# CLIMATOLOGY AND HISTORICAL SNOWCOVER OF THE BIG LEVEL PLATEAU, GROS MORNE NATIONAL PARK, NEWFOUNDLAND



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# **Climatology and Historical Snowcover of the Big Level Plateau, Gros Morne National Park, Newfoundland**

by

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## Abstract

The daily climatological record including snowcover water equivalent (SWE) was estimated for the Big Level Summit Plateau (BL), Gros Morne National Park, western Newfoundland, for the 1962-1999 study period. This was accomplished using widely known statistical techniques, i.e multiple linear regression and mean ratios, linking the Big Level Autostation (1993-1999) record with the nearby low lying coastal stations records through the climate elements mentioned above. Physically based relationships were added for the snow model, which was based on the reconstructed climatological elements (temperature, wind speed and precipitation).

The BL daily temperature record (estimated for 1946-1999) and hourly wind speed record were estimated with multiple linear regression. Precipitation was estimated using mean ratios. The estimated BL winter precipitation record was separated into snowfall and rainfall and measured snowfall at BL was corrected for wind induced undercatch. Snowcover SWE was estimated with the estimated daily precipitation record and the temperature index method, using increasing degree-day factors as the snow season progressed. Snowdrifting off the plateau and sublimation of blowing snow were also estimated using simple expressions that were a function of wind speed.

Snowcover on BL is consistently present for long periods from year to year, although its thickness varies considerably compared with the coastal plain (BL maximum seasonal SWE varied from 110 mm to 1650 mm). The single largest factor influencing the snowcover SWE,

other than the large amounts of snowfall, was found to be blowing snow sublimation, which ablates approximately 53% of the snowfall accumulation on average. Another large factor influencing snowcover SWE is snowdrifting off the plateau, accounting for approximately 10% of the average accumulated snowfall, leaving approximately 32% to be ablated by melt. The months from February to June all showed negative significant trends with respect to mean BL SWE during the study period. There were no significant linear trends in the temperature record for the period of estimation, but there was a significant linear positive trend (95% level) for the number of thawing degree-days using daily maximum temperature for the snow accumulation season (November-April) of 1.3-1.8 degree-days per year.

An anthropogenically induced warmer climate would lead to much thinner snowcovers existing during shorter periods, with snow seasons having reduced snowfall and increased rainfall values. The thinnest seasonal snowcovers of the recent past such as in the 1981 snow season could become above average if the IPCC climate change projections become reality.

## Acknowledgments

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## Table of Contents

## List of Tables

## List of Figures

## List of Abbreviations and Symbols

Abbreviations

| | |
|---|---|
| AES | Atmospheric Environment Service (now MSC) |
| BL | Big Level Summit Plateau |
| BLD | Big Level Autostation daily values |
| BLS | Big Level Autostation |
| CH | Cow Head |
| DH | Daniel's Harbour |
| GCM | Global circulation model |
| IPCC | Intergovernmental Panel on Climate Change |
| JD | Numerical day of the year, i.e. Julian day |
| LR | Lapse rate calculated from SUA |
| NAO | North Atlantic Oscillation |
| NST | Newfoundland Standard Time |
| PBSM | Prairie Blowing Snow Model |
| RH | Rocky Harbour |
| rh | Relative humidity (%) |
| SE | Standard Error |
| SOG | Snow on the ground (depth - cm) |
| SUA | Stephenville A Upper Air Station |
| SWE | Snow water equivalent (mm) |
| WE | Water equivalent (mm) |
| WMO | World Meteorological Organisation |
| Z | Zulu - Greenwich Mean Time |

