginning of the thawing seasons were defined as the first day in spring that daily mean air temperatures rose above 0 °C. The air temperature-defined freezing and thawing seasons are congruent, with the last day of the freezing season directly preceding the first day of the thawing season and *vice versa*. It is important to note that the durations of sequential thaw and freezing (or sequential freezing and thawing) seasons rarely sum to 365, because seasons are defined by temperature rather than by calendar date.

3.3.1 Freezing seasons

The freezing seasons at Yellowknife were always shorter and warmer than those at Ekati to the north (Figure 3.3). The normal freezing season at Yellowknife lasts 200 days, extends from 12 October to 28 April, has 3424 °C-d air freezing degree-days (DDF_A), and 152 cm of snowfall (Table 3.2). At Yellowknife, freezing seasons 2003-04 and 2004-05 were both similar to normal conditions. At Ekati, these first two freezing seasons were also similar to each other with approximately 35% higher DDF_A than at Yellowknife. Freezing season 2005-06 was considerably warmer at both Yellowknife and Ekati. The mean freezing season air temperature (1 September to 30 April) was approximately 4 °C higher than the previous two years at both Yellowknife and Ekati.

During the study period there were distinct differences in snow cover north and south of treeline (Figure 3.4). Snow depth was considerably higher throughout the

Table 3.2 Descriptive statistics of the study's freezing seasons for Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports including the timing and duration of the seasons, mean winter air temperature temperatures (1 September to 30 April), freezing degree days (DDF_A), and median March snow depth. Snow data for Ekati was collected by Indian and Northern Affairs Canada – Water Resources Division at Daring Lake (48 km west-northwest of Ekati). Statistics for the normal freezing season (1971-2000) at Yellowknife Airport are also included.

Freezing Season	Start of Freezing Season	End of Freezing Season	Duration of Freezing Season (days)	Mean Winter Air Temperature (°C)	DDF _A (°C-days)	Median March Snow Depth (cm)
Ekati						
2003-04	14 October	31 May	230	-18.3	4924	9
2004-05	29 September	28 May	242	-18.7	4824	24
2005-06	18 September	5 May	230	-14.4	3587	20
Yellowknife						
Normal 1971-2000	12 October	28 April	200	-13.1	3424	39
2003-04	27 October	13 May	212	-12.4	3466	36
2004-05	29 September	7 April	191	-13.3	3366	55
2005-06	27 September	8 April	194	-8.5	2278	32

Figure 3.4 Daily mean snow depth at Yellowknife Airport (Environment Canada 2010) and Daring Lake for the study period freezing seasons (2004-05 and 2005-06) and the preceding winter (2003-04). Daring Lake is located 48 km west-northwest of Ekati. Daring Lake snow data were collected by Indian and Northern Affairs Canada – Water Resources Division.

winter at Yellowknife than on the tundra at Daring Lake. Snow accumulation began at Yellowknife either before or at the same time as at Daring Lake, despite the earlier start to the freezing season on the tundra. Furthermore, snow depth at Yellowknife tended to increase rapidly in the early winter, but it remained less than 20 cm through to mid-December at Daring Lake. Though mid-winter snow depths were unavailable at Daring Lake due to instrument failure, maximum snow depths in April were lower than at Yellowknife in 2003-04 and 2004-05 and similar in 2005-06. However, the snow-depth-days – the summation of daily mean snow depth over the freezing season – were higher at Yellowknife due to the thick snow covers established early in the season. For the study period, snowmelt occurred predominantly in April at Yellowknife, and in May at Daring Lake.

3.3.2 Thaw seasons

In accordance with the freezing seasons, the thaw seasons at Yellowknife were always longer and warmer than at Ekati (Figure 3.3). The normal thaw season at Yellowknife lasts 165 days, extends from 29 April to 11 October, has 1769 °C-d, and 164 mm of rain (Table 3.3). Thaw season 2004 was slightly cooler than normal at Yellowknife. In 2004, air thawing degree days (DDT_A) for Yellowknife and Ekati were 1509 and 918 °C-d respectively. The 2004 thaw season was particularly dry, with both Yellowknife and Ekati receiving less than 70 mm of rain by the end of August (Figure 3.5). The following 2005 thaw season was similar to 2004 in terms of air temperature, but Ekati received slightly more rain, and Yellowknife received 115% more rain than the previous year. Thaw season 2006 was longer and considerably warmer at both Yellowknife and Ekati. The mean thaw season air temperature (1 May to 31 August)

Table 3.3 Descriptive statistics of the study's thawing seasons for Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports including the timing and duration of the seasons, mean summer air temperature temperatures (1 May to 31 August), thawing degree days (DDT_A), and rainfall. Rainfall data for Ekati were collected by Indian and Northern Affairs Canada – Water Resources Division at Daring Lake (48 km west-northwest of Ekati). Statistics for the normal thawing season (1971-2000) at Yellowknife Airport are also included.

Thaw Season	Start of Thaw Season	End of Thaw Season	Duration of Thaw Season (days)	Mean Summer Temperature (°C)	DDT _A (°C-days)	Rainfall (cm)
Ekati						
2004	1 June	28 September	120	4.2	918	106
2005	29 May	17 September	112	4.8	910	118
2006	6 May	27 September	145	8.9	1271	176
Yellowknife						
Normal 1971-2000	29 April	11 October	165	12.5	1769	164
2004	14 May	28 September	138	10.0	1509	65
2005	8 April	26 September	172	11.1	1590	240
2006	9 April	8 October	183	14.1	2101	188

Figure 3.5 Monthly total rainfall at Yellowknife (Environment Canada 2010) and Ekati (BHP Billiton) Airports for the study period thawing seasons (2005 and 2006) and the preceding summer (2004).

was approximately 4 °C higher than the previous two years at both Yellowknife and Ekati, and degree-day totals were approximately 30 % greater. Rainfall in 2006 was similar at Yellowknife and Ekati at 188 and 176 mm respectively. The relation between the monthly rainfall at Yellowknife and Ekati was inconsistent due to local convective storms.

3.4 Summary

The Slave Geological Province is an ideal location to examine the climate-permafrost relation. Within the study region, the geological and geomorphological setting is uniform, the three study areas lie within the same climate system, and both tundra with continuous permafrost and taiga with discontinuous permafrost exist. This study is particularly useful given the considerable industrial development and limited knowledge of permafrost conditions in this area.

The study period extended from September 2004 to October 2006, comprising two full years. Air temperatures for the first year of the study and the preceding year were remarkably close to normal at both Yellowknife and Ekati. In 2005-06, the second year of the study, air temperatures were about 4 °C higher in both the freezing and thawing seasons. This warm year is analogous to the climate warming scenario for the region in 2071-2100.

Chapter Four: Physical Characteristics of the Sites

4.1   Introduction

To examine the climate-permafrost relation in the Slave Province, several study sites were selected within the Yellowknife, Colomac and Ekati study areas.

Instrumentation was established at these study sites to measure air, surface and ground temperatures (Figure 1.4). This chapter describes the physical characteristics of the instrumented sites in terms of vegetation, soil, snow cover, and active-layer thickness.

The methodology used to characterise the sites is explained throughout the chapter.

4.2   Site selection

The instrumented sites were established at multiple locations within the Yellowknife, Colomac and Ekati study areas, in low-lying peatlands classified as bogs (National Wetlands Working Group 1997). Bogs represent a common terrain unit in the Slave Province, where climate-permafrost relations are affected by latent heat and seasonal changes in thermal conductivity. The instrumented bogs at Yellowknife and Colomac were treed basin bogs (National Wetlands Working Group 1997), and were chosen because they support permafrost in the discontinuous permafrost zone. At Ekati, the instrumented sites were located within high-centred ice-wedge polygons, and were chosen because they are the most comparable with the basin bogs at Yellowknife and Colomac. Both sets of features had fibrous organic layers at the ground surface and substantial soil moisture. At Yellowknife and Ekati, instrumented sites were also established in glaciofluvial mineral deposits.

At Yellowknife, five study sites were chosen east and north of the city (Figure 4.1). Sites Y-1 through Y-4 were treed basin bogs. Sites Y-1, Y-2, and Y-3 were 70, 50, and 30 km east of Yellowknife along the Ingraham Trail respectively, while Y-4 is 10 km north of the city near Ryan Lake. Y-5 is a treed glaciofluvial sand deposit near Reid Lake approximately 60 km east of Yellowknife. Three study sites were chosen at Colomac along the Kim Cass Road southwest of the mine (Figure 4.2). Sites C-1 and C-2 were basin bogs with relatively few trees, and were located 14 and 10 km south of Colomac respectively. Site C-3 was approximately 5 km from the mine, in a treed bog similar to those at Yellowknife. Five study sites were chosen at Ekati (Figure 4.3). Sites E-1, E-3, and E-4 were situated within isolated pockets of high-centred ice-wedge polygons. Site E-1 was located 4 km north of the Ekati camp, while sites E-3 and E-4 were located along the Misery Road 8 km and 24 km respectively southeast of the Ekati Camp. Sites E-2 and E-5 were situated adjacent to and on top of Airport Esker respectively, approximately 5 km south of the Ekati Camp.

4.3 Vegetation

Vegetation must be considered when examining the surface thermal regime because it affects radiation input and snow distribution. Vegetation surveys examined trees, tall shrubs and ground cover in a 10 x 10 m quadrat centred on each instrumented site. To determine the characteristics of the arboreal vegetation at the Yellowknife and Colomac instrumented sites, the mature trees, saplings and seedlings in the survey quadrat were tallied for each species. Mature trees were defined as those higher than 2 m, saplings between 0.5 and 2.0 m, and seedlings shorter than 0.5 m. A sample of tree heights at the instrumented sites was obtained by measuring the height of trees

Figure 4.1 The Yellowknife study area, showing locations of the five study sites. The Yellowknife Airport is located within 5 km of city-centre.

Figure 4.2 The Colomac study area, showing locations of the three study sites.

Figure 4.3 The Ekati study area, showing locations of the five study sites. The Ekati Airport is located between the Ekati camp and E-2.

within 0.5 m of a 10 x 10 m cross-transect centred on the instrumentation. The instrumented sites at Yellowknife and Colomac were categorized according to their canopy cover. Open canopy sites had few if any surrounding trees and were shaded only early and late in the day. Conversely, closed canopy sites had multiple overhanging tree branches that provided shading continuously throughout the day, and intercepted snow in winter.

The tall shrub component of the vegetation consisted of multi-stemmed woody plants greater than 15 cm tall, while the ground cover component consisted of shorter shrubs, vascular non-woody herbs, mosses, lichens, and non-vegetative cover such as exposed soil, and leaf litter. At the instrumented sites, species of tall shrubs, and ground cover, as well as non-vegetative components, were identified and coverage was visually assessed. Coverage, defined as the vertical crown area projection in the quadrat (Mueller-Dombois and Ellenberg 1974), was summarized using a cover-scale class. The cover-scale used is a modification of the Braun-Blanquet cover scale, which provides a subjective and qualitative but rapid technique to assess vegetation coverage (Goldsmith *et al.* 1986). In addition to coverage, mean height and diameter at stem base (DSB) were determined for the tall shrubs. Appendix A contains results of the vegetation survey including tall shrub heights and DSB, and vegetation and ground coverage.

4.3.1 Yellowknife Vegetation

The Yellowknife study sites had similar vegetation characteristics. All of the instrumented sites were predominantly colonized by *Picea mariana*, of different heights. The trees were randomly distributed, with small clusters of trees, and open areas devoid of trees. *Ledum groenlandicum* had high coverage at all the Yellowknife sites, and was

generally about 30 cm tall. *Chamaedaphne calyculata* and *Alnus crispa* were also present at some sites. The ground cover at the Yellowknife sites consisted of a small number of short shrubs and herbs, most notably *Andromeda polifolia*, *Vaccinium vitis-idaea*, *Empetrum nigrum*, and *Rubus chamaemorus*. Mosses and lichens covered large areas, while needle and leaf litter, and layer branches were common but did not cover large areas at most sites. There was little difference in the ground cover component of the vegetation among the instrumented sites at Yellowknife. All of the sites except Y-5a were covered with poorly defined peat mounds colonized by short shrubs. The ground at Y-5a was flat, covered with lichen, needle and leaf litter, and was the only site at Yellowknife with exposed mineral soil, and almost no moss.

At Yellowknife, instrumented sites were chosen to capture differences in canopy cover, thus sites were established in open areas with few trees, and at points shaded by trees. Instrumented sites Y-1a, Y-2a, Y-2b, Y-3a and Y-4a were classified as open canopy sites, (e.g. Figure 4.4a) because the surrounding trees were short and few in number. Instrumented sites Y-2c, Y-3b, Y-4b, and Y-5a were designated as closed canopy sites, (e.g. Figure 4.4b) because they were situated in dense clusters of mature *Picea mariana* trees, with multiple overhanging branches that provided shading throughout the day, and intercepted snow in winter. Compared to the open canopy sites, the closed canopy sites had more seedlings, saplings, and mature trees, and had higher maximum tree heights.

Figure 4.4 Photograph examples of Yellowknife instrumented sites with (a) open canopy, Y-4a; and (b) closed canopy, Y-3b.

4.3.2 Colomac Vegetation

Vegetation differed among the three Colomac study sites. Two of the study sites at Colomac, C-1 and C-2 were generally open and had few trees while the trees at C-3 were taller and more dense. *Picea mariana* was the only tree species present at the Colomac study sites. *Ledum groenlandicum* was present at all three sites, and had a similar height and coverage to that observed at Yellowknife. *Betula pumila* shrubs between 40 and 70 cm tall were present at sites C-1 and C-2, the open study sites. *Salix arbusculoides* shrubs, close to one metre tall, were found at all three instrumented sites at C-3. The ground cover at all three Colomac study sites included *Vaccinium vitis-idaea*, *Rubus chamaemorus*, *Dicranum polysetum*, and *Cladina mitis*, and most of the sites had *Sphagnum spp.* mosses. Grasses between 30 and 40 cm tall were present at the instrumented sites at study site C-3, and at C-2a. Similar to Yellowknife, the instrumented sites at Colomac were located in open areas and within clusters of trees. Instrumented sites C-1a, C-1b, C-2a, C-3a, and C-2b had open canopies (Figure 4.5 (a)), while sites C-1c, C-3b, and C-3c had closed canopies (Figure 4.5 (b)).

4.3.3 Ekati Vegetation

All of the Ekati study sites were open as there were no trees or shrubs greater than 40 cm. Study sites E-1, E-3, and E-4 were located in high-centred ice-wedge polygons, and had earth hummocks approximately 1 m apart and 0.3 m high. Site E-2 also had hummocks but they were irregularly shaped and of different sizes. The ground was even on top of the esker at site E-5. *Ledum decumbens*, rather than *Ledum groenlandicum*, was present at most of the instrumented sites, and was less than 15 cm tall. *Betula pumila* was the only tall shrub present at the Ekati study sites,

Figure 4.5 Photograph examples of Colomac instrumented sites with (a) open canopy, C-1a; and (b) closed canopy, C-3c.

but it was generally shorter than it was at Colomac. Except for E-5a, which was completely free of any vegetation, the ground cover was similar among the Ekati instrumented sites. *Andromeda polifoli*, *Empetrum nigrum*, *Vaccinium vitis-idaea*, and *Eriophorum russeolum* were present at all the instrumented sites. There were many different species of moss and lichens at the instrumented sites but *Aulacomnium turgidum*, *Rhacomitrium lanuginosum*, and *Cladina mitis* were particularly common. *Betula* leaf litter was ubiquitous. Figure 4.6 presents photos of sites E-1, which was located within a peat plateau bog, and E-5 which was located on top of the esker.

4.4 Soil

Soil characteristics at the instrumented sites were examined as they control the soil's thermal properties. In September 2005 and 2006, soil samples were collected at the instrumented sites from soil pits dug to the base of the active layer. The soil pits were dug approximately 5 m from the temperature sensors, to capture soil characteristics and minimize disturbance. At the majority of the sites, samples were obtained from two pits per instrumented site at 5 and 10 cm from the surface and at 10 cm intervals to the base of the active layer. However, at several Colomac sites only 3 samples per site could be collected due to time constraints. These three samples were collected 10 cm from the surface, and in the middle and base of the active layer. Depth to supra permafrost water table was noted if present when the samples were collected.

Johansen's (1975) model of soil thermal conductivity requires the volumetric fractions of the soil constituents, thus soil samples of known volume were collected by cutting rectangular blocks from the wall of the soil pit at regular intervals, and measuring the dimensions of the blocks. Poorly consolidated mineral soils were sampled using

Figure 4.6 Photograph examples of Ekati instrumented sites at (a) high centred polygon, E-1a; and (b) atop esker, E-5a.

a cylinder of known volume. The samples were weighed, then dried in an oven for 24 hours at 105 °C, and reweighed to determine the volumetric water content. Organic content and soil texture were determined for soils with a mineral component. The percent organic content was estimated through loss on ignition. Prew weighed samples were heated to 550° C for four hours to burn off the organic material, and the samples were subsequently weighed to determine the loss of organic material. Soil texture was determined using the Microtrac^(R) 3500 Particle Size Analyzer.

Figure 4.7 and 4.8 present bulk density and volumetric soil content at three layers in the active layer – the Surface Layer, the Middle Layer, and the Base Layer. At sites where samples were collected from multiple depths, samples were chosen to represent these three layers. The Surface Layers, sampled at 5 cm depth, were always composed exclusively of organic material, and had low bulk densities and moisture contents. Thicknesses of the Middle Layers ranged from 10 to 40 cm, and were determined from field observations of soil moisture, structure, and composition. These Middle Layers typically had higher soil moistures and bulk densities than the Surface Layers, and were usually unsaturated. Thicknesses of the Base Layers were also determined from field observations of soil properties especially soil water. These layers captured saturated conditions if present, and were 10 to 60 cm thick. Appendix A contains details of the 2006 soil survey.

4.4.1 Yellowknife Soils

The Yellowknife study sites, Y-1, Y-2, Y-3, and Y-4, had organic peat soils, and mineral material was not encountered in the active layer. The active layer at most of these sites was composed of poorly decomposed Sphagnum mosses, but Sphagnum mosses

Figure 4.7 Examples of soil bulk density values for the Surface, Middle and Base layers of the active layers at Yellowknife, Colomac, and Ekati. Soil samples for the (a) Surface Layer were taken at 0 to 5 cm depth; (b) Middle Layer were taken at 10 to 40 cm depth; (c) Base Layer were taken at base of the active layer.

Figure 4.8 Examples of soil volumetric water content for the Surface, Middle and Base layers of the active layers at Yellowknife, Colomac, and Ekati. Soil samples for the (a) Surface Layer were taken at 0 to 5 cm depth; (b) Middle Layer were taken at 10 to 40 cm depth; (c) Base Layer were taken at base of the active layer.

were not abundant on the surface. The bulk density generally increased with depth, from less than 0.10 g cm^{-3} at the surface to between 0.10 and 0.23 g cm^{-3} at depths below 20 cm. These bulk density values are comparable to published values summarized by Walmsley (1973), and recently reported by Van Asselen and Roosendaal (2009). Volumetric water contents generally increased with depth. In 2006 the instrumented sites were wetter than the previous summer and a perched water table was often found between 5 and 45 cm below the surface. In both years, instrumented sites Y-2a and Y-2b were particularly wet, while Y-3 was relatively dry. Unlike the other sites at Yellowknife, site Y-5a was located in a glaciofluvial sand deposit and did not have a well developed organic layer apart from the lichens that covered the surface. The sand was well sorted, had a low organic content, a mean bulk density of 1.4 g cm^{-3} , and was dry with volumetric water content less than 0.10 mL cm^{-3} . This site, Y-5a, was remarkably different from the other Yellowknife sites.

4.4.2 Colomac Soils

The Colomac study sites were all located in peatlands and had similar soil characteristics to the peatland sites at Yellowknife, except that undecomposed peat was more prominent. At all three Colomac study sites the bulk density and volumetric water content increased with depth, and active layers were wetter in 2006 than in 2005. Instrumented sites at C-3 had perched water tables in 2006. Ice-rich, fine-grained mineral soil was found below the active layer between 60 and 100 cm at instrumented sites C-1a and C-1c.

4.4.3 Ekati Soils

The active layers at the Ekati sites had a substantial mineral component. The organic layers at the Ekati sites were generally 5 to 10 cm thick, and had similar bulk densities to the surface peat at Yellowknife and Colomac. The volumetric water content tended to be lower in the organic layer than the underlying mineral material. Sites E-1, E-3, and E-4 were located amongst ice-wedge polygons, and active-layer soil was finer than medium sand. The bulk densities of the mineral material were generally less than 1.0 g cm^{-3} , because the soils were cryoturbated and had a substantial organic content. The base of the active layer regularly had a higher moisture content. When these study sites were instrumented, excess ice was encountered beneath the active layer to 100 cm depth. Study sites E-2 and E-5 had much different soil characteristics than the other sites as they were associated with esker sediments. The soil at site E-2 consisted of gravel and cobbles supported in a matrix of poorly sorted, coarse-grained sand with small amounts of silt and clay. The soil was unconsolidated and wet. Bulk density and volumetric water content measurements for the sites at E-2 are underestimated due to the cobbles. In contrast, site E-5 was located on top of the esker and had dry, well-sorted sand with little organic matter.

4.5 Snow

Snow characteristics were examined in April 2005 and 2006 to assess their influence on the ground thermal regime. An additional snow survey was conducted at Yellowknife and Colomac in January 2006. At the study sites, snow depths were measured using a graduated steel rod along two 70 m transects at 5 m intervals, for a total of 30 measurements per study site. Snow depths were also measured nine times along a

4x4 m cross-transect centered on the instrumented site, to determine the snow characteristics associated with the ground temperature measurements. Snow density was measured at the instrumented sites in a pit excavated to the base of the snow pack. Samples of known volume were extracted at 5 or 10 cm intervals from the wall of the snow pit and weighed in the field to determine density. Snow density was measured at Ekati in April 2006 only, but similar conditions were observed in 2005. The non-parametric Mann-Whitney U statistic was used to determine if differences in snow cover and active-layer thickness existed among the study areas, and the Wilcoxon Matched-Pairs Signed Rank statistic (W) was used to determine if differences existed between the two years (Hammond and McCullagh 1978).

4.5.1 Yellowknife and Colomac Snow

At Yellowknife and Colomac south of treeline, snow covers were thick, had low density, and snow surfaces were even (Figure 4.9a). Snow densities at Yellowknife and Colomac were between 0.15 and 0.35 g cm⁻³ throughout the snow pack with no distinct ice layers (Figure 4.10). Depth hoar layers between 5 and 15 cm thick with 3 to 5 mm crystals were regularly observed at the base of the snow pack and were more pronounced in April.

The snow-depth measurements along transects through the study sites were grouped by study area (Figure 4.11). In April, snow depths at the study sites south of treeline, were generally greater than 40 cm in both 2005 and 2006. Colomac had thicker snow covers than Yellowknife in April 2005 and 2006 (Table 4.1), as well as in January 2006 (U=11389.5; n=150 & 90; p<0.001). Snow depth measurements at Yellowknife in spring 2005 and 2006 may be underestimated due to thawing conditions, which

Figure 4.9 Photograph examples of snow conditions south of treeline at (a) Colomac study site C-2, and north of treeline at (b) Ekati study site E-2. Photos taken April 2005.

Figure 4.10 Snow density through the snow pack at Yellowknife (Y-3a), Colomac (C-3a) and Ekati (E-2a) in (a) April 2005, (b) January 2006, and (c) April 2006. Snow density at Ekati was only measured in April 2006.

Figure 4.11 Snow depths from transects through the study sites at Yellowknife, Colomac and Ekati. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths. Sample size ranged from 60 to 120. Snow surveys were not completed at Ekati in January 2006, and measurements at Ekati esker site E-5 in April 2006 were excluded because of their unusually low values.

Table 4.1 Mann-Whitney U statistics comparing snow depth measured along transects through the study sites at Yellowknife, Colomac, and Ekati in April 2005 and 2006. Significant differences with p values ≤ 0.05 are in bold.

2005			2006		
Transect Snow Depth	Colomac (n=90)	Yellowknife (n=120)	Transect Snow Depth	Colomac (n=90)	Yellowknife (n=120)
Ekati (n=120)	U=10611 p<0.001	U=133360 p<0.001	Ekati (n=60)	U=5386.5 p<0.001	U=5951 p<0.001
Yellowknife (n=120)	U=9532 p<0.001	X	Yellowknife (n=120)	U=10693.5 p<0.001	X

could have contributed to the significant differences in April snow depth found between Colomac and Yellowknife. However, the survey in January 2006 also suggests that Colomac had thicker snow cover than Yellowknife.

Snow depth varied between the instrumented sites at Yellowknife (Figure 4.12) and Colomac (Figure 4.13) depending on canopy cover. At Yellowknife, instrumented sites with closed canopies had significantly thinner snow covers than open sites in April 2005 ($U=1558.5$; $n=45$ & 36 ; $p<0.001$), January 2006 ($U=1479$; $n=45$ & 36 ; $p<0.001$), and April 2006 ($U=1273.5$; $n=45$ & 36 ; $p<0.001$). Similarly at Colomac, instrumented sites with closed canopies had significantly thinner snow covers than open sites in April 2005 ($U=983$; $n=45$ & 27 ; $p<0.001$), January 2006 ($U=783$; $n=45$ & 27 ; $p=0.0207$), and April 2006 ($U=896$; $n=45$ & 27 ; $p<0.001$). Median snow depth values in April at the instrumented sites were not significantly different between the two years at either Yellowknife ($W=-6$; $n=8$; $p=0.344$), or Colomac ($W=11$; $n=8$; $p=0.141$).

4.5.2 Ekati Snow

North of treeline at Ekati, snow covers were thin, and snow surfaces were wind-packed – considerably different than south of treeline (Figure 4.9b). The snow pack at Ekati consisted of high-density wind-packed layers above low-density depth hoar with crystals up to 5 mm. Snow density varied through the snow pack at Ekati (Figure 4.10). In April 2005 and 2006, snow covers at Ekati were generally less than 40 cm (Figure 4.11), and significantly thinner than at Yellowknife and Colomac (Table 4.1). Median snow cover at the Ekati instrumented sites ranged between 24 and 39 cm in both April 2005 and 2006, with snow cover at esker sites E-5a and E-5b between 0 and 22 cm

Figure 4.12 Snow depths at Yellowknife instrumented sites in (a) April 2005; (b) January 2006; and (c) April 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and $n=9$.

Figure 4.13 Snow depths at Colomac instrumented sites in (a) April 2005; (b) January 2006; and (c) April 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and n=9.

in April 2006 (Figure 4.14). There was no significant difference in median April snow depth at the Ekati instrumented sites between the two years ($W=6$; $n=7$; $p=0.250$).

4.6 Active-Layer Thickness

The thickness of the active layer was measured by probing to resistance with a graduated steel rod. To determine the active-layer thickness of the study sites, measurements were taken at 5 m intervals along a 115 m transect for a total of 24 points. The transect cross-cut the range of terrain and vegetation characteristics present at a study site. Active-layer thickness was also measured nine times along a 4 x 4 m cross-transect centered on the instrument site, to determine the active-layer thickness associated with the ground temperature measurements. At site E-2 where the soil contained large amounts of gravel and cobbles, two pits were dug to the top of permafrost, to verify the measurements of active-layer thickness taken by probing. Active-layer measurements were not possible at site Y-5 because of the absence of permafrost, and at E-5 because the active layer was too thick to determine by probing or digging. Frost-table depth were measured in 2005 and 2006 slightly before the end of the thaw season. At Yellowknife and Ekati, measurements of frost-table depth in both 2005 and 2006 were made after at least 90% of the DDT_A for the season had accumulated (Table 4.2). Conversely, the frost-table depth measurements made at Colomac in 2005 were taken between 18 and 22 August, after only 80% of the total DDT_A had accumulated. These early measurements may have underestimated the true active-layer thickness at Colomac in 2005.

Figure 4.15 presents active-layer thicknesses measured along the transects in each study site. Active-layer thicknesses were significantly different at the three study

Figure 4.14 Snow depths at Ekati instrumented sites in (a) April 2005; and (b) April 2006;. Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and $n=9$.

Table 4.2 Timing of active-layer thickness measurements with respect to end of the thaw season, percent of thaw season lapsed, and percent of thawing degree-days ($\sqrt{\text{DDT}_A}$) at time of measurement. Measurements of active-layer thickness at Colomac in 2005 were slightly earlier than at Yellowknife and Ekati.

	Date of Active Layer Measurement	End of Thaw Season	Percent of Season Lapsed at Time of Measurement	Percent of $\sqrt{\text{DDT}_A}$ Accumulated at Time of Measurement*
2005				
Ekati	3-4 September	16 September	92	97
Colomac	18-22 August	21 September	62	89
Yellowknife	7-8 September	26 September	78	97
2006				
Ekati	4-5 September	27 September	83	95
Colomac	19-20 September	5 October	85	98
Yellowknife	12-16 September	7 October	80	96

* According to Stefan Solution (Andersland and Ladanyi 2004, p.61) square-root of air thawing degree-days at time of measurement divided by square-root of total air thawing degree-days

Figure 4.15 Active-layer thicknesses at the study sites for summers 2005 (dark grey) and 2006 (light grey). Box and whisker plots present median, maximum, minimum and upper and lower quartile snow depths, and $n=24$.

areas in 2005 (Table 4.3), with median values of 50 cm at Yellowknife, 47 cm at Colomac, and 36 cm at Ekati measured along the transects through the study sites. The following year active layers were thicker at all three study areas with higher minimum, maximum, and median values measured at every study site (Figure 4.15). In this second year, median active-layer thicknesses were 60 cm at Yellowknife, 58 cm at Colomac, and 44 cm at Ekati, and were not significantly different at Yellowknife and Colomac (Table 4.3). The difference in active-layer thickness at Yellowknife and Colomac in 2005 may be because active-layer thicknesses were measured too early in the season (Table 4.2). The thicker active layers in 2006 across all three study areas corresponded to higher air temperatures that year (Table 3.1 and Figure 3.2). At Yellowknife, the active layers were thicker in both years at Y-2 and Y-4, than at Y-1 and Y-3. At Colomac, the active layers at C-1 and C-2, were similar to each other, and thicker than C-3. The sites located among ice-wedge polygons, E-1, E-3 and E-4, had similar active layer thicknesses, while those associated with esker material, E-2 and E-5, had considerably greater thaw depth.

4.6.1 Yellowknife Active-Layer Thickness

At Yellowknife in 2005, median active-layer thicknesses at the instrumented sites ranged from 42 to 70 cm (Figure 4.16), and were significantly thinner at closed canopy sites than open sites ($U=997$; $n=45$ & 27 ; $p<0.001$). The following year, median active-layer thicknesses were significantly higher ($W=18$; $n=8$ & 8 ; $p=0.004$), and active layers at closed canopy sites continued to be significantly thinner than open sites ($U=1020$; $n=45$ & 27 ; $p<0.001$). Active layers may have been thicker in the second year at the instrumented sites due to disturbance of temperature installations, however the thicker active layers measured in 2006 along the transects at the study sites

Table 4.3 Mann-Whitney U statistics comparing active-layer thickness measured along transects through the study sites at Yellowknife, Colomac, and Ekati in 2005 and 2006. Significant differences with p values ≤ 0.05 are in bold.

2005			2006		
Transect Active-layer thickness	Colomac (n=72)	Yellowknife (n=96)	Transect Active-layer thickness	Colomac (n=72)	Yellowknife (n=96)
Ekati (n=96)	U=4931.5 p<0.001	U=7252.5 p<0.001	Ekati (n=96)	U=4888 p<0.001	U=7048.5 p<0.001
Yellowknife (n=96)	U=4642.5 p<0.001	X	Yellowknife (n=96)	U=3803 p=0.131	X

Figure 4.16 Active-layer thicknesses at Yellowknife instrumented sites for summers 2005 (dark grey) and 2006 (light grey). Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and $n=9$.

suggests that thickening was independent of disturbance. Active layers at instrumented sites Y-3a and Y-3b were only slightly thicker in 2006. Both these sites had peat soils with low soil moisture which could have made these sites less susceptible to interannual differences in air temperature. The range of active-layer thicknesses were small at the closed canopy instrumented sites and the thickness increased only slightly in 2006, while at the open sites active-layer thicknesses high wide ranges and were considerably thicker in 2006.

4.6.2 Colomac Active-Layer Thickness

Median active-layer thicknesses at the Colomac instrumented sites ranged from 37 to 55 cm in 2005 (Figure 4.17), and were significantly thicker the following year, ranging from 46 to 85 cm ($W=18$; $n=8$ & 8 ; $p=0.004$). While the difference in active-layer thickness between the two years at Colomac could have resulted from differences in measurement timing, the higher air temperatures and thicker active layers at Yellowknife and Ekati that year suggest that active layers at Colomac were probably thicker as well. Like Yellowknife, the closed canopy instrumented sites at Colomac had significantly thinner active layers than the open sites in both 2005 ($U=851.5$; $n=8$ & 8 ; $p=0.002$), and 2006 ($U=800.5$; $n=8$ & 8 ; $p=0.012$). The active layer at C-3c was thicker than the adjacent instrumented sites in C-3, and was considerably thicker in 2006. Permafrost at this site may have been disturbed by the installation.

4.6.3 Ekati Active-Layer Thickness

Active-layer thicknesses at the Ekati instrumented sites differed between the sites located within high-centred ice-wedge polygons and those associated with esker material (Figure 4.18). Active layers at instrumented sites located at E-1, E-3,

Figure 4.17 Active-layer thicknesses at Colomac instrumented sites for summers 2005, and 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and $n=9$.

Figure 4.18 Active-layer thicknesses at Ekati instrumented sites for summers 2005, and 2006. Box and whisker plots present median, maximum, minimum and upper and lower quartile active-layer thicknesses, and n=9.

and E-4, in high-centred ice-wedge polygons, were similar to each other with median active-layer thicknesses ranging from 29 to 35 cm in 2005. Instrumented sites E-2a and E-2b, located beside an esker, had coarse-grained soils and active layers that were twice as thick as those in the high-centred ice-wedge polygons. The active layers at sites E-5a and E-5b were too thick to determine their depth by probing or excavating, and were at least 150 cm thick in September 2005 when the instruments were installed. In 2006, active layers at the Ekati instrumented sites were significantly thicker than the previous year ($W=14$; $n=7$ & 7 ; $p=0.008$).

4.7 Summary

The instrumented sites at Yellowknife and Colomac were located in treed basin bogs with open and closed canopy stands of *Picea mariana*. The active layers were composed of peat and were often saturated, especially in 2006. South of treeline, snow cover at the end of winter, and active-layer thicknesses at the end of summer were generally greater than 40 cm. The Ekati instrumented sites north of treeline had remarkably different physical characteristics. The sites were located in low-lying areas of high-centred ice-wedge polygons dominated by *Ledum decumbens* and *Betula pumila*. Active layers at Ekati were generally 40 to 60 cm thick, and composed of cryoturbated mineral soil underlying a 10 cm thick organic layer. The snow cover on the tundra was wind-packed and thinner than south of treeline. Instrumented sites Y-5a at Yellowknife and E-5a and E-5b at Ekati were considerably different from the other sites and had dry sandy soils with no organic layers.

Chapter Five: Air, Surface and Ground Temperatures in the Slave Geological Province

5.1 Introduction

The Slave Province, as described in Chapter Three, is bisected by treeline with discontinuous permafrost and the Western Taiga Shield Ecozone to the southwest, and continuous permafrost and the Southern Arctic Ecozone to the northeast. Yellowknife and Colomac lie south of treeline, and Ekati lies north of treeline. Chapter Four described the physical characteristics at the instrumented sites within the three study areas. The sites near Yellowknife and Colomac had thicker snow packs, active layers, and organic covers than at Ekati. The annual mean air temperatures (T_{Aa}) were near normal in 2004-05 and more than 4 °C higher the following year.

This chapter examines the variability of air, surface, and ground temperatures within and among the three study areas. Air temperature (T_A) and surface temperature (T_S) are examined on a daily, seasonal, and annual basis, and ground temperatures at the top of permafrost (T_{TOP}) on an annual basis (Appendix Tables B.1 to B.4). The following research questions are addressed in this chapter:

- (1) What is the variation in air temperature within and among the three study areas on a daily, seasonal, and annual basis?
- (2) What are the differences in air temperature between the study sites and the adjacent airports at Yellowknife and Ekati?
- (3) What is the variation in surface temperature at the organic instrumented sites within and among the three study areas on a seasonal and annual basis?

- (4) What is the variation in ground temperature at the organic instrumented sites within and among the three study areas on an annual basis?
- (5) What are the differences in surface and ground temperature between the mineral and adjacent organic instrumented sites at Yellowknife and Ekati?

5.2 Instrumentation

T_A , T_S , and T_{TOP} were measured at the instrumented sites every two hours using thermistors (Onset Computing, model TMC6-HA), attached to four-channel miniature data loggers (Onset Computing, HOBO^{TM}, model H08-006-04). The error associated with the temperature measurements is largely a function of instrument precision, which is ± 0.38 °C at 20 °C. However, random error associated with instrument precision is reduced in long-term averages because of the large number of measurements (4380 measurements yr^{-1}), especially for large temperature ranges such as those in the air and at the surface. The reported accuracy of the thermistors was ± 0.5 °C, but when calibrated in an ice bath the thermistors were regularly found to be within 0.16 °C, and never more than 0.3°C from 0 °C.

T_A values were measured 1.5 m above the ground by sensors shielded from radiant heating and precipitation. T_A measurements were distributed throughout the Yellowknife study area at instrumented sites Y-1a, Y-2a, Y-3a, and Y-4a (Figure 4.1). At Colomac and Ekati, T_A values were measured at the instrumented sites furthest from each other; at C-1a, and C-3c, and at E-1a, and E-4a (Figure 4.2 and 4.3).

T_S values were measured beneath the base of the live moss approximately 5 cm below the ground surface. The temperature at this depth is used in examinations of permafrost-climate relations (Klene *et al.* 2001; Karunaratne and Burn 2004), principally

because the sensor is largely shaded from radiation. In addition, this method positions the sensor close to the ground surface, rather than at the organic-mineral soil interface, which thereby allows comparison of T_S at sites with different organic-layer thicknesses. The surface temperature sensors were mounted on a wooden dowel and inserted into an augered hole. Several of the surface temperature sensors were exposed to radiative heating in summer 2006 due to subsidence of the ground and/or heaving of the wooden dowel, and had to be eliminated from the analysis. T_S records affected by radiation were identified as having midday T_S values several degrees higher than T_A .

Ground temperatures were measured at 50 cm and 100 cm from the surface at all the instrumented sites. The permafrost table was encountered within 100 cm of the surface at all of the organic instrumented sites, so the 100 cm ground temperature (T_{100}) measurement was used as T_{TOP} . In practice it is impossible to measure T_{TOP} precisely as the exact position of the top of permafrost fluctuates over time. However, the measurement at 100 cm was likely to be a close approximation, because the thermal properties in permafrost do not change greatly each year. At organic instrumented sites, where air temperatures were not measured, ground temperature was also measured at 20 cm. At mineral sites Y-5 and E-5, where permafrost was not encountered within the uppermost 100 cm, ground temperatures were also measured at 150 cm but were not within permafrost. Permafrost was absent at Y-5 and the active layer at E-5 was 2-3 m thick (Hu *et al.* 2003).

5.2.1 Instrument Installation

The instruments were installed in summers 2003, 2004 and 2005. Sites Y-1a, Y-2a, Y-3a, Y-4a, and Y-4b were established in August 2003. Sites Y-5a, E-5a, and E-5b

were established in August 2005. All of the Colomac instrumented sites, and the remaining Yellowknife and Ekati instrumented sites were established in August 2004. Temperature records began shortly after the sites were established, and terminated in May 2007 for the Yellowknife sites, on 19 September 2006 for the Colomac sites, and on 4 September 2006 for the Ekati sites.

5.2.2 Missing Data

Missing values of T_A and T_S for the final weeks of the 2006 thaw season at the Colomac and Ekati instrument sites were estimated. T_A values at the instrumented sites were estimated from the T_A measured by Indian and Northern Affairs (INAC) at Colomac, and by the airport at Ekati. A least-squares regression over July and August 2006 was used to predict T_A at the instrumented sites from the established T_A records. Missing T_S values were estimated using the estimated T_A and the thawing n-factor calculated for summer 2006 up until missing data. This method was evaluated using air and surface temperatures from the Yellowknife sites where the T_A and T_S records were complete. The difference between the estimated and measured mean thawing season air and surface temperatures was ± 0.02 °C and ± 0.03 °C respectively.

5.3 Air Temperature

Permafrost temperature is a function of both air temperature and microclimate. Comparison of permafrost and air temperature variability at different scales allows assessment of the relative control of air temperature, versus microclimate, on permafrost. Air temperatures measured at selected sites across the Slave Province were examined to assess the difference in air temperature within and between the three study areas on a daily, seasonal, and annual basis.

5.3.1 Air temperatures within the study areas

Daily mean air temperatures (T_{Ad}) at Yellowknife, Colomac, and Ekati fluctuated synchronously throughout the study period (Figure 5.1). Variations in T_{Ad} within the study areas were examined by plotting the principal axes, rather than the least-squares linear regression lines, to account for variation in both variables (Mark and Church 1977). T_{Ad} values within the three study areas were similar and highly correlated, with coefficients of determination (r^2) > 0.99 (Figure 5.2; Table 5.1). The principal axes of T_{Ad} between the instrumented sites at the three study areas had slopes close to 1.0, and y-intercepts within 0.53 of the origin. This suggests that the air temperature did not systematically vary by more than 0.5 °C (the accuracy of the thermistors) within each of the three study areas on any given day.

There was also little variation in T_A within the study areas on a seasonal basis (Figure 5.3). Mean seasonal temperatures for this chapter are defined by calendar dates such that the freezing seasons extend from 1 September to 30 April, and the thawing seasons extend from 1 May to 31 August. The maximum range in mean air temperature for the freezing season (T_{Af}) within a study area for either year was less than 0.8 °C (Figure 5.3a). The variations in mean air temperature for the thawing season (T_{At}) within the study areas were also low, with similar maximum ranges at all three study areas (Figure 5.3b). The annual mean air temperatures (T_{Aa}) were calculated from T_{Ad} between 1 September and 31 August. The variations in T_{Aa} within the study areas were also small, with ranges less than 0.6 °C for both Yellowknife and Colomac and less than 0.2 °C for Ekati (Figure 5.3c). The small differences in T_A within the study areas on a daily, seasonal, and annual basis were due to variations in the instrument precision and

Figure 5.1 Daily mean air temperatures at sites Y-1a, C-1a and E-1a, from September 2004 to September 2006. The daily mean air temperature was smoothed with a five-day running mean for clarity.

Figure 5.2 Principal axes of daily mean air temperatures within the (a) Yellowknife, (b) Colomac, and (c) Ekati study areas.

(a) Coefficient of Determination (r^2)

[illegible]

(b) Slope of Principal Axis

[illegible]

(c) Intercept of Principal Axis

[illegible]

Figure 5.3 Mean air temperatures for the (a) freezing season (1 September to 30 April), (b) thawing season (1 May to 31 August), and (c) annual period (1 September to 31 August) at the instrumented sites and the Yellowknife and Ekati airports.

accuracy, microclimate, and/or timing of the measurements. However, the T_A values within the study areas were assumed to be uniform because these differences were minimal. These small differences in T_A within the study areas show that T_A in the boundary layer at peatlands does not vary at the local scale in the Slave Province.

5.3.1.1 Airport air temperature

Climate-permafrost models employ T_A records from standard meteorological stations, such as those established at airports, and apply them to surrounding natural areas (Juliussen and Humlum 2007; Duchesne *et al.* 2008). T_A may be inappropriately applied to areas far away or with very different microclimates. The difference in T_A between standard meteorological stations and natural sites should be determined when examining climate-permafrost relations for a region.

T_A values at the Yellowknife (YZF) and Ekati (YOA) airports were slightly warmer than those measured at the adjacent instrumented sites on a daily, seasonal, and annual basis. At Yellowknife, the r^2 values for T_{Ad} at the airport and the instrumented sites were 0.99 and the principal axes had slopes of 1.0 (Figure 5.4; Table 5.1). The y-intercepts indicate that YZF was systematically 0.4 to 0.9 °C warmer than the instrumented sites. The principal axes of T_{Ad} between the YOA and the Ekati instrumented sites also had slopes of 1.0, and the y-intercept values indicate that YOA was systematically 0.2 to 0.3 °C higher than the instrumented sites. The scatter associated with the relation between YOA and the Ekati instrumented sites could be caused by T_{Ad} at YOA calculated from daily maximum and minimum air temperatures.

Figure 5.4 Principal axes of daily mean air temperatures at (a) Yellowknife Airport (YZF) and site Y-1a, and (b) Ekati Airport (YOA) and site E-1a.

T_A values at the airports were warmer than the instrumented sites in both the freezing and thawing seasons and on an annual basis (Figure 5.3), but never by more than more than 1.5 °C.

Discrepancies in daily, seasonal, and annual T_A between the airport and the instrumented sites could have resulted from differences in microclimate, the precision and accuracy of the instruments, or the frequency and timing of measurements from which the mean was calculated. While the differences between the study sites and airports in this study were minimal it is important to acknowledge this difference because air temperature from standard meteorological stations is often applied to remote regions in climate-permafrost modelling.

5.3.2 Air temperature among the study areas

T_A values differed among the study areas on a daily, seasonal, and annual basis, and were lowest north of treeline at Ekati, and highest south of treeline at Yellowknife (Figure 5.3). The difference in T_A between Colomac and Yellowknife was similar to that between Colomac and Ekati. The y-intercepts of the principal axes show that T_{Ad} at Colomac were between 2.1 and 3.2 °C lower than at Ekati and similarly higher at Yellowknife (Figure 5.5; Table 5.1). The high coefficients of determination ($r^2 > 0.89$) and principal axes slopes of unity for T_{Ad} among the study areas indicate that Yellowknife, Colomac and Ekati all lie within the same climatic system. At Colomac, T_A on a seasonal basis were also higher than at Ekati, and lower than at Yellowknife by about 2 to 3 °C (Figure 5.3). The differences in T_{Aa} among the three study areas were remarkably consistent for both 2004-05 and 2005-06 (Figure 5.3). In both years T_{Aa} at Colomac was 2.7 °C lower than at Yellowknife and 2.7 °C higher than at Ekati.

In 2004-05, the mean T_{Aa} values among the instrumented sites was -5.9°C at Yellowknife, -8.6°C at Colomac, and -11.4°C at Ekati. The following year, 2005-06, air temperatures were several degrees higher across the Slave Province, but annual and seasonal mean air temperatures increased by different amounts (Figure 5.3). T_{Aa} values were more than 4°C higher than the previous year, but in winter 2005-06,

Figure 5.5: Principal axes of daily mean air temperatures between the study areas at (a) Y-1a and C-1a, (b) C-1a and E-1a, and (c) Y-1a and E-1a.

T_{Af} values were approximately 5 °C higher than the previous winter, and T_{Ai} values were 2.0 to 3.0 °C higher than the previous summer. The warm air temperatures in 2005-06 represent a climatic anomaly similar to the 2071-2100 climate warming scenario for the region (Environment Canada 2010).

5.4 Ground Thermal Regime

The annual thermal regime for the surface, active layer, and top of permafrost in the Slave Province differs in many respects from that of the air. There was considerable variation of the ground thermal regime within the study areas, and distinct differences north and south of treeline. Figure 5.6 presents the daily mean surface, active layer and top of permafrost temperatures at sites E-3b, C-2a, and Y-2c for the study period. These three sites capture the range of thermal regimes expressed at the organic instrumented sites.

At the beginning of the freezing season, the surface, active layer, and top of permafrost were isothermal at 0 °C as the temperature gradient reversed, and freezing of soil water began at the surface and base of the active layer. Temperatures remained constant just below 0 °C in the zero-curtain period. In this study, the zero-curtain refers to the phenomenon of constant temperatures at any time of year between 0 and -0.5 °C, and is dependent on depth in the active layer as in Cook (1952), and does not describe freezeback of the entire active-layer. The length of the zero-curtain period varied spatially and interannually depending on moisture conditions and temperature gradients. At the surface, there was no systematic variation in the duration of the zero-curtain period among the three study areas. The surface zero-curtain period lasted only a few days at

instrumented sites where the top of the active layer was dry, and 4 weeks at wetter sites.

The surface and top 20

Figure 5.6 Daily mean temperatures at the surface (T_{sd}), and 20 cm, 50 cm, and 100 cm depth for sites E-3b, C-2a, and Y-2c from September 2004 to September 2006.

cm of the ground experienced zero-curtain conditions while the top of the active layer froze, after which the temperature declined slightly and fluctuated between 0 and -5 °C. At this time, the daily mean surface temperature (T_{sd}) did not drop below -5 °C because the insulating snow cover inhibited the escape of latent heat released deeper in the active layer, and isolated the surface from the cold overlying air. The temperature at 50 cm and 100 cm depths remained in the zero-curtain period for at least a month at all of the instrumented sites across the three study areas.