Symbols

| | |
|---|---|
| a | Degree-day factor (mm $°C^{-1}$ $d^{-1}$) |
| A | Snow available for transport (mm SWE) |
| c | Calibration factor (dimentionless) |
| CC | Cold Content (kJ) |
| $C_p$ | Specific heat (kJ $kg^{-1}$ $°C^{-1}$) |
| CR | Catch ratio (%) |
| d | Day |
| $D_e$ | Bulk transfer coefficient for latent heat transfer (MJ $m^{-3}$ $mb^{-1}$) |
| Dr | Drifting loss (mm SWE) |
| $D_u$, d | Durbin-Watson test coefficients |
| $e_a$ | Vapour pressure of the air at reference height (mb) |
| $e_s$ | Vapour pressure of the snow surface (mb) |
| f,h,k | Coefficients |
| M | Melt (in SWE, mm $d^{-1}$) |
| MF | Degree-day factor (mm $°C^{-1}$ $d^{-1}$) |
| MP | Snowmelt potential (mm $d^{-1}$) |
| m & b | Slope and intercept value of a linear equation, respectively |
| NCC | New snow cold content (kJ) |
| q | Total Snow transported per metre perpendicular to the wind (kg $m^{-1}$ $s^{-1}$ or g $m^{-1}$ $s^{-1}$) |
| $q_{evap}$ | Snow sublimation rate (mg $m^{-2}$ $s^{-1}$) |
| Q* | Net radiation (MJ $m^{-2}$ $d^{-1}$) |
| $Q_e$ | Latent heat (MJ $m^{-2}$ $d^{-1}$) |
| $Q_g$ | Ground heat (MJ $m^{-2}$ $d^{-1}$) |
| $Q_h$ | Sensible heat (MJ $m^{-2}$ $d^{-1}$) |
| $Q_m$ | Energy available for melt (MJ $m^{-2}$ $d^{-1}$) |
| r | Correlation coefficient |

| | |
|---|---|
| R | Rainfall (mm) |
| S | Snowfall (mm WE) |
| Sub | SWE loss to blowing snow sublimation (mm) |
| T | Air temperature (°C) |
| $T_a$ | Daily mean temperature (°C) |
| $T_b$ | Base temperature for snowmelt (°C) |
| $T_i$, $T_f$ | Initial and final temperature of the snowpack |
| $T_{max}$ | Daily maximum temperature (°C) |
| $T_{min}$ | Daily minimum temperature (°C) |
| $T_{mean}$ | Daily mean temperature (°C) |
| $T_{osnow}$ | Temperature of old snow (°C) |
| u | Wind speed (km $h^{-1}$ or m $s^{-1}$) |
| U | Internal energy (MJ $m^{-2}$ or kJ $m^{-2}$) |
| $\Delta U/\Delta t$ | Internal energy change per unit time (MJ $m^{-2}$ $d^{-1}$) |
| u* | Threshold wind speed for snow transport (m $s^{-1}$) |
| $u_{10}$ | Hourly mean wind speed at 10 m height (m $s^{-1}$) |
| $u_p$ | Mean precipitation event wind speed (m $s^{-1}$) |
| WC | Water content of the snow (mm) |
| WR | Liquid water requirement for melt to begin (mm) |
| z | Height (m) |
| $\rho$ | Snowcover density (kg $m^{-3}$) |
| $\mu$ | Mass of the snowpack (kg $m^{-2}$) |

## List of Appendices

# 1 INTRODUCTION

In Canada, snowcover is an integral part of the landscape. In most areas it is present at least one third of the year during the winter months (Potter, 1965; McKay and Gray, 1981) affecting hydrology, climate, ecosystems and human activities. Studies that are concerned with maintaining the environmental integrity of our National Parks and wilderness areas in general (e.g. Anions and Berger, 1998), can thus not ignore snow as one of the essential components of the biophysical environment.

Perhaps one of the most fundamental impacts of snow on the biosphere is its effect on vegetation. The presence of snowcover greatly influences the length of the growing season. While snowcover usually restricts vegetation growth, its insulation properties protect plants from the potentially damaging effects of winter frost, thus allowing them to be more productive during the summer. In exposed environments, such as alpine tundra, abrasion of vegetation by drifting snow is a factor in preventing vegetation growth above the winter snowpack. This favors low-lying vegetation that could otherwise be threatened by more aggressive species, such as trees, in areas where the growing season is adequate for these species (Adams, 1981). Furthermore, late-lying snowbeds can cause the formation of unique vegetation patterns (Billings and Bliss, 1959).