The active layer cannot cool substantially while the soil water within it freezes, because of the latent heat released. Freezeback of the active layer is characterized in this study by T_{sd} fluctuating between 0 °C and -5 °C and T_{TOP} constant just below 0 °C (Figure 5.6). The zero-curtain period differs from the freezeback period in that temperatures remain constant just below 0 °C during the zero-curtain period and refer to a specific depth, while the freezeback period describes a thermal period for the entire active layer and top of permafrost under two-sided freezing. Once the majority of the active-layer water is frozen, surface, active-layer, and top of permafrost temperatures typically declined noticeably, marking the end of the freezeback period as sensible heat rather than latent heat is released from the ground. Across the Slave Province and regardless of snow depth, daily mean ground surface, active-layer, and top of permafrost temperatures did not drop below -5 °C until T_{TOP} dropped out of the zero-curtain period (<-0.5 °C) in either year. Thus daily mean surface or active-layer temperatures above -5 °C indicated a partially unfrozen active layer. This observation is informative for winter land use regulations in the region (Bryant *et al.* 2008; Lily *et al.* 2008).

The relative duration of freezeback varies interannually, and north and south of treeline. At Ekati, freezeback comprised no more than a third of the freezing season, and increased only slightly in 2005-06, when T_A was higher (Figure 5.7). In both years freezeback periods were longer south of treeline at Yellowknife and Colomac than at Ekati. At several sites south of treeline, the active-layer temperature never dropped below $-5\text{ }^{\circ}\text{C}$, which suggests that the active layer never completely froze during the entire winter (Figure 5.7). In contrast to conditions at Ekati, south of treeline the length of the freezeback periods were considerably longer in 2005-06 than the previous year, when T_A was lower (Figures 5.6 (c) and 5.7). Of the 15 Yellowknife and Colomac sites with permafrost, the freezeback period comprised 100% of the freezing season at only 3 sites in 2004-05 but at 9 sites the following year (Figure 5.7). The freezeback periods ended by the first week of December at all the Ekati organic sites, and no earlier than January at the coldest instrumented sites south of treeline. Even if the active layer did completely freezeback near the end of the freezing season, near-surface ground temperatures did not drop below $-5\text{ }^{\circ}\text{C}$ because T_A was then increasing and did not support heat extraction required for substantial cooling. As a result, long freezeback periods correspond to warmer winter ground temperatures.

In April, the surface, active layer and top of permafrost became isothermal at temperatures equal to or less than $0\text{ }^{\circ}\text{C}$ as the snow melted and the temperature gradient changed direction. T_S rose above $0\text{ }^{\circ}\text{C}$ once the snow had melted and fluctuated throughout the summer in response to T_A . Ground temperatures at 20 cm depth (T_{20}) rose above $0\text{ }^{\circ}\text{C}$ approximately 2-3 weeks after T_S , and were several degrees lower than T_S on a daily basis throughout the summer (Figure 5.6). T_S and T_{20} were equivalent for

Figure 5.7 Active-layer freeze-back duration as a percent of the freezing season.

the last 2-3 weeks of the thawing season as temperature gradient changed direction.

Although relations between T_A and T_S have been explored using temperatures measured at 20 cm depth (Burn 1998; Karunaratne and Burn 2003; Taylor 1995), these measurements should be interpreted with caution because of the differences in T_S measured at various depths. This is especially true in organic material with steep temperature gradients at the surface.

Temperatures at 100 cm depth (T_{100}) did not rise above 0 °C in either year at any of the organic instrumented sites. Where the freezeback period extended over the entire freezing season, T_{TOP} remained at 0 °C throughout the thawing season (Figure 5.6(b)). Where the freezeback period comprised more than 50% of the freezing season, T_{TOP} rose quickly to 0 °C at the same time as T_S reached 0 °C (Figure 5.6(c)). Where the freezeback period comprised less than 50% of the freezing season, T_{TOP} increased slowly to 0 °C by late summer (Figure 5.6(a)).

These observations are pertinent to the active-layer thermal regime in summer. The ground cannot be completely thawed until it is almost 0 °C, since most latent heat is consumed between -1 and 0 °C. Thawing of the active layer will therefore be restricted if the majority of the summer is spent warming the ground (uptake of sensible heat), rather than thawing it (uptake of latent heat). Conversely, if the freezeback period comprised the majority of the freezing season, the majority of the thawing season is comprised of ground thawing, rather than warming, because the ground temperatures are already close to 0 °C. North of treeline where the snow is thin and active-layer freezeback is rapid, the ground thermal regime is dominated by the extraction and uptake of sensible heat and therefore ground temperatures are responsive to climate. South of treeline where a thick

snow cover retards freezeback, the ground thermal regime is dominated by the extraction and uptake of latent heat and therefore ground temperatures are less sensitive to climate. The thermal regime of the active layer in summer is therefore dependent on winter conditions.

Temperatures at 50 cm depth (T_{50}) behaved similar to T_{100} but were lower in winter and rose above 0 °C at several sites (Figure 5.6). At sites where the active layer was less than 50 cm thick, T_{50} and T_{100} were remarkably similar on an annual basis despite differences in minimum and maximum temperatures. Therefore, T_{100} was used to represent T_{TOP} for this study.

5.5 Surface Temperature

There was considerable variation in T_S within each of the study areas, and a distinct difference north and south of treeline. Since T_S is a function of both T_A and microclimate, variation in T_S where T_A was uniform, represents the influence of the microclimate. T_S varied within and among the three study areas and this variation differed throughout the year. The non-parametric Mann-Whitney U statistic was used to determine differences among the study areas (Hammond and McCullagh 1978).

Figure 5.8 displays the daily mean surface temperature (T_{Sd}) at the (a) coldest and (b) warmest organic sites at Yellowknife, Colomac and Ekati. Unlike T_{Ad} , the variability of T_{Sd} changed throughout the year due to distinct changes in the microclimate. In both 2004 and 2005, T_{Sd} was consistently below 0 °C by late September at all the instrumented sites. In winter, T_S behaved differently north and south of treeline in terms of freezeback duration, and minimum T_{Sd} . The duration of the freezeback periods were considerably shorter north of treeline than they were at Yellowknife and

Figure 5.8 Daily mean surface temperatures from September 2004 to September 2006 at Yellowknife, Colomac, and Ekati sites with (a) cold surfaces, and (b) warm surfaces.

Colomac (Figure 5.8b). All of the Ekati organic sites experienced low T_{sd} with minimum values dropping below -15°C in both years.

T_{sd} rapidly increased in April at all the instrumented sites as the snowpack began to ripen. At many sites, T_{sd} remained constant between 0°C and -0.5°C in spring as melt water percolated through the snowpack and refroze at the surface, releasing latent heat. The snowmelt period was defined as beginning on the first day T_{Ad} rose above 0°C , and ending on the first day T_{sd} rose above 0°C . The duration of this snowmelt period was variable as it was dependent on the snow depth, T_A , and the radiation regime. Typically this period was longer at Yellowknife and Colomac (10 to 35 days), than it was north of treeline at Ekati (2 to 19 days).

T_{sd} rose above 0°C at Yellowknife between one and six weeks earlier than at Ekati. Once the snow had melted, T_{sd} values fluctuated above 0°C on a daily basis throughout the summer months, and were similar to T_{Ad} . Figure 5.8 highlights the differences in T_{sd} within the study areas (Figure 5.8 (a) versus (b)), and the variation among the study areas (Figure 5.8 (a) and (b)), and shows that the variation both within and among is concentrated in the winter months. In winter, T_{sd} values south of treeline at Yellowknife and Colomac were considerably warmer than north of treeline at Ekati. In summer, differences in T_{sd} among the three study areas were similar to differences in T_A .

5.5.1 Freezing surface temperatures

In winter, T_s values were similar south of treeline at Yellowknife and Colomac despite differences in T_A . The mean surface temperatures for the freezing season (1 September to 30 April; T_{sf}) were slightly lower at Colomac than at Yellowknife where T_{Af} was over 3°C higher (Figure 5.9a). The warmest surfaces at Colomac and .

Figure 5.9 Mean surface temperatures for the (a) freezing season (1 September to 30 April), (b) thawing season (1 May to 31 August), and (c) annual period (1 September to 31 August) at the organic and mineral instrumented sites.

Yellowknife had similar T_{sf} , and did not reflect the differences in T_{Af} . At these warm sites T_{sd} fluctuated between 0 and -5 °C all winter, suggesting that the active layer released latent heat throughout the winter.

Conversely, there was a greater difference in T_{sf} at the coldest Colomac and Yellowknife sites. At these colder sites, the freezeback period took less than 50% of the freezing season, so T_s responded to cold mid-winter T_A . However, T_{sf} values at Colomac's coldest sites were no more than 1.7 °C lower than at Yellowknife. Overall, T_{sf} values were not significantly different among the organic instrumented sites at Yellowknife and Colomac in 2004-05 ($U=20$, $n= 8 \text{ \& } 7$, $p=0.397$), or 2005-06 ($U=17$, $n= 8 \text{ \& } 7$, $p=0.232$). The difference in T_{sf} between Yellowknife and Colomac corresponds to differences in T_{Af} at cold sites but not at warm sites, which suggests that T_{sf} is resistant to changes in T_{Af} at warm sites where the active layers do not freeze back the entire winter.

Despite a 3 °C difference in T_{Af} between Colomac and Yellowknife, T_{sf} were similar (Figure 5.9a) because snow conditions, which control rates of active-layer freezeback, were comparable at these two sites. A similar difference in T_{Af} existed between Colomac and Ekati, but T_{sf} values were 3.0 °C to 4.0 °C lower north of treeline at Ekati where snow cover was considerably thinner. Thus, thick snow cover inhibits the relation between T_{Af} and T_{sf} , while thin snow cover enables it.

In winter 2005-06, when T_{Af} was close to 5 °C higher across the Slave Province, T_{sf} south of treeline values were 1.7 °C higher on average than the previous winter, but the increases between years varied among the sites. Between these two years, T_{sf} increased by as little as 0.5 °C at wet sites with warm surfaces and long freezeback periods, and increased by as much as 3.1 °C at dry sites with cold surfaces and shorter

freezeback periods. At these warm, wet sites, T_S did not drop below $-5\text{ }^{\circ}\text{C}$ throughout either the warm 2005-06 winter or the cooler 2004-05 winter because the active layer was kept warm by latent heat released throughout the season. At the cold sites, T_S did not drop below $-5\text{ }^{\circ}\text{C}$ until much later in the warm 2005-06 winter, suggesting that the active layer took longer to freeze-back. At both Yellowknife and Colomac, the range of T_{Sf} contracted by more than 50% in the warm 2005-06 year, because of the varying responses of T_{Sf} to changes in T_{Af} (Figure 5.9(a)). At wet sites south of treeline where the freezeback period was prolonged, T_{Sf} were more resistant to changes in T_A than at dry sites.

T_{Sf} at the Ekati sites also responded differently to the increase in T_{Af} in 2005-06. Similar to Yellowknife and Colomac the warmest organic sites at Ekati experienced smaller increases in T_{Sf} from 2004-05 to 2005-06 than the colder sites. This resulted in a lower range of T_{Sf} in 2005-06 as compared with the previous cooler winter; however, the range decreased by only 20% as opposed to 50% south of treeline. The differences in range were less extreme north of treeline, because T_S had the opportunity to respond to T_A following the relatively short active-layer freeze-back duration.

5.5.2 Thawing surface temperatures

Unlike T_{Sf} , the difference in mean thawing season surface temperatures (1 May to 31 August; T_{St}) north and south of treeline was less distinct (Figure 5.9b). The ranges in T_{St} at Yellowknife and Colomac overlapped considerably, and the warmest surfaces at Ekati were similar to the coldest surfaces south of treeline. However there were significant differences in T_{St} among the three study areas. In 2004-05, T_{St} at Colomac was slightly lower than at Yellowknife ($U=34$, $n=6\&7$, $p=0.03$), and significantly higher than

at Ekati ($U=36$, $n=6\&7$, $p=0.002$). Although there were insufficient data at Colomac in 2005-06 to determine statistical differences in T_{St} among the three study areas, the data available suggest the differences were similar to 2004-05. In 2005-06, the median T_{St} at Colomac was 2.0 °C lower than at Yellowknife, 2.0 °C higher than at Ekati, and range of T_{St} for the three areas overlapped each other by no more than 0.5 °C. The differences in T_{At} among the three study areas, and the close agreement between T_{At} and T_{St} within the study areas suggests that climate, rather than microclimate, dominates the surface temperature regime during the thawing season. This conclusion is supported by interannual changes in T_{At} . In 2005-06 when T_{At} values were between 2.8 °C and 4.3 °C higher than the previous year, T_{St} values were between 1.5 °C and 4.7 °C higher.

5.5.3 Annual surface temperatures

Although the annual mean surface temperature (1 September to 31 August; T_{Sa}) is a summary of both T_{Sf} and T_{St} , variations of T_{Sa} across the Slave Province were similar to that of T_{Sf} , with distinct differences north and south of treeline (Figure 5.9c). In 2004-05, T_{Sa} values at the organic sites were not significantly different at Yellowknife and Colomac ($U=9$, $n=6\&7$, $p=0.101$), and were at least 4.4 °C lower north of treeline at Ekati. There were similar differences in T_{Sa} north and south of treeline the following year but insufficient data at Colomac prevented determination of statistical difference. Like both T_{Sf} and T_{St} , T_{Sa} responded to higher T_A in 2005-06. Across the Slave Province, T_{Sa} values were between 0.8 and 2.9 °C higher in 2005-06 when T_{Aa} values were approximately 4.3 °C higher. Similar to T_{Sf} , T_{Sa} increased only slightly at wet sites with long freezeback periods.

The differences in T_{Sa} , both north and south of treeline and between the two years, were similar to those exhibited by T_{Sf} . However, the differences in T_{Sa} were not as extreme as those exhibited by T_{Sf} because of the moderating influence of T_{St} . T_{Sf} was controlled by microclimate, which differs across treeline, while T_{St} was controlled by T_A , which decreased with latitude irrespective of treeline. Since the freezing season was longer than the thawing season, T_{Sa} reflected T_{Sf} but differences north and south of treeline were moderated by T_{St} .

5.6 Temperature at the Top of Permafrost

The permafrost table lay within 100 cm of the surface at all the organic sites, and the temperature at the top of permafrost (T_{TOP}) was measured at this depth. Similar to T_S , T_{TOP} differed throughout the year and across treeline (Figure 5.10). T_{TOPd} typically reached its minimum in March, following freezeback of the active layer, and was considerably lower at Ekati. At all of the Ekati sites, T_{TOPd} dropped below $-11.0\text{ }^{\circ}\text{C}$ in both years. South of treeline at Yellowknife and Colomac, minimum T_{TOPd} values were no less than $-7.0\text{ }^{\circ}\text{C}$, and T_{TOPd} remained constant just below $0\text{ }^{\circ}\text{C}$ throughout winter at several sites. T_{TOPd} began to increase in April and reached its maximum between $-1.0\text{ }^{\circ}\text{C}$ and $0\text{ }^{\circ}\text{C}$ by August. As for T_{Sd} , T_{TOPd} values were similar north and south of treeline by late summer. In 2005-06 when T_A and T_S were higher across the Slave Province, the thermal regime at the top of permafrost remained the same as the previous summer but differed in winter. In the warmer 2005-06 winter, the top of permafrost at all the instrumented sites experienced longer freezeback periods and higher minimum temperatures than the previous winter.

Figure 5.10 Daily mean ground temperatures at 100 cm depth from September 2004 to September 2006 at selected Yellowknife, Colomac, and Ekati instrumented sites with (a) cold ground, and (b) warm ground.

In 2004-05, the annual mean T_{TOP} (T_{TOPa}) values were not significantly different at Yellowknife and Colomac ($U=15$, $n= 7 \text{ \& } 6$, $p=0.445$), and were between 3.0 and 6.6 °C lower at Ekati (Figure 5.11). The following year, T_{TOPa} exhibited similar differences north and south of treeline. The increases in freezeback periods and the minimum ground temperatures from 2004-05 to 2005-06 led to increases in T_{TOPa} , however some sites increased more than others especially south of treeline. Similar to T_{Sa} , T_{TOPa} increased only slightly, less than 0.5 °C, at the warm Yellowknife and Colomac sites, while T_{TOPa} at the cold sites increased by several degrees. This led to a narrower range of T_{TOPa} at Yellowknife and Colomac in the warm 2005-06 year as compared to the previous year. Conversely, T_{TOPa} at Ekati increased between 1.0 to 2.5 °C from 2004-05 to 2005-06, which caused the T_{TOPa} range to remain approximately the same but shift towards warmer temperatures. Similar to T_{Sa} , distinct differences in the duration of the freezeback periods north and south of treeline, led to distinct differences in T_{TOPa} .

5.7 Mineral Sites

Surface and ground temperatures at the Yellowknife and Ekati mineral sites, Y-5 and E-5 respectively, were measured in 2005-06 to contrast the differences in thermal regimes with those at the organic sites. The thermal regimes at these mineral sites differ from those at the adjacent organic sites. However the relation between the mineral and organic sites at Yellowknife differs from that at Ekati because permafrost is absent at the Yellowknife mineral site.

The mineral surfaces at Ekati were much colder in winter and much warmer in summer than the adjacent organic surfaces (Figure 5.12), but since the winters were significantly longer than the summers, T_{Sa} values were similar to or slightly lower than

Figure 5.11 Annual (1 September to 31 August) mean ground temperatures at top of permafrost (100 cm depth) at the organic and mineral instrumented sites.

Figure 5.12 Daily mean surface temperatures from September 2004 to September 2006 at an organic and mineral site at (a) Ekati and (b) Yellowknife.

the cold organic sites. At Ekati, T_{100} values at the mineral sites were also much lower in winter and higher in summer (Figure 5.13(a)) but were the same as at the organic sites on an annual basis (Figure 5.11).

At the Yellowknife mineral site, the T_{Sa} was similar to that at the adjacent cold organic sites (Figure 5.9(a)), because the mineral site was dry and therefore the surface temperature in winter remained cool (Figure 5.12(b)), and because it was not greatly affected by latent heat. Conversely, the annual mean T_{100} at the Yellowknife mineral site was several degrees higher than at all the organic sites (Figure 5.11) because permafrost was absent at this site and ground freezing, rather than cooling, continued throughout the winter.

These results are significant for the basal temperature of snow (BTS) method used to predict permafrost distribution. According to this method, when T_S values measured under ≥ 80 cm snow cover in late winter are: < -3 °C, permafrost is probable; -2 to -3 °C, permafrost is possible; and > -2 °C, permafrost is improbable (Haeberli 1973; Lewkowicz and Ednie 2004). At Yellowknife under similar snow conditions, T_{Sd} in March 2006 ranged from -7.2 to -3.9 °C at the coldest site (Y-1a), and from -3.5 to -1.1 °C at Y-2a where permafrost was present but freezeback of the active layer continued throughout the freezing season. At the Yellowknife mineral site where permafrost was absent and snow covers were slightly thinner, T_{Sd} values in March 2006 were quite low because of the low soil moisture, ranging from -5.9 to -2.0 °C. Although the snow-thickness requirements for BTS were not met, the March T_{Sd} values at these Yellowknife sites show that within the discontinuous zone cold surfaces do not necessarily indicate the

Figure 5.13 Daily mean ground temperatures (150 cm depth) from September 2004 to September 2006 at and organic and mineral site at (a) Ekati and (b) Yellowknife.

presence of permafrost, and that warm surfaces may not indicate its absence. The BTS method is probability-based, and it could be refined to consider moisture conditions in the substrate.

5.8 Summary Points

This chapter examined the variability of air, surface and ground temperatures within and among the study areas in the Slave Geological Province, NWT. The following points summarize the results of geographical interest presented in this chapter:

- (1) Air temperatures were distinctly different at the three study areas and did not vary at the local scale on a daily, seasonal, or annual basis. Air temperatures at the Yellowknife and Ekati airports were higher than at the adjacent instrumented sites in both the freezing and thawing seasons and on an annual basis.
- (2) Within the three study areas, surface temperatures varied by several degrees both seasonally and annually due to differences in site characteristics that control the surface energy balance.
- (3) Surface temperatures during the freezing season and on an annual basis were not significantly different at Yellowknife and Colomac, south of treeline. These values were significantly higher than at Ekati, north of treeline. Surface temperatures during the thawing season were not significantly different among the three study areas.
- (4) Under a normal climate, permafrost can be maintained with annual mean surface temperatures greater than 0 °C south of treeline.

- (5) Annual mean temperatures at the top of permafrost were not significantly different between Yellowknife and Colomac, both sites being south of treeline. However values were significantly lower at Ekati, north of treeline.
- (6) Mineral sites at Yellowknife and Ekati had thermal regimes different from the adjacent organic sites. At Ekati the annual mean surface and ground temperatures at the mineral sites were similar to those at the organic sites, despite much larger annual amplitudes. Conversely, at Yellowknife the annual mean surface temperature at the mineral site was identical to that at the coldest organic site, but permafrost was absent.

In addition, a principal observation can be made regarding the ground thermal regime:

- (7) South of treeline, freezeback of the active layer takes longer than north of treeline due to the thicker snow cover and active layers. Long freezeback periods inhibit ground cooling and result in relatively warm surface and ground temperatures throughout the discontinuous permafrost zone irrespective of climate, and buffer the response of the near-surface ground thermal regime to temporal changes in climate.

Chapter Six: The Surface Offset: Relations between air and surface temperatures in the Slave Geological Province

6.1   Introduction

In Chapter Five, air temperature (T_A), ground surface temperature (T_S) and ground temperature at 100 cm depth (T_{100}) were examined in the Slave Geological Province, south of treeline at Yellowknife and Colomac, and north of treeline at Ekati. Across this area, T_A values were progressively lower with increasing latitude throughout the year, with annual mean air temperatures (T_{Aa}) at Colomac 2 °C lower than at Yellowknife, and 2 °C higher than at Ekati. Despite differences in T_A among the three study areas, annual mean surface (T_{Sa}) and ground (T_{100a}) temperatures at Colomac and Yellowknife were similar, and considerably higher than north of treeline at Ekati.

This chapter examines the relation between T_A and T_S on a seasonal and annual basis across the Slave Province. Surface offsets, the difference between T_A and T_S , and n-factors, the ratio of T_S to T_A , were used to explore this relation (Appendix Tables B.5 to B.11). The following research questions are addressed in this chapter.

- (1) What is the variation in surface offset and n-factors within and among the study areas?
- (2) What was the effect of the unusually high air temperatures in 2005-06 on the surface offsets and n-factors?
- (3) How do snow cover, vegetation and, active-layer water affect the n-factors?

6.2 Relation between air and surface temperature

The relation between T_A and T_S changed throughout the year across the Slave Province (Figure 6.1). Values of T_A and T_S were similar in autumn, and the daily mean surface temperature (T_{Sd}) dropped below 0 °C no more than three days after the daily mean air temperature (T_{Ad}). Once T_A and T_S were below 0 °C, the difference between them rapidly increased – T_A cooled while T_S remained just below 0°C in the zero-curtain period. The surface zero-curtain period continued for several days while water close to the surface froze. Following the surface zero-curtain period, T_{Ad} continued to decrease but T_S did not drop below -5 °C until the active layer was frozen and latent heat was no longer being supplied to the surface (see section 5.4). A rapid decrease in T_{Sd} below -5 °C typically occurred once the active layer was completely frozen. The sites at Yellowknife and Colomac had long active-layer freezeback periods that persisted past December and, in cases, lasted all winter (Figure 6.1 (b)), while the freezeback periods at Ekati were completed by early December in both years of the study (Figure 6.1 (a)).

The surface was rarely as cold as the air in midwinter because the insulating snow cover trapped sensible heat. There was a smaller difference between T_{Ad} and T_{Sd} in midwinter at Ekati than at any of the sites south of treeline because the freezeback period dominated the freezing season south of treeline, restricting ground cooling (Figure 6.1). In the second half of the winter the difference between T_{Ad} and T_{Sd} decreased as the air warmed. T_A and T_S were similar across the Slave Province in April as the temperature gradient reversed. During the snowmelt period, T_{Ad} climbed above 0 °C and T_{Sd} remained just below 0°C until the snowpack ripened and melted. Across the Slave Province, T_A and T_S were similar in the thawing season. However, at many sites

Figure 6.1 Daily mean air and surface temperature for 1 September 2004 to 31 September 2006 at (a) Ekati (E-1a) and (b) Yellowknife (Y-1a). Daily means were calculated from 12 observations. Air temperatures were measured in a radiation shield 1.5 m above the ground, and surface temperatures were measured 5 cm below the radiative surface at the base of the live moss.

especially south of treeline, T_{Ad} continued to exceed T_{Sd} by a few degrees in the early summer as surface warming was restricted by the proximity of the frost table (Karunaratne and Burn 2004). By late summer, T_{Sd} and T_{Ad} were similar because the frost table had descended and no longer restricted surface warming. The difference between T_{Ad} and T_{Sd} was smaller in summer compared to winter, especially south of treeline (Burn 1998).