Fauna is as greatly influenced by snow as vegetation. A snowpack as thin as 15 cm can be enough for small animals to build burrows, a relatively warm and safe place to live in winter. However, a deep snowpack can restrict the travels of animals, as well as making snowcovered food difficult to access. Moreover, the whitening of the environment can either expose animals to predation or hide them (Adams, 1981). Finally, late-lying snowbeds can offer fauna a cool environment to rest away from predatory insects (Brouillet *et al.*, 1998).

Snow is also a valuable water resource for all living things. The release of water during spring (or summer) snowmelt is the most important hydrological event of the year in this country. It is the cause of many floods and enhanced erosion events (McKay, 1981). In addition snowmelt at higher altitudes (or latitudes) can be an essential source of river flow later in the spring or summer for warmer and drier areas downstream (Steppuhn, 1981; Carr, 1989; Trabant and Clagett, 1990; Barry, 1994).

Due to its dominance during a large portion of the year, one can envisage that the importance of snowcover is augmented in mountain environments. In fact, higher altitude sites in general and alpine areas in particular, as defined as areas above the tree line (Löve, 1970), are particularly sensitive to changes or trends in their climate (Beniston, 1994a). Mountains are much more than just water reservoirs; they offer substantial biodiversity that is only equaled over hundreds or thousands of kilometres latitudinally. Climate warming could lead to an upward migration of species, profoundly affecting the ecological balance of these areas achieved over many centuries (Beniston, 1994a). However, relatively little is known about the climate of mountains because of the lack of observations both in space and in time caused mainly by the difficulty in accessing these sites. Consequently, this has led to a fewer number of studies dealing with climate change in these environments (Barry, 1994).

The above facts should provoke great concern in light of the Intergovernmental Panel on Climate Change referring to global change as being the greatest challenge currently facing humanity (IPCC, 1996). Global circulation models (GCMs) currently project a global mean temperature increase of 1.5 - 4.5°C by the year 2100 due to enhanced greenhouse gas concentrations in the atmosphere (IPCC, 2001a). This amount of warming in such a short period would be unprecedented. The impacts of such climate change, if one considers that there has only been a 4 to 5°C rise in global mean temperature since the last glacial maximum of 18,000 BP, could be tremendous (Grassl, 1994). Changes in hydroclimatic elements will be associated with the shifts in temperature. However, due to the complexity and smaller scale character of the moisture balance, the estimated changes in precipitation, evaporation, soil moisture and snowcover are highly uncertain (Grenell and Oerlemans, 1989; Barry, 1992; Beniston, 1994a). With respect to mountainous areas, the current resolution of GCMs is too coarse to adequately project climate change. Since the degree of parallelism and relative amplitude of climate change trends of upland versus lowland areas is not well understood, we can only make educated guesses about the changes in temperature, precipitation, winds, snowcover, and other parameters that could occur at higher elevations (Barry, 1992; Beniston, 1994a).

Despite the difficulties in determining alpine climate change, many authors have used analogies between past and potential future changes as a method of anticipating climate change (e.g Wigley *et al.*, 1980; Williams, 1980; Pittock and Salinger, 1982; Kellogg, 1982; Jaeger and Kellogg, 1983; Lough *et al.*, 1984). While the general focus of investigators is on future climate change, one must not forget that this phenomenon is continuously ongoing. The empirical approach assesses possible climate change in part by analysing extreme seasonal events in the historical record (Beniston, 1994b). Climate change in the 20$^{th}$ century has caused an increase in mean global temperatures of approximately 0.6°C (IPCC, 2001a). This warming most likely has both a natural and an anthropogenic signal, but it is not known exactly how much of it is the result of the conclusion of the Little Ice Age and of other natural factors or of anthropogenic greenhouse gases (Folland *et al.*, 1990; Schlesinger, 1991). However, the IPCC (2001a) has stated a likelihood that most of the warming of the past 50 years is attributable to human activities.

In the absence of lengthy records, reconstructed databases can be created by assuming similar statistical relationships for meteorological variables between present and past climate at lowland and upland sites and applying these relationships to historical meteorological values at lowland stations. This method is analogous to assessing future local conditions for mountainous areas by coupling the present statistical relationships to GCM outputs for the lowlands (Giorgi and Mearns, 1991). If this latter approach is assumed to be of value, than using the same statistical relationships would certainly be useful in assessing past climate variability and change prior to meteorological measurements in an alpine area and give some indication as what to expect in the future.

In attempting to evaluate climate change, snowcover as a meteorological variable is widely presented as a useful indicator because it inherently incorporates several climatic variables, including temperature, precipitation and winds (Carr, 1989; Brown and Braaten, 1998). When comparing sites of varying elevations, these factors result in snowcover conditions (depth and duration) at upland locations often being poorly correlated with those at low altitude (Pfister, 1985). However, by using the historical record of individual meteorological variables that determine the state of the local snowpack (i.e. temperature, precipitation, winds etc.), it is possible to reconstruct a more realistic database of snowcover conditions that can serve to indicate past climate variability and change at a given upland location.

## 1.1 Objectives

This research aims to estimate continuous snowcover duration and snow water equivalent (SWE - a function of depth and density) on the Big Level Summit Plateau, Gros Morne National Park, Newfoundland (Canada) from 1962-1999. This was accomplished by utilizing data from a climate autostation in operation on this alpine plateau in question since 1993, plus neighbouring coastal station records. Statistical relationships, including linear regressions, between the Big Level recent record and longer running climate stations with varying record lengths on the adjacent coastal plain were used to estimate the needed climate variables required for determining the SWE of the snowcover. Other specific objectives therefore include:

1) Estimating daily maximum, mean and minimum temperatures on the Big Level

Plateau from 1946-1999,

2) Estimating mean hourly wind speed on Big Level from 1962-1999,

3) Estimating daily snowfall, rainfall and total precipitation on Big Level from 1962-1999.

These reconstructed data records were compared to the coastal plain record and analysed for trends on a monthly or seasonal basis to assess climate variability and to extract any evidence of climate change. The results served to build a daily SWE record by approximating its accumulation, redistribution and ablation on the alpine plateau. The snow depth and derived snow cover duration record are compared on an interannual basis with the record from the coastal plain and examined for trends. Finally, the seasonal variability and sensitivity of the snowcover to future changes in climate is discussed.

# 2 PRESENTATION OF THE REGION AND CLIMATE STATIONS

## 2.1 The Study Region

Gros Morne National Park (GMNP) (49° N, 58°W) is located on the west coast of the island of Newfoundland, on the eastern shore of the Gulf of St. Lawrence (Figure 2.1). Its 55 km of coastline (linear approximation) is oriented SSW-NNE and is consistent with the orientation of the Great Northern Peninsula as a whole. With an area of 1805 $km^2$, Gros Morne is Atlantic Canada's largest national park. The territory can be subdivided into two main physiographic entities, the narrow coastal plain and the Long Range Mountains occupying most of the interior, both generally oriented SSW-NNE.

The coastal plain averages approximately 6-7 km in width, from the coast to the base of the Long Range Mountains. However, it is wider in the northern part of the Park and virtually disappears at its southern edge. In the transition zone between the lowland and the mountains, and dissecting the latter, lie numerous small lakes and ponds and four larger ones, namely Western Brook Pond, Bakers Brook Pond, Ten Mile Pond and Trout River Pond, which are all former fjords opening to the west. The vegetation cover of the park's lowlands is well mixed between sphagnum bogs and black spruce forests on boggy ground, and balsam fir forests on better drained slopes, each typical of the boreal forest (Berger *et al.*, 1992). Most local human activity occurs in this area of the Park.

The Long Range Mountains start their sharp ascent near 150 m a.s.l. and are, for the