6.2.1 Thermal orbits

Differences in air-surface temperature relations among the study sites were summarized using plots of monthly mean surface temperature plotted against monthly mean air temperature (Figure 6.2). At most sites, the relation between T_A and T_S exhibited hysteresis during winter. For a given T_A , T_S was higher as the surface cooled in early winter than as the surface warmed in late winter. The shape and orientation of these hysteretic paths, known as thermal orbits (Beltrami 1996), describe the relation between T_{Sd} and T_{Ad} throughout the winter. Wide orbits represent large differences in the T_A and T_S relation throughout the winter, with both a substantial freezeback period followed by a substantial cooling period. Narrow or collapsed orbits that lie parallel to the x-axis represent surfaces where the freezeback period dominated the freezing season, and orbits oriented close to the 1:1 line represent sites where T_A and T_S are the same throughout the freezing season due to insignificant freezeback periods. In summer, plots of monthly mean T_A and T_S exhibit only slight hysteresis and are oriented along the 1:1 line. Though the shape of the thermal orbits is often different in winter and summer, the slope of the principal axes through the seasonal thermal orbits approximates the freezing and thawing n-factors.

Figure 6.2 Plots of monthly mean surface vs. monthly mean air temperature for 2004-05. Plots (a), (c) and (e) are examples of sites with low mean annual surface temperatures (i.e. cold), while plots (b), (d) and (f) are sites with high mean annual surface temperatures (i.e. warm). September 2004 (SEP), February 2005 (FEB), and July 2005 (JUL) are labelled on each plot.

Figure 6.2 presents 2004-05 thermal orbits for the coldest and warmest organic sites at Yellowknife, Colomac and Ekati to describe the range of T_A and T_S relations across the study region. The freezing thermal orbits differ in their shape and orientation north and south of treeline. South of treeline at Yellowknife and Colomac, the freezing thermal orbits were narrow and oriented along the x-axis especially at warm sites (Figure 6.2 (b) and (d)). At all the Ekati organic sites, the freezing thermal orbits were wider and had greater hysteresis than the sites south of treeline Figure 6.2 (e) and (f)). There is more variation in the shape of the thermal orbits north and south of treeline than there is within each of the study areas or between Yellowknife and Colomac, which both lie south of treeline. This supports the observation that the air-surface thermal regime during the freezing season differs north and south of treeline. Conversely, the thawing thermal orbits were oriented along the 1:1 line at all the sites, which suggests similarity in air-surface thermal regimes during the thawing season across treeline (Figure 6.1).

The thermal orbits at the Yellowknife and Ekati mineral sites (Y-5a, E-5a and E-5b) also suggest different surface thermal regimes north and south of treeline (Figure 6.3). All three mineral sites had collapsed thermal orbits, but they were oriented along the x-axis at Yellowknife and along the 1:1 line at Ekati. The relation between T_A and T_S at the Yellowknife mineral site was similar to that at the adjacent organic sites. The relation between T_A and T_S at the Ekati mineral sites were considerably different from that at the organic sites.

Thermal orbits are employed in climate-permafrost models to describe the air-surface thermal regime throughout the freezing and thawing seasons (Sushama *et al.* 2007), and use the same rationale as mean monthly n-factors (Buteau *et al.* 2004, 2010).

Figure 6.3 Plots of monthly mean surface vs. monthly mean air temperature for 2005-06 at sites with exposed mineral soil at the surface at (a) Yellowknife Y-5a, and (b) Ekati E-5a. September 2005 (SEP), February 2006 (FEB), and July 2006 (JUL) are labelled on each plot.

Both recognize that the relation between T_A and T_S changes not only between the freezing and thawing seasons but also within them. Like n-factors, the shape and orientation of a thermal orbit is dependent on the microclimate and could therefore be characterized for different surfaces. However this would require that conditions be monitored throughout the year. Variation in the timing of the seasons must also be considered if monthly mean n-factors are used in modeling, because the physical system does not consistently align with the calendar.

6.3 Surface Offsets

The annual surface offset ($S.O._a$) is the difference between T_{Aa} and T_{Sa} (i.e. $T_{Sa} - T_{Aa}$). Although the climate-permafrost relation is expressed on an annual basis (Brown 1963; Lachenbruch *et al.* 1988; Smith and Riseborough 1996), the relation must be examined on a seasonal basis because of the extreme differences in the relation between summer and winter. The seasonal relations between T_A and T_S are regularly summarized by the freezing and thawing n-factors (n_f and n_t respectively). Freezing and thawing surface offsets ($S.O._f$ and $S.O._t$ respectively) have not been widely reported in the literature but are important to consider as they relate directly to $S.O._a$.

For the freezing season, n_f is the ratio of freezing degree-days for the surface (DDF_S) to those for the air (DDF_A) (i.e. DDF_S / DDF_A), while $S.O._f$ is the difference between mean surface temperature for the freezing season (T_{Sf}) and the mean freezing air temperature (T_{Af}) (i.e. $T_{Sf} - T_{Af}$). Since T_{Sf} is always less negative than T_{Af} , n_f is always between 0 and 1.0, and $S.O._f$ is always positive. Close relations between T_A and T_S during the freezing season are represented by high n_f and small $S.O._f$ values. For the thawing season, n_t is the ratio of thawing degree-days for the surface (DDT_S) to those for the air

(DDT_A), while S.O._t is the difference between mean surface temperature for the thawing season (T_{St}) and the mean thawing air temperature (T_{At}). During the thawing season, vegetated surfaces are slightly cooler than the air on a daily basis. Thus n_t values are typically between 0.5 and 1.0, and S.O._t values are regularly negative. Positive S.O._t and n_t values greater than 1.0 are possible for surfaces that are warmer than the air on a seasonal basis (Karunaratne and Burn 2004; Cannone and Guglielmin 2009). On an annual basis, S.O._a values are positive with high values representing large differences between air and surface temperatures. The n-factor, rather than surface offset, should be used on a seasonal basis when examining interannual change and variation among the study areas because it is a ratio and therefore standardized.

The definition of the freezing and thawing seasons – when they begin and end – affects seasonal temperature averages and degree-days used to calculate the surface offsets and n-factors. The seasons can be defined using calendar dates. However this method does not accommodate interannual or spatial variation in the timing or duration of freezing/thawing. A physically based approach to season definition uses T_A or T_S, and the difference between these two approaches depends on the duration of the snowmelt period. T_S, rather than T_A, is usually used to define the seasons (Taylor 1995; Karunaratne and Burn 2004; Juliussen and Humlum 2007), because T_{Sd} fluctuates above and below 0 °C only moderately at the onset of the freezing and thawing seasons, and is more appropriate for surfaces with late-lying snow packs. For this research, the freezing season began when T_{Sd} values were consistently below 0 °C in autumn, and the thawing season began when T_{Sd} first rose above 0 °C in spring. These definitions imply that fluctuations of T_{Ad} and T_{Sd} above and below 0 °C in either spring or autumn were

included in the thawing season and that the annual period began with the onset of the freezing season and ended with close of the thawing season.

The duration and timing of the seasons varied between instrumented sites since each site had a unique surface thermal regime and therefore a unique season definition. Within the study areas, freezing seasons typically started within 1-2 days of each other, but the start of the thawing seasons varied by up to 30 days. The timing of the thawing season was dependent on snowmelt, as T_{sd} could not rise above 0 °C until the snow had melted (Taras *et al.* 2002). The start of the thawing season was more variable south of treeline at Yellowknife and Colomac due to the thicker snow covers and shading from trees.

Annual and seasonal surface offsets were examined at eight organic instrumented sites at Yellowknife, Colomac and Ekati in 2004-05 and 2005-06. Unfortunately logger failure or radiant heating of the surface temperature sensor resulted in the loss of data from several sites. Therefore, pair-wise analysis could not be used to determine interannual differences because data from the sites were not always available for both years. The non-parametric Mann-Whitney U statistic was used to determine differences in surface offsets among the study areas and interannually (Hammond and McCullagh 1978). Surface offsets at the Yellowknife and Ekati mineral sites (Y-5a, E-5a, and E-5b) in 2005-06 were considered separately and not included in comparison testing.

6.4 The Freezing Surface Offset

Across the Slave Province, $S.O_f$ ranged from 8.5 to 16.8 °C and n_f ranged from 0.09 to 0.57 among all the organic instrumented sites over the two-year study period

(Figure 6.4; Table 6.1). In both years there was a distinct difference in $S.O._f$ and n_f across treeline (Table 6.2). Both $S.O._f$ and n_f values were similar south of treeline at Yellowknife and Colomac, while Ekati had significantly lower $S.O._f$ values and higher n_f values. In 2004-05 when the climate was close to normal, $S.O._f$ values ranged from 10.4 to 16.8 °C south of treeline, and were lower, between 8.5 and 10.9 °C, north of treeline at Ekati. Similarly, n_f was lower south of treeline, between 0.09 and 0.34, and higher north of treeline, between 0.47 and 0.57.

Published values of n_f tend to be higher for tundra than for the boreal forest. A ten year study of n_f at eight sites in a 200-km transect along the North Slope of Alaska, found average n_f values for tundra and shrub landscapes between 0.32 and 0.58 (Klene *et al.* 2008). Along a 600-km transect in the boreal forest of the Mackenzie Valley, NWT, n_f measured at 27 sites ranged from 0.13 to 0.48 (Taylor 1995).

The present research represents the first systematic survey showing the difference in n_f north and south of treeline for comparable terrain repeated over two years. The similarity in the $S.O._f$ and n_f values south of treeline at Yellowknife and Colomac, suggests that the surface offset is independent of air temperature. The 2 °C difference in T_{Af} between Yellowknife and Colomac was comparable to that between Colomac and Ekati north of treeline, yet the surface offset values were much different at Ekati than at the sites south of treeline. These results support the hypothesis proposed by Smith and Riseborough (2002) that the northern limit of discontinuous permafrost, which closely aligns with treeline, is accompanied by a change in the surface offset.

In 2005-06 when T_{Af} was 5.0 °C higher across the Slave Province, $S.O._f$ values decreased as T_{Sf} responded to the change in T_{Af} . $S.O._f$ values were statistically significant

Figure 6.4 Surface defined values of freezing (a) surface offsets and (b) n-factors at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are represented by (+). Data from the instrumented sites were not necessarily available for both years.

Table 6.1 Median and range of the surface-defined freezing surface offsets ($S.O._f$) and n-factors (n_f) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Data from the instrumented sites were not necessarily available for both years.

2004-05				2005-06		
$S.O._f$	n	Median	Range	n	Median	Range
Ekati	6	10.1	8.5 to 10.9	9	7	5.7 to 8.5
Colomac	7	14.2	12.1 to 16.8	7	11.3	9.9 to 11.9
Yellowknife	8	12.5	10.4 to 16.0	9	9.5	9.0 to 10.1
n_f						
Ekati	6	0.51	0.47 to 0.57	9	0.55	0.48 to 0.61
Colomac	7	0.20	0.10 to 0.34	7	0.15	0.08 to 0.26
Yellowknife	8	0.23	0.09 to 0.31	9	0.14	0.10 to 0.19

Table 6.2 Mann-Whitney U test comparing freezing surface offsets (S.O._f) and n-factors (n_f) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Significant differences with p values ≤ 0.05 are in bold.

2004-05			2005-06		
S.O. _f	Colomac (n=7)	Yellowknife (n=8)	S.O. _f	Colomac (n=7)	Yellowknife (n=8)
Ekati (n=6)	U=0; p=0.001	U=2; p=0.003	Ekati (n=7)	U=0; p=0.001	U=0; p<0.001
Yellowknife (n=8)	U=14; p=0.121	X	Yellowknife (n=8)	U=3; p=0.004	X
n _f	Colomac (n=7)	Yellowknife (n=8)	n _f	Colomac (n=7)	Yellowknife (n=8)
Ekati (n=6)	U=0; p=0.001	U=0; p=0.001	Ekati (n=7)	U=0; p=0.001	U=0; p<0.001
Yellowknife (n=8)	U=25; p=0.779	X	Yellowknife (n=8)	U=26; p=0.867	X

different in 2005-06 than in the previous year at Yellowknife ($U=0$, $n=8$ & 8 , $p<0.001$), Colomac ($U=0$, $n=7$ & 7 , $p<0.001$), and Ekati ($U=1$, $n=6$ & 7 , $p<0.002$). However, south of treeline $S.O_f$ changed more than 6°C at some sites, and changed as little as 0.5°C at other sites. This resulted in a considerable decrease in the $S.O_f$ range at Yellowknife and Colomac for 2005-06 (Figure 6.4 (a); Table 6.1). Conversely, the range of $S.O_f$ at Ekati remained constant between the two years with $S.O_f$ values decreasing between 2.8 and 3.8°C . Unlike $S.O_f$, there was no significant difference in n_f between the two years at Yellowknife ($p=0.195$), Colomac ($p=0.259$) or at Ekati ($p=0.366$), because n_f is a ratio and is more stable. In 2005-06, warm sites had low n_f , while cold sites experienced a decrease in n_f of up to 0.15 from 2004-05 to 2005-06.

South of treeline, the clear differences in the $S.O_f$ distribution between the two years resulted from varying responses of T_{sf} to higher T_{Af} . $S.O_f$ changed considerably at warm Yellowknife and Colomac sites where T_{sf} were higher than -2°C for both years. T_{sf} at these warm sites remained between 0 and -5.0°C – in the freezeback period – for most of both winters, therefore the higher T_{Af} in 2005-06 substantially reduced the difference between T_A and T_S and resulted in a lower $S.O_f$. Alternatively, T_{sf} at sites where the freezeback period ended by mid-winter responded to T_{Af} , which resulted in only a slightly lower $S.O_f$ in the second warmer winter.

The warm year illustrates the effect of latent heat on the surface offset. At dry sites with short freezeback periods, T_{sf} can respond to T_{Af} because sensible heat dominates the system. At wet sites with long freezeback periods, T_{sf} is held between 0 and -5.0°C by latent heat released at depth throughout the winter, and therefore does not respond to T_{Af} . Thus $S.O_f$ is resistant to changes in T_{Af} where sensible heat dominates

(dry active layers), and sensitive to changes in T_{Af} where latent heat dominates (wet active layers).

Relatively small surface offsets are associated with low T_{100} , but this is not always the case. At the Yellowknife mineral site (Y-5a), where permafrost was absent, the 2005-06 $S.O._f$ was smaller (8.6 °C) and n_f was higher (0.23) than at any of the adjacent organic sites (Figure 6.4). At the Ekati mineral sites, E-5a and E-5b, where the active-layer is over 150 cm deep, $S.O._f$ were several degrees lower, 1.2 and 3.4 °C respectively, and n_f were higher, 0.92 and 0.79 respectively, than the organic sites where the permafrost tables lay within 1 m of the surface.

6.5 Snow Cover and the Surface Offsets

Snow cover is an important component of the surface offset system because it traps heat from leaving the ground, and regional and local variations in T_{Sf} are often attributed to differences in snow cover (Desrochers and Granberg 1988; Juliussen and Humlum 2007; Karunaratne and Burn 2003; Smith 1975; Taras *et al.* 2002). Figure 6.5 presents n_f as a function of April snow depth measured at the instrumented sites for the (a) 2004-05 and (b) 2005-06 freezing seasons. In both years, n_f was higher at Ekati, north of treeline, where snow cover was thinner than at Yellowknife and Colomac, south of treeline. The strength of the relation between n_f and snow depth was evaluated using the Kendall Tau (τ) rank correlation coefficient (Table 6.3). The Kendall Tau compares the number of concordant and discordant pairs and is an alternative to the Pearson Correlation (r) procedure. Although τ tends to be slightly lower than the Pearson r or the Spearman rho, it was chosen because it is more robust with small sample sizes and can cope with the nonparametric distribution of snow (Wilkie 1980).

Figure 6.5 Relation between freezing n-factor (n_f) and April snow depth (cm) at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.

Table 6.3 Kendall Tau (τ) correlation between n_f and April snow depth for freezing seasons (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values ≤ 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.

(a) 2004-05	n	τ	p
Ekati	6	-0.74	0.050
Colomac	7	-0.52	0.099
Yellowknife	8	-0.37	0.209
Slave Province	21	-0.59	0.001
(b) 2005-06	n	τ	P
Ekati	9	-0.82	0.002
Colomac	7	-0.65	0.046
Yellowknife	9	-0.09	0.747
Slave Province	25	-0.66	0.001

There was a significant negative correlation between n_f and snow depth across all the instrumented sites in the Slave Province for both 2004-05 and 2005-06 (Table 6.3). Snow isolates the surface from the cold overlying air and inhibits heat loss from the ground, such that relatively cold surfaces with high n_f have thin snow covers. The correlations across the Slave Province demonstrate that snow controls n_f on a regional scale. However, the correlations were discontinuous north and south of treeline due to the distinct differences in n_f . North of treeline at Ekati, the correlation between n_f and snow depth was strong and significant both years. In 2005-06, the relation between n_f and snow depth at Ekati and across the Slave Province was influenced by mineral sites E-5a and E-5b at Ekati. These sites are outliers and have both high n_f and thin snow covers. Although the exclusion of E-5a and E-5b from the analysis weakens the strength of the correlations between n_f and snow depth, they remain significant at Ekati ($\tau = -0.68$; $p = 0.03$), and across the Slave Province ($\tau = -0.60$; $p < 0.0001$), which shows that these extreme sites strengthen rather than distort the relation.

South of treeline at Yellowknife, the correlations between n_f and snow depth were weak and non-significant for both years. Although the relation between n_f and snow depth was stronger at Colomac, the correlation was non-significant in 2004-05 and marginally non-significant the following year. Since the insulating effect of snow decreases exponentially at depths greater than 30 cm (Smith 1975), the poor relation between n_f and snow south of treeline could be due to the thick snow covers. However this would result in similar n_f south of treeline, and the range is actually larger south of treeline than on the tundra. Therefore n_f is not primarily controlled by snow in the boreal forest.

Across treeline, the relations between n_f and snow depth are scale-dependent. While snow is a critical component of the surface thermal regime both north and south of treeline, its disparate form results in a distinctly different function within the system. The thin tundra snow covers control the spatial variability in n_f in the continuous permafrost zone, but the thick snow covers characteristic of the boreal forest do not explain the variability in n_f , south of treeline.

6.6 The Thawing Surface Offset

As described in section 6.2, the relation between T_A and T_S differed between summer and winter. Across the Slave Province, values of $S.O._t$ were similar, and variation occurred within each study area rather than between them (Figure 6.6; Table 6.4). In 2004-05, the $S.O._t$ ranged from -0.8 to -4.6 °C across the Slave Province, and the median $S.O._t$ were similar at the three study areas; -2.0 °C at Yellowknife, -2.3 °C at Colomac, and -2.7 °C at Ekati. Values of n_t were also similar among the study areas. In 2004-05, n_t ranged from 0.46 to 0.91 across the Slave Province, and the median n_t was 0.75 at Yellowknife and Colomac and slightly lower, 0.67, at Ekati. There was no significant difference among the study areas in either $S.O._t$ ($H=0.626$; $p=0.731$) or n_t ($H=2.137$; $p=0.344$) in summer 2005. Although differences in $S.O._t$ and n_t among the study areas could not be determined in summer 2006 due to insufficient sample size, the data available suggest similarities among the study areas in this year as well. Unlike in the freezing season, $S.O._t$ and n_t did not respond to elevated $T_{A,t}$ in the second year. There was no significant difference between 2005 and 2006 in either $S.O._t$ ($U=141$; $n=19$ & 15 ; $p=0.974$) or n_t ($U=109$; $n=19$ & 15 ; $p=0.255$) across the Slave Province.

Figure 6.6 Surface defined values of thawing (a) surface offsets and (b) n-factors at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are represented by (+). Data from the instrumented sites were not necessarily available for both years.

Table 6.4 Median and range of the surface-defined thawing surface offsets (S.O._t) and n-factors (n_t) at Yellowknife, Colomac and Ekati organic sites for 2005 and 2006. Data from the instrumented sites were not necessarily available for both years.

2004-05				2005-06		
S.O. _t (°C)	n	Median	Range	n	Median	Range
Ekati	6	-2.7	-1.2 to -4.5	4	-2.5	-0.8 to -4.0
Colomac	6	-2.3	-0.9 to -3.8	2	-2.9	-2.4 to -3.3
Yellowknife	7	-2.0	-0.8 to -4.6	6	-3.1	-1.1 to -4.4
n _t						
Ekati	6	0.67	0.46 to 0.85	4	0.74	0.59 to 0.91
Colomac	6	0.75	0.62 to 0.88	2	0.73	0.69 to 0.77
Yellowknife	7	0.75	0.58 to 0.91	6	0.74	0.63 to 0.91

In 2006, $S.O._t$ and n_t at the Yellowknife mineral site (Y-5a) were close to the median values for the organic sites (Figure 6.6). Alternatively, the mineral sites at Ekati (E-5a and E-5b) had $S.O._t$ values that were at least 2.0 °C and n_t values that were above 1.0. Unlike Y-5a at Yellowknife, the surface temperatures at the Ekati mineral sites were not moderated by evaporation because they were unshaded, dry and did not have any ground cover. Dry, unvegetated surfaces with n_t values greater than 1.0 have been reported previously (Kade *et al.* 2006; Juliussen and Humlum 2007; Cannone and Guglielmin 2009).

6.7 Vegetation and the Surface Offsets

Vegetation is regularly cited as a significant controlling factor of the surface thermal regime (Brown 1963; Taylor 1995; Klene *et al.* 2001, 2008; Karunaratne and Burn 2003, 2004; Kade *et al.* 2006; Cannone and Guglielmin 2009). Indeed, cold region energy balances are highly influenced by vegetation in terms of incoming shortwave radiation (Yang and Friedl 2003), snow redistribution (Sturm *et al.* 2001) and interception (Hedstrom and Pomeroy 1998), and moisture regimes (Liston *et al.* 2002).

The organic sites in this study had similar ground cover because they were all in peatlands. However, the Yellowknife and Colomac sites were in treed peatlands and had varying degrees of shading, and the Ekati sites were on the tundra with no shading from trees or tall shrubs. The Yellowknife and Colomac sites were divided into two groups – “shaded” and “open”. Sites in the ‘shaded’ group, were predominantly shaded by trees throughout the day, while sites in the ‘open’ group were in open areas and exposed to the sky. All of the Ekati sites were categorized as “tundra” and were not subdivided.

Vegetation is thought to be particularly important in determining the thawing surface offset as it controls incoming shortwave radiation through shading, therefore differences in n_t between the shaded, open and tundra sites were examined (Figure 6.7; Table 6.5). Values of n_t were significantly different at the open and shaded sites with n_t being higher in 2004-05 at the open sites. However, the importance of shading was not evident when n_t at the open and shaded sites were compared with n_t at the Ekati tundra sites. There was no significant difference in n_t between the shaded and tundra sites despite extreme differences in shading. This suggests that vegetation alone does not directly control n_t . Karunaratne and Burn (2004) suggested that n_t was primarily dictated by the rate of thawing in early summer. Shaded and tundra sites may have similar n_t because their thawing fronts descended gradually compared to the open sites because of reduced incoming shortwave radiation caused by shading south of treeline and latitude north of treeline. Values for n_t could also be lower north of treeline because the surfaces were colder and therefore required considerable sensible heat to raise the active-layer temperature to 0 °C before thawing could begin.

Vegetation was also associated with n_f south of treeline. There was a significant difference in n_f between the Open and Shaded sites at Yellowknife and Colomac in both years (Figure 6.7), with n_f lower at the Open sites. In 2004-05, the Shaded sites had significantly thinner snow covers ($U=9$, $n=10$ & 5 , $p=0.022$) and active layers ($U=9$, $n=10$ & 5 , $p=0.022$) than the open sites. Differences in snow and active-layer depth between the open and shaded sites were not detected in 2005-06. Although neither snow depth nor active-layer thickness were directly correlated with n_f south of treeline they are important components of the surface energy balance and ground thermal regime and

Figure 6.7 Comparison of (a) freezing n-factors (n_f), and (b) thawing n-factors (n_t) between open and shaded sites at Yellowknife and Colomac and the Ekati tundra sites. Data from the instrumented sites were not necessarily available for both years.

Table 6.5 Mann-Whitney U test comparing n-factors between open canopy sites, shaded sites and Ekati Tundra sites. Significant differences with p values ≤ 0.05 are in bold.

	2004-05		2005-06	
(a) Freezing n-factor	Shaded	Tundra	Shaded	Tundra
Open	U=49; p=0.011	U=54; p=0.002	U=58; p=0.006	U=81; p=0.000
Shaded	X	U=36; p=0.005	X	U=63; p=0.001
(b) Thawing n-factor	Shaded	Tundra	Shaded	Tundra
Open	U=5; p=0.034	U=8; p=0.045	N/A	N/A
Shaded	X	U=15; p=0.928	X	N/A

covary with each other. Shaded sites tended to have thinner snow covers and active layers, and higher n_f .