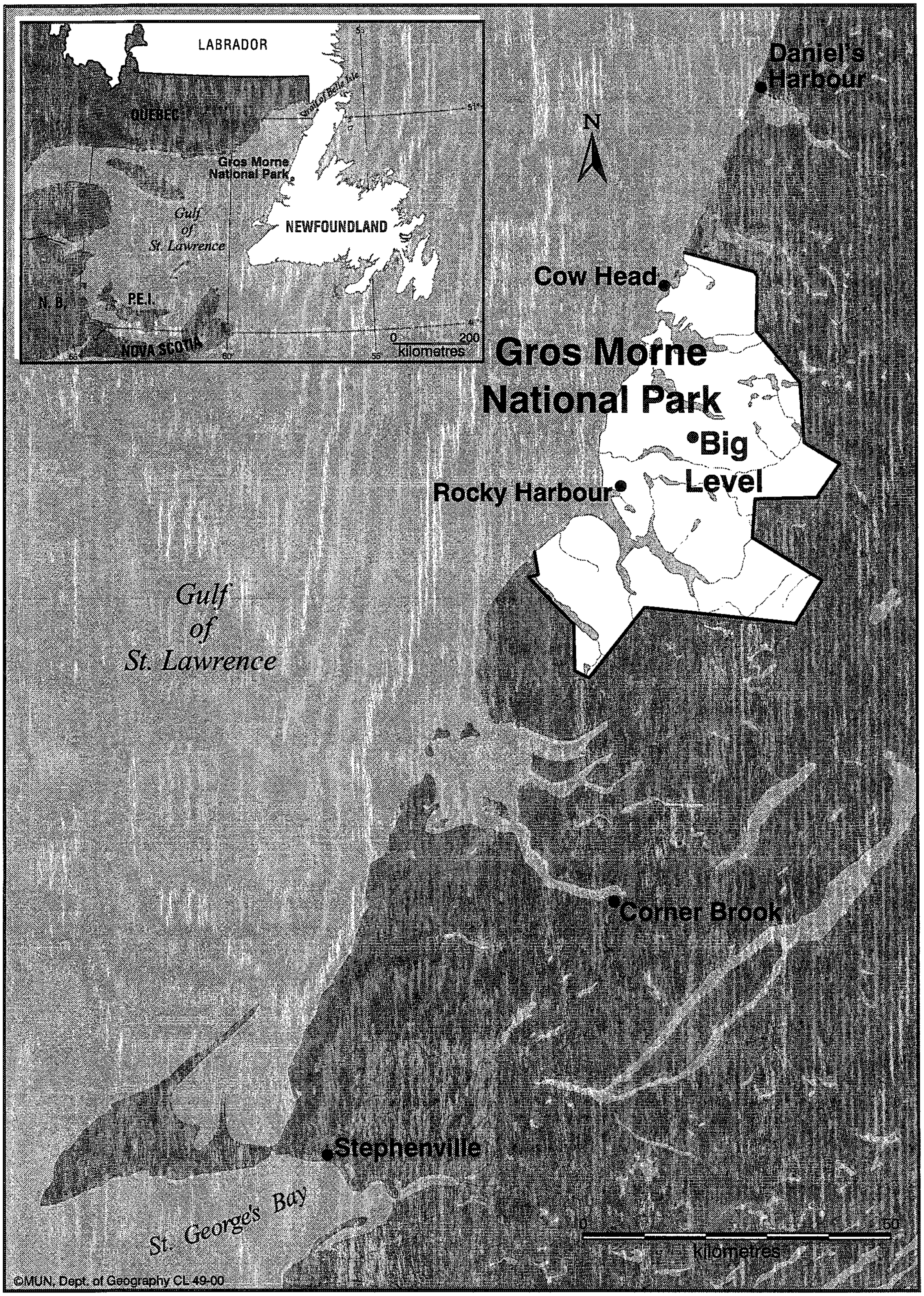


Figure 2.1 Gros Morne National Park and Newfoundland's west coast.

most part, well defined on their occidental margin by very steep slopes or cliffs of approximately 300 m relief. The previously named former fjords cut deeply into the range. The maximum elevations in these highlands are limited to near 800 m a.s.l., which is approximately 150 m above the local tree line. At these upper elevations, the alpine vegetation consists of heath-lichen tundra in exposed areas, with sedge fens and bogs in poorly drained areas. The dominantly gneissic bedrock often protrudes to the surface (Berger *et al.*, 1992, Brouillet *et al.*, 1998). Krummholz is often present near the upper tree limit, in the transition between the alpine barrens and the balsam fir forest of the lower and more protected areas (Berger *et al.*, 1992).

The most extensive area of high ground within the park is the Big Level Plateau. Its highest point of 797 m is located at 49°41' N, 57° 44' W, with the plateau's longest axis being perpendicular to the coast (WNW-ESE) (Figure 2.2). This highland is extensively dissected by small rivers and streams. It is clearly delimited to the north and south by Western Brook Pond and Bakers Brook Pond, respectively. The western and eastern margins are not as well defined, as more hills or mountains extend beyond the plateau limits. The plateau is thus defined here as the continuous land area above 650 m, giving it an area of roughly 60 $km^2$. Its western edge is separated from the coastal plain by a narrow valley and several hills reaching above 700 m in altitude. The eastern half of the plateau is somewhat lower and more uneven while the western half, which includes the summit, is higher and flatter (Brouillet *et al.,* 1998). Hence, a sub plateau can be defined, the Big Level Summit Plateau, which is the focal point of this thesis.

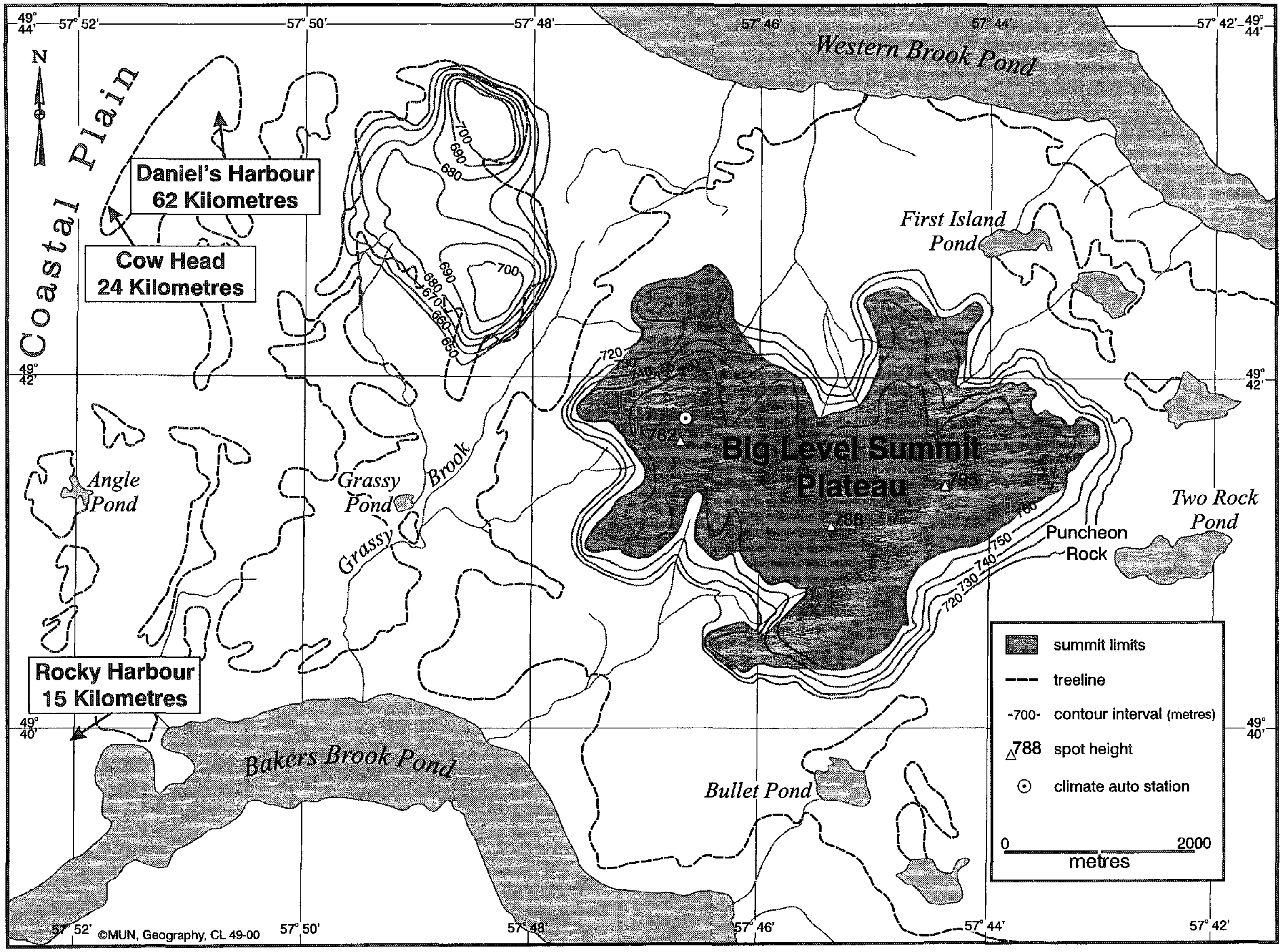


Figure 2.2 Big Level Summit Plateau

The Big Level Summit Plateau is delimited by steep slopes: the edge of the plateau ranges in elevation from 720 m in the NE to 760 m in the SW (Figure 2.2). It covers an area of 12 $km^2$ and undulates gently, revealing three high points from W to E: 782 m, 788 m and the summit of 797 m. Along its longest axis (WNW-ESE), the summit plateau spans 4.9 km, while the shortest distance across the plateau of 1.6 km is perpendicular to the previously mentioned axis, in the western portion of this highland. To the northeast of the sub-plateau edge, near the shortest distance across it, is a deep bowl-shaped depression capable of accumulating considerable amounts of drifting snow. Moreover, the plateau is the site of many erratic boulders, a sign of previous glaciation.

The vegetation of the summit plateau is consistent with the details mentioned above concerning the Long Range Mountains in general. Details of the plateau's vegetation are discussed in Brouillet *et al.* (1998), including reference to rare alpine vascular plant species. Twenty-one of these rare species are encountered on the summit plateau and they are all associated with the plateau's late-lying snowbeds (Brouillet *et al.*, 1998). The vegetation cover provides a summer feeding ground for a large population of caribou. Finally, the Big Level Plateau is considered to be a special conservation zone within the Park, consisting of rare or unique elements that could be altered by visitor activities (Brouillet *et al.*, 1998).