6.8 Soil Water and the Surface Offsets

Moisture availability influences the summer surface energy balance, and is thought to affect T_{Sf} through evaporative cooling, and its influence on thermal admittance. Soil moisture has been included in the metadata for some published n_t studies (Jorgenson and Kreig 1988; Klene *et al.* 2001), and standing surface water on tundra was associated with n_t close to 1.0 due to its low albedo at high sun angles. Moisture varies both spatially and temporally, and therefore is challenging to incorporate into n-factor characterization. However, local spatial variation in moisture is relative because it is controlled by microtopography, ground cover and vegetation, slope, and permeability of the soil, which are generally invariable.

Volumetric soil moisture was measured throughout the active layer at regular intervals within 5 m of each instrumented site in late August 2006 (Chapter Four). From these measurements, average volumetric water for the active layer ($\text{cm}^3 \text{cm}^{-3}$), total active-layer water (cm), and average weighted volumetric water for the active layer ($\text{cm}^3 \text{cm}^{-3}$) were calculated. Total active-layer water is the sum of the products between volumetric soil moisture ($\text{cm}^3 \text{cm}^{-3}$) and the depth of the interval (cm) for the soil samples. Average weighted volumetric water for the active layer is the total water divided by the active-layer depth.

In this research, relations were not found, either within or across the study areas, between n_t at the organic sites and (1) average volumetric water content for the active layer, (2) total active-layer water, (3) average weighted volumetric water for the active

layer, (4) volumetric water content in top 10 cm, or (5) suprapermafrost water table. The absence of a relation between n_t and soil moisture may be because the surfaces were relatively dry and covered in mosses and lichens. Only one site, Y-2b, had visible water at the surface and the suprapermafrost water table was at least 15 cm from the surface at all other sites. Thus evaporative cooling and thermal admittance may have been fairly constant as water was drawn up to the surface by capillary action of the mosses (Brown 1963). Soil moisture did not control n_t at the peatland sites in the Slave Province.

Total active-layer water was correlated with n_f (Figure 6.8; Table 6.6). Like the relation between n_f and snow depth, the relation between n_f and total active-layer water was scale dependent. In both years there was a negative correlation between n_f and total active-layer water across the Slave Province (Table 6.6), but at the local scale of the study areas the relation between n_f and total active-layer water was opposite to that between n_f and snow depth. The correlation between n_f and total active-layer water was not significant either year north of treeline at Ekati, even with the exclusion of clear outliers, sites E-5a and E-5b ($\tau=0.29$, $p=0.362$). South of treeline at Yellowknife and Colomac, there was a significant negative correlation between n_f and total active-layer water in 2004-05, which became marginally significant the following year when the range of n_f collapsed (Table 6.6).

The freezing season is divided into the freezeback period and the cooling period. The ground and surface cannot substantially cool during the freezeback period, because of the latent heat released at depth. The freezeback period is followed immediately by the cooling period, when sensible heat is released and the surface and ground rapidly cools. The ratio of cooling period duration to freezeback period duration

Figure 6.8 Relation between freezing n-factor (n_f) and active-layer water at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.

Table 6.6 Kendall Tau (τ) correlation between n_f and August 2006 volumetric soil moisture for freezing seasons (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values < 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.

(a) 2004-05	n	τ	p
Ekati	6	-0.21	0.559
Colomac	7	-0.62	0.051
Yellowknife	8	-0.76	0.009
Slave Province	21	-0.63	0.000
(b) 2005-06	n	τ	p
Ekati	9.00	0.34	0.206
Colomac	7.00	-0.59	0.068
Yellowknife	9.00	-0.53	0.054
Slave Province	25.00	-0.46	0.001

(cooling/freezeback) is 1.0 when freezeback period comprises half the freezing season, is 0 when freezeback period comprises the entire freezing season, and exceeds 1.0 when freezeback period is short – less than half the freezing season.

In 2004-05, n_f was directly correlated with cooling/freezeback across the Slave Province ($\tau=0.85$, $p<0.000$), south of treeline at Yellowknife and Colomac ($\tau=0.73$, $p<0.000$), and north of treeline at Ekati ($\tau=0.78$, $p=0.032$). South of treeline, cooling/freezeback was correlated with total active-layer water ($\tau=-0.65$, $p=0.001$) but not with snow depth ($\tau=0.03$, $p=0.882$), and was better correlated with snow ($\tau=0.69$, $p=0.056$) than with total active-layer water ($\tau=0.20$, $p=0.573$) north of treeline. These relations were also present in 2005-06, but they were not significant because cooling/freezeback equalled 0 at more than 55% of the sites south of treeline. This indicates that n_f is dependent on duration of the freezeback period, and that the freezeback period is controlled by snow north of treeline and by total active-layer water south of treeline.

Even though there was variation in snow in the boreal forest, the snow covers at all the sites were greater than 40 cm. Since the insulative property of snow decreases exponentially at depths greater than 30 cm (Smith 1975), the thick snow covers south of treeline provided an insulating barrier that reduced the temperature gradient between the surface and the overlying air. This left variation in freezeback period dependent on total active-layer water rather than snow depth in the boreal forest. North of treeline, freezeback period were short regardless of total active-layer water because the thin snow covers resulted in steep temperature gradients between the surface and the overlying air. On the tundra, the majority of the freezing season was dominated by the cooling period,

resulting in low T_{sf} . The importance of subsurface conditions on n_f was suggested by Karunaratne and Burn (2004). This research supports that suggestion and shows that the relations between snow depth, active-layer moisture, and surface offsets are different north and south of treeline.

6.9 The Annual Surface Offset

The $S.O._a$ assimilates the relation between T_A and T_S for the freezing and thawing seasons. In 2004-05, $S.O._a$ ranged from 5.0 to 9.7 °C across the Slave Province (Figure 6.9; Table 6.7). The following year when T_{Aa} were 5 °C higher, $S.O._a$ were between 1.5 and 3.9 °C smaller, ranging from 2.6 to 6.3 °C at the organic sites. The reduction in $S.O._a$ during the warm 2005-06 year was because the T_{Sa} did not increase to the same extent as T_{Aa} did. Similar to $S.O._f$, the range of $S.O._a$ at Yellowknife shrank from 3.4 °C in 2004-05 to 1.4 °C the next year. The range changed dramatically because T_{sf} at the warmest sites stayed the same and did not respond to the change in T_{Aa} as it did at the colder sites.

$S.O._a$ varied within and between the study areas. In 2004-05, the $S.O._a$ at Colomac was significantly higher than at Yellowknife ($U=34$; $n=6$ & 7 ; $p_1=0.037$) and Ekati ($U=36$; $n=6$ & 6 ; $p_1=0.001$). Surprisingly, the $S.O._a$ at Yellowknife and Ekati were not significantly different ($U=10$; $n=6$ & 7 ; $p=0.138$). Although there were insufficient data the following year to test statistically the difference in $S.O._a$ between the study areas, the data available suggest a similar relation. In 2005-06, the highest $S.O._a$ values measured were at Colomac, the median values were remarkably similar at Yellowknife and Ekati. The similar $S.O._a$ at Yellowknife and Ekati, and the high $S.O._a$ at Colomac

Figure 6.9 Surface defined values of the annual surface offsets at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Values for mineral sites Y-5a and E-5a and E-5b are represented by (+). Data from the instrumented sites were not necessarily available for both years.

Table 6.7 Median and range of the surface-defined annual surface offsets (S.O._a) at Yellowknife, Colomac and Ekati organic sites for 2004-05 and 2005-06. Data from the instrumented sites were not necessarily available for both years.

S.O. _a (°C)	2004-05			2005-06		
	n	Median	Range	n	Median	Range
Ekati	6	6.2	5.0 to 6.8	4	3.8	2.6 to 5.0
Colomac	6	8.2	7.3 to 9.7	2	6.1	5.9 to 6.3
Yellowknife	7	7.0	5.1 to 8.5	6	3.9	3.5 to 4.9

result from annual air and surface temperature cooling at different rates with increasing latitude. Snow cover and total active-layer water are also related to $S.O._a$ and exhibit the same scale dependency north and south of treeline as n_f (Figure 6.10; Figure 6.11; Table 6.8). $S.O._a$ was correlated with both snow cover and active-layer water across the Slave Province, but snow cover explained the variation in $S.O._a$ north of treeline, while active-layer water explained the variation south of treeline.

6.10 Summary Points

This chapter examined the spatial variability of the surface offset within and between the study areas in the Slave Geological Province, NWT. The following points summarize the results presented in this chapter.

- (1) Values of the freezing surface offset were similar south of treeline at Yellowknife and Colomac despite differences in air temperature. This suggests that the surface offset is independent of air temperature, and supports the hypothesis that the northern limit of discontinuous permafrost is accompanied by a change in surface offset.
- (2) Snow cover controlled the spatial variability in the freezing surface offset in the continuous permafrost zone of the tundra where snow was thin, but not in the boreal forest where snow covers were thicker.
- (3) There was no significant difference in thawing surface offsets among the study areas.

Figure 6.10 Relation between annual surface offset ($S.O._a$) and April snow depth (cm) at the Yellowknife, Colomac and Ekati study sites for (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years.

Figure 6.11 Relation between annual surface offset ($S.O._a$) and active-layer soil moisture at the Yellowknife, Colomac and Ekati study sites for freezing seasons (a) 2004-05 and (b) 2005-06. Data from the instrumented sites were not necessarily available for both years. There was a significant correlation between $S.O._a$ and active-layer water in 2004-05 ($\tau=0.47$; $p=0.005$), but not in 2005-06.

Table 6. 8 Kendall Tau (τ) correlation between annual surface offset (S.O._a) and April snow depth for (a) 2004-05 and (b) 2005-06. Correlations were computed for the Yellowknife, Colomac, and Ekati study regions, and for all the study sites across the Slave Province. Significant correlations with p values ≤ 0.05 are in bold. Data from the instrumented sites were not necessarily available for both years.

(a) 2004-05	n	τ	p
Ekati	6	0.97	0.007
Colomac	6	0.47	0.188
Yellowknife	7	0.39	0.224
Slave Province	19	0.64	0.001
(b) 2005-06	n	τ	p
Ekati	6	0.87	0.015
Colomac	2	N/A	N/A
Yellowknife	7	0.49	0.129
Slave Province	15	0.65	0.001

- (4) South of treeline, shaded sites had lower thawing surface offsets than open sites. However there was no significant difference in thawing surface offsets between the shaded sites and the tundra sites north of treeline at Ekati. This suggests that vegetation does not directly control thawing surface offsets through shading but may be a component of the mechanism that does.
- (5) Total active-layer water controlled the spatial variability in the freezing surface offset in the discontinuous permafrost zone where thick snow covers retarded the freezing of the active layer, but not in the continuous permafrost zone, where thin snow covers allowed rapid freezing of the active layer.
- (6) Annually, air and surface temperatures decreased at different rates with increasing latitude, resulting in anomalously large annual surface offsets at the northern boundary of the discontinuous permafrost zone.

Chapter Seven: The Thermal Offset: Relations between Surface and Permafrost Temperatures in the Slave Geological Province

7.1   Introduction

In Chapter Six, relations between air temperature (T_A) and ground surface temperature (T_S) were examined south of treeline at Yellowknife and Colomac, and north of treeline at Ekati. Distinct differences across treeline in the freezing surface offset ($S.O._f$) and n-factor (n_f) were caused by differences in snow cover. On the tundra where snow cover was thin, snow depth controlled the variation in n_f . In the boreal forest south of treeline where snow was thick, n_f was controlled by the moisture content of the active layer. Conversely, the thawing surface offset ($S.O._t$) and thawing n-factor (n_t) were similar across treeline despite differences in vegetation.

This chapter examines the thermal regime of the active layer through the thermal offset (T.O.), the difference between the annual mean temperature at the ground surface (T_{Sa}) and at the top of permafrost (T_{TOPa}). The T.O. is a function of rk , the ratio of thawed to frozen thermal conductivity (k_t and k_f respectively) in the active layer. Field measurements of T_S , T_{TOP} , and soil moisture were used to examine the thermal offset (Appendix Tables B.12 and B.13). The following research questions are addressed in this chapter:

- (4) What is the dominant mode of heat transfer in the active layer at the instrumented sites?
- (5) What is the variation in thermal offsets within and among the study areas?

- (6) What is the effect of the unusually high air temperatures in 2005-06 on the thermal offset?
- (7) What controls the thermal offset?
- (8) Can r_k be determined from field measurements?

7.2 Active Layer Thermal Regime

Relations between T_S and T_{TOP} varied throughout the year at all the instrumented sites (Figure 7.1). At the beginning of the freezing season, the entire active layer and the top of permafrost were isothermal, just below 0 °C as latent heat was released from freezing soil water. T_S fluctuated slightly below 0 °C once the surface was frozen but did not drop below -5 °C during the active-layer freezeback period. The freezeback phase of the freezing season was described in Chapter Five. Following freezeback period, T_S and T_{TOP} declined as sensible heat flowed out of the ground to the cold overlying air. During the freezing season, differences between T_S and T_{TOP} were larger at Ekati than south of treeline at Yellowknife and Colomac. Daily differences between T_S and T_{TOP} at Ekati were occasionally greater than 8 °C, and were generally around 5 °C. South of treeline daily differences between T_S and T_{TOP} were only occasionally greater than 5 °C.

In spring when T_A rose above 0 °C, T_S and T_{TOP} increased rapidly as T_S exceeded T_{TOP} but remained just below 0 °C until snowmelt was complete. During the thawing season T_S was above 0 °C, while T_{TOP} slowly increased to 0 °C, or remained at 0 °C if freezeback comprised the entire previous winter. On a daily basis, T_S values were regularly between 5 °C and 15 °C higher than T_{TOP} during the thawing season. The daily differences between T_S and T_{TOP} were slightly lower at Ekati than south of treeline at Yellowknife and Colomac.

Figure 7.1 Daily mean temperature at the surface and top of permafrost from 1 September 2004 to 31 September 2006 at (a) Ekati (E-1a) and (b) Yellowknife (Y-1a). Daily means were calculated from 12 observations. Surface temperatures were measured 5 cm below the radiative surface at the base of the live moss, and temperatures at the top of permafrost were measured at 100 cm.

7.2.1 Nonconductive Heat Transfer

The TTOP Model assumes that heat flow in the active layer is entirely conductive (Smith and Riseborough 1996). However, nonconductive heat flow in the active layer can be substantial, especially if associated with phase change of water (Kane *et al.* 2001). To identify episodes of nonconductive heat flow in the active layer at the instrumented sites, changes in the thermal gradient between 20 and 50 cm were examined (Burn 2000). Figure 7.2 presents the difference in the gradient every two hours – the frequency of ground temperature measurements. Under conductive heat flow, the change in the gradient over time would be either 0 if the gradient were constant, or ± 0.015 to 0.024 °C/cm – a function of the thermistors' precision. Sudden changes in the temperature gradient between 20 and 50 cm of 0.02 to 0.1 °C/cm correspond to temperature changes of 0.5 to 3.4 °C respectively, and indicate nonconductive heat transfer events.

There were few nonconductive heat flow events at the instrumented sites during the freezing season. In unsaturated soils, forced convection was minimal due to the presence of snow, and free convection was limited by low porosity. Heat transfers through advected water were minimal because the sites were located in basins with no relief to promote lateral movement of water in the active layer. Following freezeback, the movement of unfrozen water in soils less than 0 °C was limited by low ground temperatures at Ekati, and by the low unfrozen water content of peat at the Yellowknife and Colomac sites. During freezeback, nonconductive heat transfers could occur between the 0 °C isotherms, where water would be above 0 °C. However, these transfers would be internal to the system. Water could also migrate and transfer heat at Ekati

Figure 7.2 Change to the thermal gradient between 20 and 50 cm every 2 hours from 1 September 2004 to 31 August 2006 at (a) Ekati (E-4b), (b) Colomac (C-3b), and (c) Yellowknife (Y-2c).

during freezeback behind the 0 °C isotherms because the active layers are predominantly composed of silt, which typically has high unfrozen water content at temperature above -1 °C. These nonconductive heat transfers were limited to the freezeback period, which was only 30 % of the freezing season at Ekati.

Nonconductive heat transfer was more prevalent during the thawing seasons when surface and soil water were unfrozen, and the soil water vapour pressure gradients were steep in the unsaturated active layers. Nonconductive heat transfer associated with snowmelt in spring and rain events in late summer was identified at several of the instrumented sites between 20 and 50 cm (Figure 7.2). On average, 5 to 10 nonconductive events per year were identified at the instrumented sites, with wet sites experiencing far fewer events than dry sites. Low intensity nonconductive heat transfer events may be more pervasive than this analysis suggests, especially if they were brief and not captured at the 2 hour temperature measurement frequency, or if they were less intense and not captured at the resolution of the thermistors. Nonconductive fluxes of heat such as internal distillation that are pervasive but subtle may be important to the thermal regime of the active layer but are difficult to quantify (Hinkel and Outcalt 1994). These processes are included in determination of the apparent thermal conductivity.

7.3 The Thermal Offset

The annual thermal offset (T.O.) is the difference between T_{TOPa} and T_{Sa} (i.e. $T_{TOPa} - T_{Sa}$). Where the ground thaws seasonally at the surface, T_{TOPa} is lower than T_{Sa} , and T.O. is negative, with the lowest values representing large differences in temperature across the active layer. Thermal offsets were calculated for the organic instrumented sites from values of T_{TOPa} and T_{Sa} measured at the sites in 2004-05 and 2005-06. The

annual period used to determine T_{TOPa} and T_{Sa} was defined by surface temperature. The thermal offset was not examined at Y-5 because permafrost was absent, nor at E-5 because measurements of T_{TOP} were not recorded.

Unfortunately, T.O. data were not available for several sites, especially in 2005-06, due to logger failure or radiant heating of the surface sensor. Values of T.O. were collected from 21 sites in total, from 18 sites in 2004-05 and 11 sites in 2005-06. Only 8 sites had T.O. values for both years. As a result, pair-wise analysis could not be used to determine interannual differences in T.O. The nonparametric Mann-Whitney U statistic was used to determine differences in T.O. between the three study areas (Hammond and McCullagh 1978).

Across the Slave Province, T.O. ranged from 0.0 to -3.1 °C in 2004-05 when T_A were near normal (Figure 7.3). Within each of the study areas that year, T.O. varied considerably with ranges over 1.0 °C. There were no distinct differences north and south of treeline as there were for T_{Sa} and T_{TOPa} . Rather, differences in T.O. among the study areas were similar to those of T_{St} , with progressively smaller, less negative, T.O. values going north. In 2004-05, median T.O. at Yellowknife, Colomac and Ekati were -2.5, -1.3, and -0.6 °C respectively, and T.O. at Yellowknife were significantly different from both Ekati ($U=0$, $n=6$ & 6 , $p=0.002$) and Colomac ($U=1.0$, $n=6$ & 6 , $p=0.004$). Values of T.O. at Colomac and Ekati were not significantly different ($U=7.0$, $n=6$ & 6 , $p=0.093$), despite differences in median values.

The following year, when T_A , T_S , and T_{TOP} were higher, T.O. values were lower (larger negative numbers) because T_S increased more than T_{TOP} . Of the 8 sites with T.O. values for both years, T_S and T_{TOP} increased by median values of 2.1 °C and 1.0 °C

Figure 7.3 Annual thermal offset (T.O.) at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. Annual period was defined by surface temperature. Values for the two years were not necessarily for the same sites.

respectively, while T.O. decreased by a median value of 1.0 °C. Unfortunately, there were insufficient data in 2005-06 to determine statistical differences among the three study areas. However, T.O. values at Yellowknife were at least 0.5 °C less (more negative) than at Ekati in both years, suggesting a similar trend going north during the abnormally warm year.

The thermal offset model under seasonally thawed conditions is:

$$T.O. = \frac{DDT_s (rk-1)}{P} \quad [7.1]$$

where, DDT_s is the thawing degree-days at the surface, rk is the ratio of ground thermal conductivity in the thawed to frozen state, and P is the annual period (Romanovsky and Osterkamp 1995). The timing of the seasons used to determine DDT_s and P were defined by surface temperature. Figure 7.4 presents the negative relation between T.O. and thawing degree-days at the surface (DDT_s). The relation between T.O. and DDT_s is significant in both 2004-05 ($\tau = -0.73$; $p < 0.000$) and 2005-06 ($\tau = -0.66$; $p = 0.005$), with larger T.O. values related to warmer surface temperatures in summer. This spatially generated relation supports the thermal offset model that directly relates T.O. to surface temperatures in summer under equilibrium conditions. These results support the hypothesis that increasing T_A through climate change summer warming will increase the magnitude of T.O., which would make permafrost more resistant to warming (Smith and Riseborough 2002).

The T.O. is thought to be an important control of permafrost distribution at its southern boundary, where MAAT is close to 0 °C (Smith and Riseborough 2002). However, these results suggest that T.O. also controls permafrost distribution at the northern extent of discontinuous permafrost. Yellowknife lies in the extensive

Figure 7.4 Relation between annual thermal offset (T.O.) and thawing degree-days at the surface (DDT_s) at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=-0.73$, $p<0.000$) and (b) 2005-06 ($\tau=-0.66$, $p=0.005$).

discontinuous permafrost zone, where climate is thought to be sufficient to cause permafrost in 50 to 90% of the landscape (Heginbottom *et al.* 1995). However, at Yellowknife, permafrost is absent in bedrock, and exists in peatlands under steep thermal offsets. At Yellowknife, T.O. was sufficient to support permafrost at T_{sa} above 0 °C in both years. This suggests that climate is not conducive to permafrost at Yellowknife, and that the thermal offset is required to support permafrost even at mean annual air temperatures well below 0 °C. According to Shur and Jorgenson's (2007) classification of the climate-permafrost relation, permafrost at Yellowknife is described as "climate-driven, ecosystem-protected". This class of permafrost is characterized by subarctic climate and late ecosystem succession, and does not regenerate after disturbance causes degradation.

7.4 Thermal Conductivity Ratio (rk)

According to Equation 7.1, T.O. is also directly related to rk – a physically-based parameter that summarizes changes to the thermal conductivity of the active layer in summer due to the thawing of soil water. rk is a function of soil water alone and is unrelated to climate. The slope of the linear least squares line through the relation between T.O. and DDT_s in Figure 7.4, was used to determine the rk value for the Slave Province using the thermal offset model rearranged as:

$$rk = \left(\frac{T.O.}{DDT_s} \times P \right) + 1 \quad [7.2]$$

The annual period ranged from 352 to 375 days among the instrumented sites and between the two years, thus the standard period of 365 days was used in this calculation. The slopes (T.O./ DDT_s) were similar in both years, approximately -0.0033 days⁻¹, and yielded rk values of -0.20, which are inappropriately negative. If a positive relation

between T.O. and DDT_s was assumed using the absolute value of T.O., the thermal offset model produced positive rk values of 2.2, which are also inappropriate as it implies larger k_i than k_f . The thermal offset model did not describe the thermal offsets on a regional scale in the Slave Province.

7.4.1 $rk_{Temperature}$

Equation 7.2 was also used to determine rk at each of the instrumented sites where, DDT_s , and P were defined by surface temperature. These rk values are referred to as $rk_{Temperature}$ because they were derived directly from temperature data, particularly the T.O. determined from field measurements of T_{Sa} and T_{TOPa} .

Across the Slave Province, $rk_{Temperature}$ values ranged from 0.08 to 1.02 in 2004-05 (Figure 7.5). Similar to T.O., there was considerable variation in $rk_{Temperature}$ within the study areas, and though the ranges overlapped, the median values were lowest at Yellowknife and highest at Ekati. In 2004-05, median $rk_{Temperature}$ at Yellowknife, Colomac and Ekati were 0.22, 0.49, and 0.59, respectively. Values of $rk_{Temperature}$ at Yellowknife had a smaller range and were significantly different from values at Colomac ($U=3$, $n=6\&6$, $p=0.015$), and Ekati ($U=0$, $n=6\&6$, $p=0.002$). The following year when T_A was higher, $rk_{Temperature}$ values were considerably lower, ranging from 0.00 to 0.49 across all the instrumented sites, with median values of 0.05, and 0.25 for Yellowknife and Ekati respectively.

Site E-3a had a particularly high $rk_{Temperature}$ value of 1.02, resulting from a low T.O. value of 0.0 °C. High values of rk and low values of T.O. imply dry conditions, such as bedrock, in which the thermal conductivity of the active layer does not change between seasons because of the freezing of soil water. Soil water was observed at

Figure 7.5 Thermal conductivity ratio based on temperatures ($rk_{Temperature}$) at Yellowknife, Colomac and Ekati for 2004-05 and 2005-06. The annual thermal offset (T.O.), thawing degree-days at the surface (DDT_S), and annual period (P) were used to calculate $rk_{Temperature}$. Surface temperature was used to define P, T.O., and DDT_S. Values for the two years were not necessarily for the same sites.

E-3a, though this site was drier than others at Ekati. Site E-3b, less than 50 m away from E-3a, also had relatively high values of $rk_{Temperature}$ and T.O. The thermal conductivity ratio is a measure of the change in thermal conductivity due to the freezing of soil water, it is therefore a function of soil water content of the active layer. Across the Slave Province there were significant negative relations between $rk_{Temperature}$ and total active-layer water in 2004-05 ($\tau=-0.62$, $p=0.000$), and 2005-06 ($\tau=-0.62$, $p=0.008$) (Figure 7.6). The relation between $rk_{Temperature}$ and total active-layer water supports the claim that rk is a physically-based parameter independent of climate. According to the thermal offset model, an active layer composed of saturated peat should have an rk value of approximately 0.30, regardless of where that peat is located.

In the warm year, five of eleven sites had $rk_{Temperature}$ values less than 0.05. Such values challenge the thermal offset model, because improbable changes in moisture conditions between the freezing and thawing seasons are required to effect such values for rk . The collapse of the thermal offset model in the second warmer year suggests that the rk parameter is dependent on climate being under equilibrium conditions.

7.5 Field Measurements of rk

The thermal offset model has been validated on the North Slope of Alaska (Romanovsky and Osterkamp 1995), and been used to reproduce the permafrost map of Canada (Henry and Smith 2001). These results are promising and suggest that the thermal offset model may be used to determine permafrost conditions at larger map scales. Values of rk were determined from soil samples collected in the field, using Johansen (1975) to calculate k_t and k_f . These values of rk are referred to as rk_{Soil} as they

Figure 7.6 Relation between thermal conductivity ratio based on temperatures ($rk_{Temperature}$) and total active-layer soil water at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=-0.62$, $p<0.000$), and (b) 2005-06 ($\tau=-0.62$, $p=0.008$).

were determined from physical soil properties. The rk parameter is sensitive to soil moisture, which varies spatially, temporally and with depth. This variation was considered in determining rk_{Soil} . Soil surveys were conducted in September 2005 and 2006, and moisture conditions at the time of sampling were assumed representative of that summer and the previous winter. Thus, k_t was calculated based on the moisture conditions at the time of the soil survey, and k_f was calculated by assuming that the soil water at the time of sampling was present as ice the previous winter. This assumption captures the spatial variability in moisture conditions.

7.5.1 Thermal conductivity measurements

In 2006, thermal conductivity measurements were also made in situ prior to extraction of the soil samples using the TPSYS02 Thermal Conductivity Measurement System by Hukseflux®. The system is based on the Non-Steady-State technique using a probe that contains both a heating wire and a temperature sensor. The probe is inserted into the soil, a heating step is applied, and the thermal resistivity (the inverse of conductivity) can be calculated (Hukseflux 2003). The manufacturer states that resolution of the probe was $0.02 \text{ Wm}^{-1}\text{K}^{-1}$ and the accuracy was 2%. However these specifications are based on a good contact between the probe and the soil. A Hukseflux measurement was taken for every soil sample extracted so that the estimate of thermal conductivity from the soil sample could be compared with that from the probe.

There was considerable scatter about the 1:1 line between k_t measurements from the Hukseflux and k_t estimates using Johansen's method (Figure 7.7). This scatter could be due to poor estimates of soil water or bulk density from the soil samples, poor performance of Johansen's method, or inaccurate Hukseflux measurements.

Figure 7.7 Measurements of thermal conductivity obtained using Hukseflux TPSY02 Thermal Conductivity Measurement System compared with estimates of thermal conductivity obtained from soil samples using Johansen's Model (1975).

Hukseflux measurements of k_t were more than $0.5 \text{ Wm}^{-1}\text{K}^{-1}$ greater than estimates of k_t using Johansen's method for approximately 10% of the samples. These high values probably represent nonconductive heat transfer. Similarly, at the surface where k_t were typically low for organic soils with low bulk densities, instrument measurements were systematically higher than estimates by Johansen's method. There was no systematic difference between the probe and Johansen's values for k_t between 0.3 and $0.8 \text{ Wm}^{-1}\text{K}^{-1}$. Of the 145 pairs of k_t values, 45% had differences less than $0.1 \text{ Wm}^{-1}\text{K}^{-1}$, and 70% had differences less than $0.2 \text{ Wm}^{-1}\text{K}^{-1}$. In view of the apparent questions regarding instrument accuracy, Johansen's estimates of k_t were used in determining rk_{Soil} .

7.5.2 rk_{Soil}

Thermal properties of the active layer were considered for three layers – the Surface Layer, the Middle Layer, and the Base Layer. At sites where samples were collected from multiple depths, samples were chosen to represent these three layers (see section 4.4). Values of k_t and k_f varied considerably across these three layers due to differences in soil water, bulk density, and composition. Across the Slave Province, values of k_t and k_f increased with depth, and were higher in 2006 than in 2005 due to increases in soil moisture. Values of rk decreased with depth at Yellowknife and Colomac but remained constant throughout the active layer at Ekati (Figure 7.8). Across all three study areas, the median value of rk for the Surface Layer was 0.60 in 2005. At the Base Layer, median values of rk were 0.47 south of treeline, and similar to the Surface Layer north of treeline. The following year, values of rk for all three layers across the Slave Province were lower due to higher moisture contents, but again rk values

Figure 7.8 Thermal conductivity ratio (r_k) values for the Surface, Middle and Base Layers of the active layers at Yellowknife, Colomac and Ekati. Soil samples were collected in September 2005 and 2006.

were much lower for the Base Layer than for the Surface Layer at Yellowknife and Colomac. Low rk values represent high soil water contents, where k_f greatly exceeds k_t . Values of rk were regularly as low as 0.30 at the base of the active layers south of treeline.

The thermal offset model assumes a single value of rk for the active layer, which is presumably based on single values of k_t and k_f . At each site, the different thermal conductivities of the three layers were averaged into a single value for the active layer using the harmonic mean, weighted with the thickness of each layer:

$$\frac{x_{A.L.}}{k_{A.L.}} = \frac{x_S}{k_S} + \frac{x_M}{k_M} + \frac{x_B}{k_B} \quad [7.3]$$

where, $x_{A.L.}$, x_S , x_M , and x_B are the thicknesses of the active layer, Surface, Middle, and Base Layers respectively, and $k_{A.L.}$, k_S , k_M , and k_B are the thermal conductivities of the active layer, Surface, Middle, and Base Layers respectively. The weighted harmonic mean was used to calculate bulk kt and kf from the three layers, and their ratio was determined to obtain rk_{Soil} values for the active layers. The harmonic mean was chosen to summarize thermal conductivities of the active layer because it describes energy flow through layers arranged in series, which is analogous to heat flow perpendicular to layers in the active layer (Farouki 1981). The harmonic mean is sensitive to low values, such as k_t for the Surface Layer, which limit heat entering the ground in summer.

Average k_t and k_f values for the three layers were also obtained using the arithmetic mean, the weighted arithmetic mean, and the harmonic mean without weights. Compared to the other averages, values of rk_{Soil} determined using the weighted harmonic mean covaried strongly with values of $rk_{Temperature}$, and were similar to values of

$rk_{Temperature}$ for several sites. Values of rk_{Soil} based on 6 layers as opposed to 3 layers were calculated for 5 sites in 2006 and the difference was less than 0.03.

Values of rk_{Soil} varied within and among the study areas and between the two years of study (Figure 7.9). Across the Slave Province values of rk_{Soil} ranged from 0.47 to 0.73 in 2005. The following year when T_A were higher and active layers were thicker and wetter, values of rk_{Soil} were lower at all the instrumented sites, and ranged between 0.32 and 0.69. In both years, values of rk_{Soil} were similar at Yellowknife and Colomac and higher at Ekati (Table 7.1). This difference in rk_{Soil} across treeline is expected due to the higher porosity and subsequent higher water content of the active layers composed entirely of peat south of treeline at Yellowknife and Colomac.

7.6 Comparison of rk_{Soil} and $rk_{Temperature}$

In 2005, the relation between rk_{Soil} and $rk_{Temperature}$ was strong ($\tau=0.52$, $p=0.003$) (Figure 7.10(a)). However, the range of $rk_{Temperature}$ was almost four times larger than the range of rk_{Soil} . Values of rk_{Soil} were similar, or within 0.12, at 8 of the 18 instrumented sites for 2005. All of these sites were either at Ekati or Colomac. Values of $rk_{Temperature}$ were lower than rk_{Soil} for the remainder of the sites, including all of the Yellowknife sites.

Values of rk_{Soil} should approach those of $rk_{Temperature}$ if: (1) temperature measurements of the thermal offset were accurate; (2) methods to estimate rk from soil samples are adequate, and; (3) the thermal offset model is valid. The close agreement between $rk_{Temperature}$ and rk_{Soil} at the majority of the Ekati and Colomac sites suggests that field measurements of the thermal offset, and the thermal conductivities were appropriate for both mineral and peat sites, and that the thermal offset model is robust under normal

Figure 7.9 Thermal conductivity ratio based on field measurements of soil (rk_{Soil}) at Yellowknife, Colomac and Ekati. Soil samples were collected throughout the active layers in September 2005 and 2006. Values of k_t and k_f were obtained using Johansen (1975), and the weighted harmonic mean was used to summarize k_t and k_f for the active layer.

Table 7. 1 Results of Mann-Whitney U test to determine statistical difference in rk_{Soil} between study areas in 2005 and 2006. Statistically different pairs are presented in bold for one-tailed test, confidence level of 0.05. Values of rk_{Soil} were higher at Ekati in both years.

	2005		2006	
rk_{Soil}	Colomac (n=8)	Yellowknife (n=8)	Colomac (n=8)	Yellowknife (n=8)
Ekati (n=7)	U=12; p=0.036	U=9; p=0.014	U=1; p<0.000	U=10; p=0.020
Yellowknife (n=8)	U=20; p=0.117	X	U=24; p=0.221	X

Figure 7.10 Relation between thermal conductivity ratio based on temperatures ($rk_{Temperature}$) and thermal conductivity ratio based on soil samples (rk_{Soil}) at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=0.52$, $p=0.003$), and (b) 2005-06 ($\tau=0.46$, $p=0.050$)

climate conditions at these more northerly study areas. The thermal offset model did not perform well at Yellowknife.

At Yellowknife, values of rk_{Soil} were 57 to 148 % higher than values of $rk_{Temperature}$. The sensitivity of rk to water, and the temporal variability of soil water suggests that the timing of the soil survey in 2005 may not have captured representative moisture conditions of the 2004-05 freezing and thawing seasons at Yellowknife. The low $rk_{Temperature}$ values at Yellowknife in 2004-05 represent large differences in thermal conductivity between the freezing and thawing seasons. For the rk_{Soil} to approach $rk_{Temperature}$ would require either wetter conditions in winter, higher k_f values, or drier conditions in summer, lower k_t values, or both higher k_f and lower k_t values, than the observed conditions.

In 2005-06 when air temperatures were warmer, both $rk_{Temperature}$ and rk_{Soil} were lower, the relation between them was weaker ($\tau=0.46$, $p=0.050$), and their relative distributions differed from that of the previous year (Figure 7.10(b)). Apart from E-3b, which had high $rk_{Temperature}$ values in both years, the ranges of $rk_{Temperature}$ and rk_{Soil} did not intersect as they did the previous year. Values of $rk_{Temperature}$ ranged from 0 to 0.49, while values of rk_{Soil} ranged from 0.36 to 0.69. Similar to the Yellowknife sites the previous year, values of $rk_{Temperature}$ in 2005-06 for the majority of the instrumented sites across the Slave Province are unreasonably low.

Extreme differences in moisture conditions between summer and winter are required to obtain the 2005-06 values of $rk_{Temperature}$. It is possible that for freezing season 2005-06, active layers across the Slave Province were wetter than in September 2006 when the soil surveys were conducted. Both Yellowknife and Ekati received high

amounts of rain in summer 2005, especially in August and September, which would have resulted in high soil moisture contents for that winter. However it is improbable that dry conditions were experienced in summer 2006 as both Yellowknife and Ekati had higher than normal rainfall between June and August. Also, many of the active layers were saturated to varying degrees in September 2006 when the soil surveys were conducted. Across the Slave Province, the moisture conditions required for rk_{Soil} to reach 2005-06 $rk_{Temperature}$ values were not observed nor probable. This suggests that the thermal offset model did not describe the thermal regime of the active layer outside of equilibrium conditions, since T_A in 2005-06 were considerably warmer than normal.

7.7 Latent Heat and the TTOP Model

The relation between $rk_{Temperature}$ and total active-layer water (Figure 7.6), supports the claim that rk is a physically-based parameter that describes seasonal changes in thermal conductivity due to the freezing and thawing of soil water. In 2004-05, the difference between $rk_{Temperature}$ and rk_{Soil} was also related to total active-layer water ($\tau=0.54$, $p=0.002$) (Figure 7.11(a)), and the duration of freezeback ($\tau=0.47$, $p=0.007$) (Figure 7.12(a)). The duration of freezeback is a function of total active-layer water, such that wet active layers release more latent heat and take longer to freeze than dry active layers with similar temperature gradients (Chapter Five). Large differences between $rk_{Temperature}$ and rk_{Soil} at Yellowknife due to unreasonably low values of $rk_{Temperature}$ are associated with wet sites that had long freezeback periods. While the thermal offset is a function of soil water, the thermal offset model does not account for the release and uptake of latent heat that accompanies the freezing and thawing of soil water. Under normal climate conditions in 2004-05, the thermal offset model did not

Figure 7.11 Relation between the difference between thermal conductivity ratio based on temperatures and those based on soil samples ($rk_{Soil} - rk_{Temperature}$) versus total active-layer water at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=0.55$, $p=0.002$), and (b) 2005-06 ($\tau=0.22$, $p=0.349$).

Figure 7.12 Relation between the difference between thermal conductivity ratio based on temperatures and those based on soil samples ($rk_{Soil} - rk_{Temperature}$) versus the active-layer freeze-back period (FBP) as a percentage of the freezing season duration at the Yellowknife, Colomac, and Ekati instrumented sites for (a) 2004-05 ($\tau=0.47$, $p=0.007$), and (b) 2005-06 ($\tau=0.30$, $p=0.217$).

describe field conditions where freezeback comprised more than 40% of the freezing season.

In 2005-06, when T_A values were higher and active layers were wetter, the difference between $rk_{Temperature}$ and rk_{Soil} increased at most sites and ranged between 0.01 and 0.42 because of the unreasonably low values of $rk_{Temperature}$ this year. In this abnormal year, the thermal offset model did not describe the thermal regime of the active layer even at relatively dry sites where freezeback was less than 30 % of the freezing season. Consequently, the difference between $rk_{Temperature}$ and rk_{Soil} was not related to either total active-layer water ($\tau=0.22$, $p=0.349$) (Figure 7.11(b)), or the duration of freezeback ($\tau=0.30$, $p=0.217$) (Figure 7.12(b)). These results suggest that the thermal offset model did not describe the thermal regime of the active layer when the system was out of equilibrium, regardless of moisture regime.

7.8 Summary Points

This chapter examined field values of the thermal offset, and the thermal offset model at the study areas in the Slave Geological Province. The following points summarize the results presented in this chapter:

- (1) Heat is primarily transferred by conduction in the active layers at the instrumented sites. Nonconductive heat transfers associated with infiltration of snowmelt in spring and rain events in late summer were infrequent. These and other pervasive, low-intensity nonconductive heat transfers not captured by the instrumentation, would be included in the apparent thermal conductivity.

- (2) Across the Slave Province, thermal offsets varied from 0.0 to -3.1 °C in 2004-05, and ranged by at least 1.0 °C within the three study areas. Despite distinct differences in active-layer composition across treeline, thermal offset values were progressively higher (less negative) going north, similar to surface temperature in summer.
- (3) In 2005-06 when air, surface, and top of permafrost temperatures were higher, thermal offsets were lower (more negative). Thermal offsets were unstable under transient conditions because surface temperatures increased more than temperatures at the top of permafrost.
- (4) In both years, the thermal offset was inversely related to the surface temperature in the thawing season. This relation is explicit in the thermal offset model.
- (5) In both years, the thermal conductivity ratio determined through the thermal offset model ($rk_{Temperature}$) is inversely related to the total active-layer water. This relation is implicit in the thermal offset model.
- (6) Values of rk for the active layer were determined from soil samples (rk_{Soil}) using a weighted harmonic mean. Under normal climate conditions in 2004-05, $rk_{Temperature}$ compared well to rk_{Soil} at Ekati and several sites at Colomac, but not at Yellowknife. In 2005-06 when temperatures were higher, values of $rk_{Temperature}$ were unreasonably low.
- (7) The difference between $rk_{Temperature}$ and rk_{Soil} was related to total active-layer water and the duration of the active-layer freezeback in 2004-05 but not in 2005-06. The thermal offset model does not describe field conditions in

which freezeback comprises more than 40 % of the freezing season, or when the climate-permafrost system is not in equilibrium.

Chapter Eight: Summary and Conclusions

This thesis examined climate-permafrost relations across treeline using field data from the Slave Geological Province. The surface and thermal offsets, parameterized in the TTOP model, were used as the framework for this investigation. Field data were collected at three study areas in the Slave Province: Yellowknife, Colomac Mine, and Ekati Diamond Mine. The Yellowknife and Colomac study areas lie south of treeline in the extensive discontinuous permafrost zone, while Ekati lies on the tundra north of treeline in the continuous permafrost zone.

Eight instrumented sites were established at Yellowknife and Colomac in treed basin bogs with *Picea mariana*, and thick active layers composed of wet peat. The eight Ekati instrumented sites were located in low-lying areas of high-centred ice-wedge polygons, and had thinner active layers composed of cryoturbated mineral soil with a thin organic layer at the surface. At each instrumented site, measurements of air, ground surface, active layer, and top of permafrost temperature were recorded every two hours using miniature data loggers. Data collection began in September 2004 and spanned two years. In the first year (2004-05), air temperatures across the Slave Province were close to climate normals (1971-2000), and were 4 °C higher on an annual basis in the second year (2005-06).

The research objectives were threefold. The first objective was to determine the variability of the surface and thermal offsets within the three study areas and among them, as well as between the normal and warm years of the study. This investigation examined the relative control of climate and microclimate on permafrost north and south of treeline. The second objective was to identify the physical characteristics that control

the surface and thermal offsets. Of particular interest was the influence of active-layer wetness on the offsets, with respect to release of latent heat during freezeback of the active layer in early winter. The final objective was to evaluate the TTOP model with respect to the first two objectives, and through validation of the thermal conductivity ratio (rk) with field data.

The variability of air, ground surface, and ground temperatures within and among the three study areas was examined in Chapter Five. The study areas lay within the same climate system but had distinctly different annual and seasonal mean air temperatures, which decreased progressively northwards. South of treeline, the ground thermal regime was similar and did not respond greatly to spatial or temporal changes in air temperature. North of treeline, surface and ground temperatures were considerably lower, and the climate-permafrost relation was stronger. These results support Shur and Jorgensen's (2007) understanding that permafrost in the discontinuous zone is largely driven or protected by the microclimate.

As the active layer freezes, ground cooling is inhibited by the release of latent heat. Long freezeback periods result in high mean ground temperatures on an annual basis and for the freezing season. South of treeline, freezeback of the active layer regularly took between 30 and 100 % of the freezing season because of thick snow covers and saturated active layers. North of treeline, where snow covers and active layers are thinner, freezeback did not take more than a third of the freezing season, which allowed the ground to cool for the majority of the winter. The duration of the active-layer freezeback period is the most important factor for variation in the ground thermal regime both within the discontinuous permafrost zone, across treeline, and interannually.

The surface offset and the physical characteristics that control it were examined in Chapter Six. Snow cover controlled the spatial variability in the freezing surface offset across the Slave Province and in the continuous permafrost zone of the tundra, but not in the discontinuous permafrost zone. At the continental scale, snow cover controls the freezing surface offset. However, at local scale south of treeline, thick snow covers prolongs freezeback, and variation in snow cover depth is less important than total active-layer water. Total active-layer water controlled the spatial variability in the freezing surface offset as thick snow covers prolonged the freezeback period. These field results confirm Goodrich's (1982) proposal that active-layer conditions influence surface temperatures in winter, and will impact application of freezing n -factors (n_f) in permafrost mapping endeavours.

The thermal offset and its model were examined in Chapter Seven. Across the Slave Province, thermal offsets varied both within and among the three study areas, similar to surface temperature in summer, despite distinct differences in active-layer composition across treeline. This suggests that local variation is controlled by changes in rk , and regional scale variation is driven by climate. To evaluate the thermal offset model, values of rk determined from soil samples (rk_{Soil}) were compared with values obtained directly from field temperature measurements ($rk_{Temperature}$). The model performed well under normal climate conditions, but only to the north where freezeback comprised less than 40% of the freezing season. This result supports Romanovsky and Osterkamp's (1995) validation of the thermal offset model using data from the north slope of Alaska. The thermal offset model did not describe field conditions where

freezeback comprised more than 40 % of the freezing season, or when the system was out of equilibrium, regardless of moisture conditions.

The results of this research have three implications for the use and operationalization of the TTOP model, particularly its application to mapping permafrost distribution.

- (1) Values of n_f in the discontinuous permafrost zone are controlled by subsurface conditions which are not easily estimated remotely.
- (2) The thermal offset model does not perform well for wet active layers in the discontinuous permafrost zone. Unfortunately, this region is of great interest because permafrost conditions are variable and vulnerable to change under a warming climate.
- (3) The thermal offset model does not perform well for transient conditions, regardless of moisture regime or permafrost temperature. Substantial errors should be expected if it is used to predict changes in permafrost conditions under a warming climate. It may not predict permafrost conditions accurately at the southern margin of permafrost zone where permafrost is “ecosystem-protected” and not in equilibrium with the current climate.

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Appendix A: Vegetation and Soil Surveys

Table A.1 Mean height, and diameter at stem base (DSB) of tall shrubs at the instrumented sites.

Site	Shrub Species	Mean Height (cm)	Mean DSB (cm)	Sample Size
(a) Yellowknife				
Y-1a	<i>Ledum groenlandicum</i>	32	0.5	15
Y-2a	<i>Ledum groenlandicum</i>	29	0.4	15
	<i>Chamaedaphne calyculata</i>	19	0.2	5
Y-2b	<i>Ledum groenlandicum</i>	30	0.5	15
	<i>Chamaedaphne calyculata</i>	17	0.2	15
Y-2c	<i>Ledum groenlandicum</i>	33	0.4	15
Y-3a	<i>Ledum groenlandicum</i>	24	0.4	15
Y-3b	<i>Ledum groenlandicum</i>	31	0.4	15
Y-4a	<i>Ledum groenlandicum</i>	37	0.5	15
Y-4b	<i>Chamaedaphne calyculata</i>	37	0.4	15
Y-5a	<i>Ledum groenlandicum</i>	27	0.4	10
	<i>Alnus crispa</i>	142	4	3
(b) Colomac				
C-1a	<i>Ledum groenlandicum</i>	26	0.4	15
	<i>Betula pumila</i>	65	0.8	15
C-1b	<i>Ledum groenlandicum</i>	27	0.5	15
	<i>Betula pumila</i>	69	0.8	15
C-1c	<i>Ledum groenlandicum</i>	27	0.4	15
	<i>Betula pumila</i>	59	0.6	15
C-2a	<i>Ledum groenlandicum</i>	28	0.4	15
C-2b	<i>Ledum groenlandicum</i>	31	0.4	15
	<i>Betula pumila</i>	41	0.6	15
C-3a	<i>Ledum groenlandicum</i>	29	0.3	15
	<i>Salix arbusculoides</i>	88	0.8	15
C-3b	<i>Ledum groenlandicum</i>	30	0.4	15
	<i>Salix arbusculoides</i>	83	0.8	15
C-3c	<i>Ledum groenlandicum</i>	36	0.4	15
	<i>Salix arbusculoides</i>	82	0.9	15

Table A.1 continued

Site	Shrub Species	Mean Height (cm)	Mean DSB (cm)	Sample Size
(c) Ekati				
E-1a	<i>Betula pumila</i>	36	0.7	15
E-2a	<i>Betula pumila</i>	34	0.6	15
E-2b	<i>Betula pumila</i>	31	0.4	15
E-3a	<i>Betula pumila</i>	40	0.8	15
E-3b	<i>Betula pumila</i>	40	1.0	15
E-4a	<i>Betula pumila</i>	35	0.7	15
E-4b	<i>Betula pumila</i>	31	0.6	15
E-5a	None			
E-5b	<i>Betula pumila</i>	35	0.7	5

Table A.2 Classes used to estimate vegetation cover (after Goldsmith et al. 1986).

Coverage Class	Range of Cover
1	Present
2	< 1 %
3	1-5 %
4	6-15 %
5	16-30 %
6	31-50 %
7	51-75 %
8	76-100 %

Table A.3 Coverage classes for vegetation and ground cover at the Yellowknife instrumented sites.

Species	Y-1a	Y-2a	Y-2b	Y-2c	Y-3a	Y-3b	Y-4a	Y-4b	Y-5a
Trees									
<i>Picea mariana</i>	5	5	3	7	5	7	3	7	5
<i>Pinus banksiana</i>			2						3
<i>Betula papyrifera</i>									1
Tall Shrubs									
<i>Ledum groenlandicum</i>	6	6	6	6	5	6	6		3
<i>Chamaedaphne calyculata</i>		4	5					8	
<i>Alnus crispa</i>									5
Short Shrubs									
<i>Vaccinium vitis-idaea</i>	6	4	5	5	4	5	4		5
<i>Empetrum nigrum</i>			3		5	4	5	3	3
<i>Andromeda polifolia</i>		4	5	3	3	3			
Herbs									
<i>Geocaulon levidum</i>							5		3
<i>Equisetum arvense</i>								2	
<i>Rubus chamaemorus</i>	4	4	5	3	3	3	4	3	
<i>Eriophorum russeolum</i>								3	
<i>Meesia uliginosa</i>	3	3		3			3	6	
<i>Ptilium crista-castrensis</i>									4
Mosses									
<i>Aulacomnium turgidum</i>			3					4	
<i>Polytrichum strictum</i>				3				5	
<i>Sphagnum capilifolium</i>	3	5		3					
<i>Sphagnum fuscum</i>		4	4	5	5	4			
Lichens									
<i>Cladina mitis</i>	5	7	7	7	6	6	4		6
<i>Cladina rangiferina</i>	5		4		6	5		3	3
<i>Cetraria nivalis</i>	5	5	4	4	6	4	3	3	3
<i>Cladonia fimbriata</i>	5	4	3				4	4	
<i>Cladonia cornuta</i>	3	2					3	3	
<i>Hypogymnia physodes</i>	3				5	5	3	3	
<i>Cladonia borealis</i>	4		3	3			3	4	
Non-vegetative									
Exposed mineral soil									2
Exposed organic soil	3						5		
Needle litter	8	3	3	4	4	5	6	8	5
Leaf litter	4	4	4	4	3	4	6	6	6
Layering branches	3	3	3	2	2	3	4	6	3

Table A.4 Coverage classes for vegetation and ground cover at the Colomac sites.

Species	C-1a	C-1b	C-1c	C-2a	C-2b	C-3a	C-3b	C-3c
Trees								
<i>Picea mariana</i>		3	5	5	3	6	7	7
Tall Shrubs								
<i>Betula pumila</i>	3	5	3		5			
<i>Ledum groenlandicum</i>	6	6	6	4	6	4	6	6
<i>Salix arbusculoides</i>						5	4	5
Short Shrubs								
<i>Vaccinium caespitosum</i>			3					
<i>Vaccinium vitis-idaea</i>	5	5	5		5	4	4	3
<i>Arctostaphylos rubra</i>				6		5	5	5
<i>Empetrum nigrum</i>				6		5	5	5
<i>Salix herbacea</i>				1	4			
<i>Vaccinium uliginosum</i>			4	5	4	6	7	6
<i>Andromeda polifolia</i>	6		4			5	4	4
Herbs								
<i>Equisetum scirpoides</i>				4		3		3
<i>Equisetum arvense</i>		3						
<i>Rubus chamaemorus</i>	5	5	6		5	4	4	4
<i>Eriophorum russeolum</i>				5	4			
Grasses (30-40cm tall)				7		3	4	4
<i>Meesia uliginosa</i>	4	5	6					
<i>Dicranum polysetum</i>		3	3	5	4	3	4	4
<i>Polytrichum strictum</i>	5		3		3			
Mosses								
<i>Polytrichum commune</i>		4	5	5	5			
<i>Sphagnum capilifolium</i>		3	3			4	6	6
<i>Sphagnum fuscum</i>			4	5	3			
Lichens								
<i>Cladina mitis</i>	7	6	7	5	5	4	3	4
<i>Cladina rangiferina</i>			5	5		4	3	
<i>Cetraria nivalis</i>	5	4	6	5	5			4
<i>Cladonia fimbriata</i>	5		4	5	5			
<i>Cladonia cornuta</i>					4		4	
<i>Hypogymnia physodes</i>				4		3		
<i>Cladonia borealis</i>	4	4	4		3			
Non-vegetative								
Needle litter		5	7	4		6	6	6
Leaf litter	5	6	7	3	6	6	6	6
Layering branches		2	2	1		2	3	3

Table A.5 Coverage classes for vegetation and ground cover at the Ekati instrumented sites.

Species	E-1a	E-2a	E-2b	E-3a	E-3b	E-4a	E-4b	E-5a	E-5b
Tall Shrubs									
<i>Betula pumila</i>	6	4	5	6	6	6	6		5
Short Shrubs									
<i>Vaccinium vitis-idaea</i>	7	6	5	3	5	5	6		4
<i>Ledum decumbens</i>	7	6	5	7	5	4	5		4
<i>Arctostaphylos ruba</i>									3
<i>Empetrum nigrum</i>	4	2	3		3	4			5
<i>Vaccinium uliginosum</i>	5	5			3	4	4		
<i>Andromeda polifolia</i>	5	3	3	3			3		
Herbs									
<i>Rubus chamaemorus</i>		4	4						
<i>Eriophorum russeolum</i>	6	5	6	4	6	7	5		
<i>Pedicularis labradorica</i>	1	1	2			2			
Mosses									
<i>Sphagnum grigensohnii</i>	5								
<i>Aulacomnium turgidum</i>	4	3	2	4	3		3		
<i>Polytrichum strictum</i>		1	2	2	2				
<i>Sphagnum augustifolium</i>		3	2	4	3	5	3		
<i>Rhacomitrium lanuginosum</i>	4	3	3	3	4	3	3		
<i>Sphagnum capilifolium</i>							4		
Lichens									
<i>Cladina mitis</i>	4	4	4	3	3	4	2		4
<i>Cladina rangiferina</i>					2	3	2		
<i>Cetraria nivalis</i>	3	4	4	4	2	5	3		3
<i>Cladonia fimbriata</i>				3	2	2	3		
<i>Cladonia cornuta</i>		2		2		2			
<i>Hypogymnia physodes</i>				2	3				
<i>Cladonia borealis</i>			1	3					
Non-vegetative									
Exposed mineral soil				4	3	3		8	7
Exposed organic soil	1		2	5	3	4			
Leaf litter	6	6	4	6	7	6	6		

Table A.6 Soil characteristics at the Yellowknife instrumented sites. Samples collected in September 2006.

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
Y-1a	5	Moist, fibric peat	0.11	0.50
	5	Moist, fibric peat	0.08	0.38
	10	Moist, fibric peat	0.06	0.49
	10	Moist, fibric peat	0.07	0.56
	20	Wet, mesic peat	0.10	0.78
	20	Wet, mesic peat	0.08	0.85
	30	Wet, mesic peat	0.09	0.77
	30	Wet, mesic peat	0.10	0.61
	40	Water table		
	40	Saturated, mesic peat	0.14	0.68
	40	Saturated, mesic peat	0.12	0.71
	50	Saturated, mesic peat	0.18	0.73
	50	Saturated, mesic peat	0.21	0.77
Y-2a	5	Moist, mesic peat	0.09	0.52
	5	Moist, mesic peat	0.08	0.46
	10	Moist, fibric peat	0.06	0.60
	10	Moist, fibric peat	0.08	0.46
	20	Wet, fibric peat	0.06	0.70
	20	Wet, fibric peat	0.07	0.77
	25	Water table		
	30	Saturated, fibric peat	0.11	0.85
	30	Saturated, fibric peat	0.11	0.75
	50	Saturated, mesic peat	0.14	0.83
	50	Saturated, mesic peat	0.19	0.83
Y-2b	5	Water table		
	5	Wet, fibric peat	0.12	0.85
	5	Wet, fibric peat	0.07	0.50
	10	Saturated, fibric peat	0.07	0.81
	10	Saturated, fibric peat	0.05	0.64
	20	Saturated, fibric peat	0.09	0.87
	30	Saturated, fibric peat	0.10	0.74
	50	Saturated, mesic peat	0.12	0.69

Table A.6 continued

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
Y-2c	5	Moist, mesic peat	0.08	0.40
	5	Moist, mesic peat	0.09	0.53
	10	Moist, fibric peat	0.12	0.75
	10	Moist, fibric peat	0.09	0.70
	20	Wet, fibric peat	0.09	0.96
	20	Wet, fibric peat	0.06	0.72
	25	Water table		
	30	Saturated, mesic peat	0.12	0.77
	30	Saturated, mesic peat	0.10	1.01
	50	Saturated, mesic peat	0.13	0.72
	50	Saturated, mesic peat	0.26	1.22
Y-3a	5	Moist, fibric peat	0.04	0.27
	10	Moist, fibric peat	0.05	0.39
	20	Moist, fibric peat	0.06	0.37
	30	Moist, fibric peat	0.05	0.40
	50	Moist, fibric peat	0.14	0.61
Y-3b	5	Moist, fibric peat	0.06	0.18
	10	Moist, fibric peat	0.11	0.36
	20	Moist, fibric peat	0.05	0.41
	30	Moist, fibric peat	0.05	0.47
Y-4a	5	Moist, mesic peat	0.11	0.39
	5	Moist, mesic peat	0.10	0.40
	10	Moist, mesic peat	0.12	0.45
	10	Moist, mesic peat	0.15	0.55
	20	Moist, mesic peat	0.19	0.84
	20	Moist, mesic peat	0.25	0.97
	35	Water table		
	50	Saturated, humic peat	0.22	0.76
	50	Saturated, humic peat	0.16	0.71

Table A.6 continued

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
Y-4b	5	Moist, mesic peat	0.10	0.36
	5	Moist, mesic peat	0.12	0.41
	10	Moist, mesic peat	0.09	0.26
	10	Moist, mesic peat	0.11	0.45
	20	Moist, mesic peat	0.21	0.93
	20	Moist, mesic peat	0.16	0.72
	45	Water table		
	50	Wet, humic peat	N/A	N/A
	50	Wet, humic peat	0.23	0.89
Y-5a	5	Moist sand	1.28	0.10
	10	Moist sand		
	20	Dry sand	1.53	0.03
	50	Dry sand	1.50	0.04

Table A.7 Soil characteristics at the Colomac instrumented sites. Samples collected in September 2006.

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
C-1a	5	Moist, fibric peat	0.13	0.49
	20	Moist, fibric peat	0.13	0.69
	50	Moist, fibric peat	0.24	0.93
C-1b	5	Moist, fibric peat	0.17	0.78
	20	Moist, fibric peat	0.12	0.80
	50	Moist, fibric peat	0.14	0.86
C-1c	5	Moist, fibric peat	0.09	0.31
	20	Moist, mesic peat	0.17	0.85
	40	Moist, mesic peat	0.40	0.87
C-2a	5	Moist, fibric peat	0.08	0.51
	10	Moist, fibric peat	0.06	0.52
	20	Moist, fibric peat	0.05	0.53
	50	Moist, mesic peat	0.24	0.88
C-2b	5	Moist, fibric peat	0.10	0.39
	10	Moist, fibric peat	0.09	0.62
	20	Moist, mesic peat	0.18	1.14
	50	Moist, mesic peat	0.09	0.77
C-3a	5	Moist, fibric peat	0.05	0.36
	5	Moist, fibric peat	0.04	0.23
	10	Moist, fibric peat	0.08	0.79
	10	Moist, fibric peat	0.08	0.63
	15	Water table		
	20	Saturated, mesic peat	0.13	1.07
	20	Saturated, mesic peat	0.13	0.83
	30	Saturated, mesic peat	0.21	0.99
	30	Saturated, mesic peat	0.20	0.99

Table A.7 continued

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
C-3b	5	Moist, fibric peat	0.09	0.39
	5	Moist, fibric peat	0.04	0.26
	10	Moist, mesic peat	0.09	0.41
	10	Moist, mesic peat	0.06	0.43
	20	Wet, mesic peat	0.17	0.97
	20	Wet, mesic peat	0.13	0.80
	25	Water table		
	30	Saturated, mesic peat	0.14	0.97
	30	Saturated, mesic peat	0.16	0.82
C-3c	5	Moist, fibric peat	0.04	0.17
	5	Moist, fibric peat	0.09	0.30
	10	Moist, mesic peat	0.07	0.26
	10	Moist, mesic peat	0.11	0.43
	20	Moist, mesic peat	0.14	0.73
	20	Moist, mesic peat	0.16	0.85
	30	Water table		
	30	Wet, mesic peat	0.18	0.90
	30	Wet, mesic peat	0.15	0.94

Table A.8 Soil characteristics at the Ekati instrumented sites. Samples collected in September 2006.

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
E-1a	5	Moist organics	0.09	0.43
	5	Moist organics	0.08	0.43
	5	Moist organics	0.09	0.46
	7	Moist organics	0.06	0.42
	13	Moist organics	0.14	0.64
	13	Moist organics	0.56	0.66
	18	Moist turbic organics and mineral	0.25	0.72
	19	Moist turbic organics and mineral	0.53	0.66
	33	Wet turbic mineral	1.20	0.88
	25	Wet turbic mineral	0.85	0.69
	34	Wet turbic mineral	0.51	0.84
	40	Wet turbic mineral	0.43	0.59
E-2a	3	Moist organic and mineral	0.09	0.26
	3	Moist organic and mineral	0.07	0.23
	8	Moist organic and mineral	0.06	0.34
	8	Moist organic and mineral	0.20	0.70
	15	Moist poorly sorted mineral	0.26	0.78
	15	Moist poorly sorted mineral	1.33	0.09
	30	Moist poorly sorted mineral	0.52	0.51
	30	Moist poorly sorted mineral	1.24	0.07
	40	Wet poorly sorted mineral	1.53	0.16
	40	Wet poorly sorted mineral	1.35	0.09
	50	Wet poorly sorted mineral	1.57	0.12
	50	Wet poorly sorted mineral	1.33	0.12
	60	Wet poorly sorted mineral	1.38	0.24
E-2b	5	Moist organics	0.09	0.45
	5	Moist organics	0.27	0.79
	15	Wet poorly sorted mineral	0.66	0.89
	15	Wet poorly sorted mineral	0.66	0.14
	25	Wet poorly sorted mineral	1.23	0.31
	30	Water table and cobbles		

Table A.8 continued

Site	Depth	Description	Bulk Density (g cm ⁻³)	Volumetric Water Content (cm ³ cm ⁻³)
E-3a	5	Wet organics	0.07	0.41
	5	Wet organics	0.10	0.51
	10	Wet turbic organics and mineral	0.11	0.51
	10	Wet turbic organics and mineral	0.16	1.00
	15	Wet turbic organics and mineral	0.21	0.76
	15	Wet turbic organics and mineral	1.02	0.66
	20	Wet turbic organics and mineral	0.22	0.85
	20	Wet turbic organics and mineral	1.07	0.65
	27	Water table		
	30	Saturated mineral	0.23	0.81
	30	Saturated mineral	1.09	0.70
E-3b	5	Moist turbic organics and mineral	0.40	0.32
	5	Moist turbic organics and mineral	0.26	0.48
	15	Wet turbic mineral	0.92	0.69
	15	Wet turbic mineral	0.45	0.70
	25	Wet turbic mineral	0.85	0.69
	25	Wet turbic mineral	0.79	0.67
E-4a	3	Wet organics	0.10	0.58
	5	Wet organics	0.13	0.88
	10	Wet organics	0.12	0.88
	15	Wet turbic organics and mineral	0.25	0.78
	20	Wet turbic organics and mineral	0.71	0.70
	30	Wet mineral	0.54	0.68
E-4b	3	Moist organics	0.07	0.45
	10	Moist organics	0.13	0.73
	20	Wet turbic organics and mineral	0.27	0.77
	30	Wet mineral	0.96	0.49
E-5a	3	Dry sand	1.33	0.04
	5	Dry sand	1.25	0.06
	15	Dry sand	1.31	0.04
	25	Damp sand	1.40	0.11
E-5b	N/A			

Appendix B: Temperature Data

Table B.1 Mean air temperatures for the (a) freezing season (T_{Af} ; 1 September to 30 April), (b) thawing season (T_{At} ; 1 May to 31 August), and (c) annual period (T_{Aa} ; 1 September to 31 August) at the instrumented sites and the Yellowknife and Ekati airports.

	2004-05			2005-06		
	$T_{Af}(^{\circ}\text{C})$	$T_{At}(^{\circ}\text{C})$	$T_{Aa}(^{\circ}\text{C})$	$T_{Af}(^{\circ}\text{C})$	$T_{At}(^{\circ}\text{C})$	$T_{Aa}(^{\circ}\text{C})$
Ekati						
E-1a	-19.6	4.6	-11.4	-15.2	8.8	-7.1
E-4a	-19.3	4.6	-11.3	-14.9	8.8	-6.9
Airport	-18.7	4.8	-10.8	-14.4	8.9	-6.5
Colomac						
C-1a	-16.9	8.5	-8.4	-12.0	11.6	-4.0
C-3c	-17.4	8.0	-8.8	-12.4	11.1	-4.5
Yellowknife						
Y-1a	-14.7	10.7	-6.1	-9.3	13.6	-1.6
Y-2a	-14.5	10.2	-6.1	-9.2	13.0	-1.8
Y-3a	-13.9	10.9	-5.5	-9.0	13.8	-1.3
Y-4a	-14.0	10.4	-5.8	-9.2	13.3	-1.6
Airport	-13.3	11.1	-5.1	-8.5	14.1	-0.9

Table B.2 Mean surface temperatures for the (a) freezing season (T_{sf} ; 1 September to 30 April), (b) thawing season (T_{st} ; 1 May to 31 August), and (c) annual period (T_{sa} ; 1 September to 31 August) at the instrumented sites.

	2004-05			2005-06		
	$T_{sf}(^{\circ}\text{C})$	$T_{st}(^{\circ}\text{C})$	$T_{sa}(^{\circ}\text{C})$	$T_{sf}(^{\circ}\text{C})$	$T_{st}(^{\circ}\text{C})$	$T_{sa}(^{\circ}\text{C})$
Ekati						
E-1a	-9.9	3.9	-5.4	-8.3	-	-
E-2a	-	-	-	-7.0	7.9	-2.0
E-2b	-8.5	2.5	-4.8	-8.0	7.3	-2.9
E-3a	-10.9	3.0	-6.2	-8.6	-	-
E-3b	-10.8	2.4	-6.4	-8.9	5.0	-4.2
E-4a	-8.9	3.2	-4.8	-7.4	-	-
E-4b	-8.9	1.3	-5.5	-7.4	4.7	-3.4
E-5a*	N.A.	N.A.	N.A.	-14.3	11.2	-5.6
E-5b*	N.A.	N.A.	N.A.	-12.2	10.7	-4.4
Colomac						
C-1a	-2.5	7.8	1.0	-1.4	-	-
C-1b	-1.6	-	-	-0.7	8.7	2.5
C-1c	-3.2	5.3	-0.3	-2.1	-	-
C-2a	-1.5	6.2	1.1	-0.6	7.7	2.2
C-2b	-	-	-	-1.1	-	-
C-3a	-4.4	6.4	-0.8	-	-	-
C-3b	-5.8	6.5	-1.6	-2.9	-	-
C-3c	-5.5	5.6	-1.7	-3.0	-	-
Yellowknife						
Y-1a	-2.9	8.0	0.8	-1.4	10.3	2.5
Y-2a	-1.0	8.8	2.3	-0.5	10.3	3.1
Y-2b	-1.0	7.6	1.9	-0.7	-	-
Y-2c	-3.2	6.4	0.1	-1.3	9.3	2.3
Y-3a	-3.3	9.3	1.0	-0.7	12.9	3.9
Y-3b	-4.1	-	-	-1.3	9.7	2.4
Y-4a	-2.1	9.6	1.8	-0.8	-	-
Y-4b	-3.2	7.5	0.4	-1.2	8.2	2.0
Y-5a*	N.A.	N.A.	N.A.	-2.2	10.0	2.3

* Mineral site with minimal organic layer

Table B.3 Maximum, minimum and annual (T_{50a} ; 1 September to 31 August) ground temperatures at 50 cm depth (T_{50}) at the instrumented sites.

	2004-05			2005-06		
	Maximum T_{50} (°C)	Minimum T_{50} (°C)	T_{50a} (°C)	Maximum T_{50} (°C)	Minimum T_{50} (°C)	T_{50a} (°C)
Ekati						
E-1a	-0.5	-15.4	-6.4	0.3	-14.1	-5.1
E-2a	-	-	-	2.9	-12.9	-4.0
E-2b	0.9	-15.4	-5.9	2.9	-14.1	-4.7
E-3a	-0.2	-16.0	-6.4	-	-14.3	-
E-3b	-0.2	-16.0	-6.7	-0.2	-14.7	-5.1
E-4a	-0.4	-14.1	-5.7	-0.2	-13.5	-4.5
E-4b	-0.6	-14.7	-5.9	-0.2	-12.9	-4.7
E-5a*	N.A.	N.A.	N.A.	18.1	-31.9	-5.5
E-5b*	N.A.	N.A.	N.A.	17.5	-24.0	-4.3
Colomac						
C-1a	0.7	-2.9	-0.8	2.0	-1.5	-0.3
C-1b	1.2	-2.4	-	2.9	-0.6	0.2
C-1c	-0.2	-4.3	-1.3	-	-2.9	-
C-2a	0.3	-1.5	-0.5	1.6	-0.6	-0.1
C-2b	-	-	-	4.2	-0.6	0.4
C-3a	-0.2	-7.3	-2.6	-	-	-
C-3b	-0.2	-7.8	-2.7	-	-5.3	-
C-3c	0.5	-10.0	-2.8	6.2	-7.3	-0.8
Yellowknife						
Y-1a	0.3	-6.3	-1.7	1.2	-4.3	-0.6
Y-2a	2.0	-1.5	-0.1	4.2	-0.2	0.5
Y-2b	2.9	0.3	0.8	5.9	-0.2	1.1
Y-2c	-0.2	-6.8	-1.7	1.2	-2.4	-0.4
Y-3a	1.2	-4.8	-1.0	3.3	-0.2	0.5
Y-3b	-0.2	-4.8	-1.4	-0.2	-1.5	-0.5
Y-4a	1.6	-3.4	-0.7	4.2	-0.6	0.4
Y-4b	0.7	-6.3	-1.6	2.5	-0.6	0.0
Y-5a*	N.A.	N.A.	N.A.	8.2	-2.9	1.4

* Mineral site with minimal organic layer

Table B.4 Maximum, minimum and annual (T_{100a} ; 1 September to 31 August) ground temperatures at 100 cm depth (T_{100}) at the instrumented sites.

	2004-05			2005-06		
	Maximum T_{100} (°C)	Minimum T_{100} (°C)	T_{100a} (°C)	Maximum T_{100} (°C)	Minimum T_{100} (°C)	T_{100a} (°C)
Ekati						
E-1a	-1.1	-13.5	-6.3	-1.1	-12.3	-5.1
E-2a	-	-	-	-0.2	-11.1	-4.2
E-2b	-0.6	-14.1	-5.9	-0.2	-12.9	-5.1
E-3a	-0.5	-14.1	-6.2	-	-12.3	-
E-3b	-1.1	-14.7	-6.6	-1.1	-12.9	-5.2
E-4a	-1.1	-12.9	-5.7	-1.1	-11.7	-4.6
E-4b	-1.1	-13.5	-6.0	-0.7	-11.1	-4.7
E-5a*	N.A.	N.A.	N.A.	13.2	-24.3	-5.0
E-5b*	N.A.	N.A.	N.A.	11.9	-20.2	-4.3
Colomac						
C-1a	-0.2	-1.5	-0.6	-0.2	-0.6	-0.4
C-1b	-0.6	-1.5	-	-0.6	-0.6	-0.6
C-1c	-0.6	-2.9	-1.2	-	-2.0	-
C-2a	-0.2	-0.6	-0.4	-0.6	-0.6	-0.6
C-2b	-	-	-	-0.2	-0.2	-0.2
C-3a	-0.2	-6.3	-2.5	-	-	-
C-3b	-0.2	-6.3	-2.5	-	-3.9	-
C-3c	-0.2	-6.8	-2.3	-0.2	-4.3	-1.3
Yellowknife						
Y-1a	-0.6	-5.2	-1.9	-0.6	-2.9	-1.0
Y-2a	-0.6	-1.1	-0.7	-0.6	-0.6	-0.6
Y-2b	-0.2	-0.6	-0.2	-0.2	-0.2	-0.2
Y-2c	-0.2	-4.8	-1.7	-0.6	-1.5	-0.7
Y-3a	-	-	-	-	-	-
Y-3b	-0.2	-3.4	-1.2	-0.2	-0.6	-0.3
Y-4a	-0.6	-2.0	-0.8	-0.6	-0.6	-0.6
Y-4b	-0.6	-4.8	-1.8	-1.1	-1.5	-1.1
Y-5a*	N.A.	N.A.	N.A.	4.6	-0.2	0.8

* Mineral site with minimal organic layer

Table B.5 Monthly mean air temperatures at the instrumented sites.

(a) 2004-05	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
Ekati												
E-1a	1.5	-9.8	-21.6	-31.7	-32.0	-29.4	-23.1	-10.7	-7.7	6.3	10.7	9.2
E-4a	1.9	-9.3	-21.1	-31.2	-31.4	-29.5	-23.5	-10.7	-7.6	6.2	10.7	9.4
Colomac												
C-1a	3.3	-7.1	-19.1	-30.5	-30.5	-26.6	-18.8	-6.1	0.0	9.8	13.1	10.9
C-3c	2.7	-7.3	-19.4	-31.4	-31.0	-26.7	-19.3	-6.5	-0.4	9.3	12.7	10.2
Yellowknife												
Y-1a	4.8	-4.8	-15.8	-29.0	-29.0	-24.6	-16.1	-3.1	2.4	12.5	15.4	12.5
Y-2a	4.1	-4.8	-15.6	-28.7	-28.8	-24.0	-15.2	-2.7	2.6	11.9	14.6	11.7
Y-3a	4.9	-4.2	-15.2	-27.9	-27.8	-23.6	-15.0	-2.4	3.1	12.6	15.3	12.5
Y-4a	4.4	-4.3	-15.4	-27.7	-27.4	-23.5	-15.4	-2.6	2.9	12.0	14.9	11.9
(b) 2005-06	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
Ekati												
E-1a	0.7	-6.2	-16.6	-20.1	-26.3	-23.7	-17.4	-12.1	1.1	11.4	11.7	11.2
E-4a	1.0	-5.6	-16.2	-20.0	-25.8	-23.3	-17.7	-12.0	1.0	11.1	11.9	11.4
Colomac												
C-1a	2.8	-2.5	-12.5	-16.5	-25.6	-21.0	-14.4	-6.7	5.7	14.7	14.0	12.1
C-3c	2.3	-3.0	-13.0	-16.8	-25.5	-21.4	-14.8	-7.0	5.2	14.1	13.4	11.8
Yellowknife												
Y-1a	4.9	-0.7	-9.7	-14.9	-22.5	-18.7	-10.9	-2.3	7.7	16.4	15.9	14.4
Y-2a	4.3	-0.9	-9.7	-14.6	-22.6	-18.6	-10.4	-1.7	7.5	15.9	15.2	13.5
Y-3a	5.0	-0.4	-9.3	-14.1	-22.2	-18.1	-10.7	-2.1	8.0	16.5	16.1	14.5
Y-4a	4.8	-0.6	-9.5	-14.0	-22.4	-18.3	-11.4	-2.5	7.7	15.8	15.8	14.0

Table B.6 Monthly mean surface temperatures at the instrumented sites.

(a) 2004-05	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
Ekati												
E-1a	1.9	-1.0	-3.8	-14.0	-17.3	-18.2	-16.4	-10.5	-6.6	5.2	9.3	7.8
E-2a	-	-	-	-	-	-	-	-	-	-	-	-
E-2b	1.6	-0.8	-2.1	-9.2	-14.8	-16.5	-15.5	-11.2	-7.5	3.2	7.6	6.9
E-3a	1.8	-2.0	-5.4	-15.7	-18.8	-19.6	-16.9	-10.5	-4.5	3.0	7.2	6.6
E-3b	1.6	-1.4	-5.6	-15.9	-18.7	-19.4	-17.3	-10.2	-5.0	2.0	6.5	6.1
E-4a	2.3	-1.3	-2.6	-11.0	-15.6	-17.0	-15.5	-10.7	-6.4	3.9	8.1	7.5
E-4b	1.6	-0.5	-2.0	-11.3	-16.1	-17.5	-15.3	-10.7	-6.1	1.5	5.1	5.0
E-5a*	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.
E-5b*	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.
Colomac												
C-1a	2.6	-0.5	-1.2	-2.8	-4.8	-5.7	-4.6	-2.8	0.9	7.6	12.3	10.3
C-1b	2.6	-0.5	-0.7	-1.5	-2.8	-3.8	-3.5	-2.5	-	-	-	-
C-1c	2.3	-0.9	-1.7	-4.7	-6.3	-6.6	-5.1	-2.8	0.0	4.8	8.6	7.9
C-2a	2.3	-0.4	-1.1	-3.1	-2.7	-2.8	-2.5	-1.6	0.3	6.1	9.9	8.6
C-2b	2.5	-1.1	-	-	-	-	-	-	-	-	-	-
C-3a	2.5	-0.9	-1.5	-5.9	-8.4	-9.1	-8.0	-4.3	0.3	6.4	10.1	8.8
C-3b	2.3	-1.6	-3.7	-9.6	-10.2	-10.3	-8.5	-4.5	0.2	6.5	10.5	8.6
C-3c	1.8	-0.8	-2.3	-6.6	-10.3	-11.6	-9.4	-4.8	-1.1	5.7	9.4	8.3
Yellowknife												
Y-1a	3.5	-0.4	-1.0	-2.2	-4.9	-8.2	-7.1	-3.2	1.2	8.5	12.3	10.0
Y-2a	3.2	-0.4	-0.9	-2.0	-2.3	-2.3	-2.5	-0.9	4.3	9.8	11.6	9.4
Y-2b	3.0	-0.5	-1.3	-2.3	-2.5	-2.4	-1.7	-0.3	1.1	7.6	11.7	10.0
Y-2c	3.0	-0.6	-1.1	-2.3	-7.2	-8.6	-6.4	-2.4	0.6	6.4	10.1	8.7
Y-3a	3.6	-1.0	-2.6	-6.1	-5.8	-6.6	-5.9	-2.0	2.7	10.2	13.6	10.8
Y-3b	3.2	-1.4	-3.4	-6.8	-8.1	-8.1	-6.1	-1.8	-	-	-	-
Y-4a	3.4	-0.6	-1.3	-3.0	-4.7	-5.3	-4.3	-1.4	3.0	10.7	13.7	10.9
Y-4b	3.5	-0.7	-1.7	-2.8	-6.5	-8.6	-6.6	-2.4	1.3	7.4	11.1	10.0
Y-5a*	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.

* Mineral site with minimal organic layer

Table B.6 continued

(b) 2005-06	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
Ekati												
E-1a	1.8	-1.0	-3.7	-10.2	-13.2	-15.4	-14.5	-10.8	-	-	-	-
E-2a	2.4	-0.2	-2.8	-8.2	-11.1	-13.3	-13.1	-10.1	0.6	9.7	10.8	10.7
E-2b	1.6	-0.9	-3.8	-9.5	-12.7	-14.5	-13.9	-11.0	-0.9	8.8	10.7	10.6
E-3a	1.6	-1.4	-5.0	-10.2	-12.4	-15.3	-15.1	-11.6	-1.2	-	-	-
E-3b	1.7	-1.1	-6.2	-11.1	-12.6	-15.6	-15.2	-11.5	-1.2	5.1	8.0	8.3
E-4a	2.2	-0.8	-2.8	-7.7	-11.3	-14.4	-14.1	-10.6	-	-	-	-
E-4b	1.4	-0.4	-2.5	-8.1	-11.6	-13.9	-13.5	-11.0	-1.6	4.8	7.2	8.3
E-5a*	1.2	-5.4	-16.2	-18.8	-21.4	-23.4	-18.0	-10.9	3.7	14.1	14.3	12.9
E-5b*	1.0	-3.5	-11.8	-16.1	-16.4	-20.4	-17.8	-11.1	2.8	13.6	14.0	12.7
Colomac												
C-1a	3.3	-0.9	-1.6	-2.1	-3.0	-2.7	-2.9	-1.6	-	-	-	-
C-1b	3.0	-0.5	-0.7	-0.8	-1.8	-1.8	-1.8	-0.9	1.8	10.3	12.1	10.5
C-1c	2.7	-0.8	-1.4	-1.6	-4.9	-4.5	-4.0	-2.4	-	-	-	-
C-2a	2.8	-0.4	-0.7	-1.0	-2.0	-1.7	-1.5	-0.7	1.7	8.7	10.6	9.7
C-2b	3.0	-1.0	-1.3	-1.6	-2.8	-2.2	-1.9	-1.1	-	-	-	-
C-3a	-	-	-	-	-	-	-	-	-	-	-	-
C-3b	2.6	-1.0	-1.4	-1.9	-4.1	-7.1	-6.8	-3.8	3.0	-	-	-
C-3c	2.7	-0.6	-1.2	-1.7	-4.3	-7.0	-7.7	-4.1	-	-	-	-
Yellowknife												
Y-1a	4.1	-0.4	-0.6	-0.7	-1.7	-3.7	-5.6	-2.5	3.9	12.1	13.3	12.0
Y-2a	4.1	-0.6	-0.6	-0.7	-1.4	-2.3	-1.8	-0.8	4.4	11.5	13.0	12.3
Y-2b	4.2	-0.4	-0.8	-1.1	-2.2	-3.2	-2.1	-0.5	-	-	-	-
Y-2c	3.5	-0.6	-0.8	-1.1	-2.3	-3.6	-3.7	-2.1	3.7	10.7	11.8	11.2
Y-3a	4.1	-0.8	-1.1	-1.5	-1.9	-1.9	-1.7	-1.0	6.3	15.4	16.3	13.8
Y-3b	3.6	-0.5	-1.7	-2.1	-3.5	-2.8	-2.3	-1.2	4.1	11.3	12.1	11.2
Y-4a	4.3	-0.6	-1.0	-1.3	-2.8	-2.6	-1.8	-0.7	-	-	-	-
Y-4b	4.4	-0.3	-1.3	-1.9	-3.7	-3.8	-2.2	-0.9	3.2	9.3	10.6	9.7
Y-5a*	2.8	-0.2	-1.0	-2.8	-4.9	-4.9	-3.6	-1.3	4.8	11.4	12.3	11.8

* Mineral site with minimal organic layer

Table B.7 Surface defined values of freezing air (T_{Af}) and surface (T_{Sf}) temperature and freezing surface offsets (S.O._f).

	2004-05			2005-06		
	T_{Af} (°C)	T_{Sf} (°C)	S.O. _f (°C)	T_{Af} (°C)	T_{Sf} (°C)	S.O. _f (°C)
Ekati						
E-1a	-20.8	-11.0	9.8	-15.9	-9.0	6.9
E-2a	-	-	-	-16.5	-7.9	8.5
E-2b	-20.4	-9.5	10.9	-15.8	-8.7	7.1
E-3a	-20.3	-11.5	8.8	-14.7	-8.8	6.0
E-3b	-19.8	-11.3	8.5	-14.7	-9.0	5.7
E-4a	-20.5	-10.0	10.5	-15.8	-8.1	7.7
E-4b	-20.0	-9.7	10.3	-14.8	-7.6	7.2
E-5a*	N.A.	N.A.	N.A.	-16.4	-15.2	1.2
E-5b*	N.A.	N.A.	N.A.	-16.3	-12.9	3.4
Colomac						
C-1a	-18.8	-3.0	15.8	-13.3	-2.0	11.3
C-1b	-18.8	-2.1	16.8	-13.0	-1.1	11.9
C-1c	-17.8	-3.6	14.2	-13.7	-2.7	11.0
C-2a	-17.8	-1.8	16.0	-13.0	-1.1	11.9
C-2b	-	-	-	-13.3	-1.6	11.6
C-3a	-19.1	-5.1	14.0	-	-	-
C-3b	-18.5	-6.3	12.3	-13.4	-3.5	9.9
C-3c	-18.0	-5.9	12.1	-13.6	-3.6	10.0
Yellowknife						
Y-1a	-16.8	-3.7	13.1	-11.2	-2.1	9.1
Y-2a	-17.2	-1.6	15.6	-11.2	-1.2	10.1
Y-2b	-17.7	-1.6	16.0	-11.4	-1.5	9.9
Y-2c	-15.7	-3.7	12.0	-11.3	-2.0	9.2
Y-3a	-16.0	-4.2	11.9	-10.8	-1.4	9.4
Y-3b	-17.1	-5.3	11.8	-11.0	-2.0	9.0
Y-4a	-16.4	-2.9	13.5	-11.1	-1.5	9.6
Y-4b	-14.9	-3.7	11.2	-11.4	-2.0	9.4
Y-5a*	N.A.	N.A.	N.A.	-11.2	-2.6	8.6

* Mineral site with minimal organic layer

Table B.8 Surface defined values of thawing air (T_{At}) and surface (T_{St}) temperature and thawing surface offsets ($S.O._t$).

	2004-05			2005-06		
	T_{At} (°C)	T_{St} (°C)	$S.O._t$ (°C)	T_{At} (°C)	T_{St} (°C)	$S.O._t$ (°C)
Ekati						
E-1a	8.3	7.1	-1.2	-	-	-
E-2a	-	-	-	8.8	8.1	-0.8
E-2b	8.1	5.6	-2.5	9.1	7.9	-1.2
E-3a	8.1	5.3	-2.9	-	-	-
E-3b	8.3	4.9	-3.5	9.8	6.0	-3.8
E-4a	7.9	6.0	-1.9	-	-	-
E-4b	8.4	3.9	-4.5	9.8	5.8	-4.0
E-5a*	N.A.	N.A.	N.A.	8.7	10.9	2.3
E-5b*	N.A.	N.A.	N.A.	8.8	10.8	2.0
Colomac						
C-1a	8.9	8.0	-0.9	-	-	-
C-1b	-	-	-	10.8	8.3	-2.4
C-1c	10.1	6.3	-3.8	-	-	-
C-2a	10.0	7.1	-2.9	10.8	7.4	-3.3
C-2b	-	-	-	10.6	-	-
C-3a	8.5	6.7	-1.8	-	-	-
C-3b	9.2	7.2	-2.0	10.3	-	-
C-3c	9.6	7.0	-2.6	10.2	-	-
Yellowknife						
Y-1a	10.2	7.8	-2.5	12.2	9.2	-3.0
Y-2a	9.0	7.9	-1.1	11.5	9.1	-2.4
Y-2b	8.7	6.7	-2.0	11.8	-	-
Y-2c	10.6	6.9	-3.7	11.5	8.3	-3.2
Y-3a	10.4	8.9	-1.5	12.4	11.3	-1.1
Y-3b	-	-	-	12.2	8.6	-3.7
Y-4a	9.8	8.9	-0.8	-	-	-
Y-4b	10.9	8.1	-2.9	11.8	7.4	-4.4
Y-5a*	N.A.	N.A.	N.A.	12.1	9.1	-2.9

* Mineral site with minimal organic layer

Table B.9 Surface defined values of annual air (T_{Aa}) and surface (T_{Sa}) temperature and annual surface offsets ($S.O._a$).

	2004-05			2005-06		
	T_{Aa} (°C)	T_{Sa} (°C)	$S.O._a$ (°C)	T_{Aa} (°C)	T_{Sa} (°C)	$S.O._a$ (°C)
Ekati						
E-1a	-11.7	-5.4	6.4	-	-	-
E-2a	-	-	-	-6.8	-1.8	5.0
E-2b	-11.7	-4.9	6.8	-6.7	-2.6	4.1
E-3a	-11.6	-6.3	5.2	-	-	-
E-3b	-11.6	-6.6	5.0	-6.6	-4.0	2.6
E-4a	-11.5	-4.9	6.6	-	-	-
E-4b	-11.6	-5.7	5.9	-6.5	-3.1	3.4
E-5a*	N.A.	N.A.	N.A.	-6.7	-5.1	1.6
E-5b*	N.A.	N.A.	N.A.	-6.7	-3.8	2.9
Colomac						
C-1a	-8.5	1.1	9.6	-	-	-
C-1b	-	-	-	-3.7	2.6	6.3
C-1c	-8.5	-0.3	8.2	-	-	-
C-2a	-8.5	1.2	9.7	-3.7	2.3	5.9
C-2b	-	-	-	-	-	-
C-3a	-9.0	-0.8	8.2	-	-	-
C-3b	-9.0	-1.6	7.3	-	-	-
C-3c	-9.0	-1.7	7.3	-	-	-
Yellowknife						
Y-1a	-6.2	0.8	7.0	-1.2	2.7	3.9
Y-2a	-6.2	2.4	8.5	-1.4	3.3	4.7
Y-2b	-6.2	2.0	8.2	-1.4	-	-
Y-2c	-6.2	0.1	6.3	-1.4	2.4	3.8
Y-3a	-5.6	1.0	6.6	-0.9	4.0	4.9
Y-3b	-	-	-	-0.9	2.6	3.5
Y-4a	-5.8	1.9	7.7	-	-	-
Y-4b	-5.7	0.5	6.2	-1.3	2.1	3.3
Y-5a*	N.A.	N.A.	N.A.	-1.2	2.4	3.6

* Mineral site with minimal organic layer

Table B.10 Surface defined values of freezing degree-days for the air (DDF_A) and surface (DDF_S) and freezing n-factors (n_f).

	2004-05			2005-06		
	DDF _A (°C d)	DDF _S (°C d)	n _f	DDF _A (°C d)	DDF _S (°C d)	n _f
Ekati						
E-1a	5039	2656	0.53	3771	2126	0.56
E-2a	-	-	-	3761	1813	0.48
E-2b	5040	2349	0.47	3771	2062	0.55
E-3a	4989	2820	0.57	3722	2190	0.59
E-3b	4989	2827	0.57	3722	2257	0.61
E-4a	4989	2419	0.48	3720	1901	0.51
E-4b	4987	2403	0.48	3722	1900	0.51
E-5a*	N.A.	N.A.	N.A.	3771	3487	0.92
E-5b*	N.A.	N.A.	N.A.	3770	2977	0.79
Colomac						
C-1a	4251	676	0.16	3049	455	0.15
C-1b	4251	465	0.11	3048	253	0.08
C-1c	4275	857	0.20	3039	595	0.20
C-2a	4275	436	0.10	3048	244	0.08
C-2b	-	-	-	3049	369	0.12
C-3a	4344	1156	0.27	-	-	-
C-3b	4366	1471	0.34	3119	793	0.25
C-3c	4369	1417	0.32	3119	809	0.26
Yellowknife						
Y-1a	3729	812	0.22	2474	462	0.19
Y-2a	3639	340	0.09	2440	246	0.10
Y-2b	3624	335	0.09	2444	315	0.13
Y-2c	3671	857	0.23	2444	428	0.18
Y-3a	3546	910	0.26	2392	304	0.13
Y-3b	3517	1078	0.31	2392	432	0.18
Y-4a	3554	620	0.17	2444	332	0.14
Y-4b	3569	877	0.25	2441	328	0.13
Y-5a*	N.A.	N.A.	N.A.	2474	565	0.23

* Mineral site with minimal organic layer

Table B. 11 Surface defined values of thawing degree-days for the air (DDT_A) and surface (DDT_S) and thawing n-factors (n_t).

	2004-05			2005-06		
	DDT _A (°C d)	DDT _S (°C d)	n _t	DDT _A (°C d)	DDT _S (°C d)	n _t
Ekati						
E-1a	891	756	0.85	-	-	-
E-2a	-	-	-	1255	1147	0.91
E-2b	880	606	0.69	1245	1077	0.87
E-3a	891	575	0.65	-	-	-
E-3b	873	505	0.58	1226	755	0.62
E-4a	894	670	0.75	-	-	-
E-4b	876	402	0.46	1230	729	0.59
E-5a*	N.A.	N.A.	N.A.	1258	1586	1.26
E-5b*	N.A.	N.A.	N.A.	1257	1516	1.21
Colomac						
C-1a	1212	1066	0.88	-	-	-
C-1b	-	-	-	1591	1231	0.77
C-1c	1205	750	0.62	-	-	-
C-2a	1205	856	0.71	1591	1100	0.69
C-2b	-	-	-	-	-	-
C-3a	1147	885	0.77	-	-	-
C-3b	1145	889	0.78	-	-	-
C-3c	1125	814	0.72	-	-	-
Yellowknife						
Y-1a	1485	1112	0.75	1951	1468	0.75
Y-2a	1416	1204	0.85	1860	1475	0.79
Y-2b	1426	1065	0.75	-	-	-
Y-2c	1394	899	0.64	1868	1347	0.72
Y-3a	1508	1277	0.85	1983	1814	0.91
Y-3b	-	-	-	1996	1395	0.70
Y-4a	1450	1313	0.91	-	-	-
Y-4b	1415	1044	0.74	1919	1201	0.63
Y-5a*	N.A.	N.A.	N.A.	1951	1498	0.77

* Mineral site with minimal organic layer

Table B.12 Surface defined annual mean temperatures at the surface (T_{sa}) and 100 cm depth(T_{100a}), and the thermal offset (T.O.).

	2004-05			2005-06		
	T_{sa} (°C)	T_{100a} (°C)	T.O. (°C)	T_{sa} (°C)	T_{100a} (°C)	T.O. (°C)
Ekati						
E-1a	-5.4	-6.4	-1.0	-	-5.2	-
E-2a	-	-	-	-1.8	-4.4	-2.6
E-2b	-4.9	-6.0	-1.1	-2.6	-5.2	-2.6
E-3a	-6.3	-6.3	0.0	-	-	-
E-3b	-6.6	-6.7	-0.2	-4.0	-5.3	-1.3
E-4a	-4.9	-5.8	-0.9	-	-4.7	-
E-4b	-5.7	-6.1	-0.4	-3.1	-4.8	-1.7
E-5a*	N.A.	N.A.	N.A.	-5.1	-5.2	N.A.
E-5b*	N.A.	N.A.	N.A.	-3.8	-4.5	N.A.
Colomac						
C-1a	1.1	-0.6	-1.7	-	-0.4	-
C-1b	-	-	-	2.6	-0.6	-3.2
C-1c	-0.3	-1.2	-0.9	-	-	-
C-2a	1.2	-0.4	-1.6	2.3	-0.6	-2.9
C-2b	-	-	-	-	-0.2	-
C-3a	-0.8	-2.7	-1.9	-	-	-
C-3b	-1.6	-2.6	-0.9	-	-	-
C-3c	-1.7	-2.4	-0.7	-	-1.2	-
Yellowknife						
Y-1a	0.8	-1.9	-2.7	2.7	-1.0	-3.7
Y-2a	2.4	-0.7	-3.1	3.3	-0.6	-3.9
Y-2b	2.0	-0.2	-2.2	-	-0.2	-
Y-2c	0.1	-1.7	-1.8	2.4	-0.7	-3.1
Y-3a	1.0	-	-	4.0	-	-
Y-3b	-	-1.2	-	2.6	-0.3	-2.8
Y-4a	1.9	-0.8	-2.7	-	-	-
Y-4b	0.5	-1.8	-2.3	2.1	-1.1	-3.2
Y-5a*	N.A.	N.A.	N.A.	2.4	1.1	N.A.

* Mineral site with minimal organic layer

Table B.13 Surface defined thermal offset (T.O.), thawing degree-days for the surface (DDT_s), annual period (P), and thermal conductivity ratio determined from T.O., DDT_s and P.

2004-05					2005-06			
	T.O. ^a (°C)	DDT _s (°C d)	P (d)	rk _{Temperature}	T.O. ^a (°C)	DDT _s (°C d)	P (d)	rk _{Temperature}
Ekati								
E-1a	-1.0	756	354	0.52	-	-	-	-
E-2a	-	-	-	-	-2.6	1147	370	0.17
E-2b	-1.1	606	354	0.35	-2.6	1077	375	0.10
E-3a	0.0	575	354	1.02	-	-	-	-
E-3b	-0.2	505	354	0.88	-1.3	755	375	0.36
E-4a	-0.9	670	355	0.55	-	-	-	-
E-4b	-0.4	402	352	0.63	-1.7	729	376	0.13
E-5a*	N.A.	N.A.	N.A.	N.A.	N.A.	1586	376	N.A.
E-5b*	N.A.	N.A.	N.A.	N.A.	N.A.	1516	374	N.A.
Colomac								
C-1a	-1.7	1066	358	0.42	-	-	-	-
C-1b	-	-	-	-	-3.2	1231	377	0.02
C-1c	-0.9	750	358	0.57	-	-	-	-
C-2a	-1.6	856	359	0.32	-2.9	1100	377	0.01
C-2b	-	-	-	-	-	-	-	-
C-3a	-1.9	885	359	0.22	-	-	-	-
C-3b	-0.9	889	359	0.62	-	-	-	-
C-3c	-0.7	814	358	0.70	-	-	-	-
Yellowknife								
Y-1a	-2.7	1112	363	0.12	-3.7	1468	375	0.05
Y-2a	-3.1	1204	363	0.08	-3.9	1475	374	0.01
Y-2b	-2.2	1065	362	0.26	-	-	-	-
Y-2c	-1.8	899	362	0.27	-3.1	1347	375	0.13
Y-3a	-	-	-	-	-	-	-	-
Y-3b	-	-	-	-	-2.8	1395	376	0.23
Y-4a	-2.7	1313	362	0.24	-	-	-	-
Y-4b	-2.3	1044	362	0.20	-3.2	1201	373	0.00
Y-5a*	N..	N.A.	N.A.	N.A.	N.A.	1498	377	N.A.

* Mineral site with minimal organic layer