2.1.1 Climate

The climate of Newfoundland in general and Gros Morne National Park in particular are topics well covered by Banfield (1983 and 1990 respectively), although the climate of the Long Range Mountains is recognized as requiring further study through substantially more meteorological observations . The Gulf of St. Lawrence exerts a significant influence on the climate of the west coast of Newfoundland, affecting the thermal and moisture characteristics of the airstreams from the North American continent (Banfield, 1983). Furthermore, the waters off this coast are kept relatively cool by a ribbon of the Labrador Current encircling the Island. The waters off Gros Morne Park do not normally exceed approximately 13°C even in July (Hare, 1952). This maritime influence normally keeps the diurnal range of temperatures relatively small, especially in the fall (Hare, 1952). During the winter, northwesterly airflows often bring blizzard conditions as moisture is picked up from this water body (Banfield, 1983). However, the freeze-up of the Gulf in mid- to late winter (February in most years) cuts off or reduces this moisture source and the sea ice can, furthermore, result in coastal air temperatures nearly as low as for continental areas to the west (Hare, 1952; Banfield, 1990). The low water temperatures contribute to the late arrival of the spring season. Precipitation is plentiful year round. The climate of the Park is also markedly influenced by the relief of the Long Range Mountains (Banfield, 1983). The principal effect is to bring about a general decrease in temperature with height, along with increased cloudiness and precipitation and an increase in wind velocity as one proceeds vertically up the range (Banfield, 1990).

As with other mid-latitude locations, the synoptic weather conditions are often a dominant factor determining local weather conditions, especially in winter. The mean position of the Arctic Front in the cold season determines the relative frequency of frontal cyclones versus anticyclones. When the Arctic Front is to the south of Newfoundland, precipitation bearing cyclones tend to remain to the south and east of the Gros Morne Park area, allowing cold and mostly clear weather to prevail, from anticyclones extending from north-central Canada or Quebec-Labrador (Banfield, 1990). However, snow squalls or blizzards can develop during onshore W-NWly flows during these conditions, as long as the Gulf is open. However, the largest portion of the annual precipitation falling over Gros Morne National Park is due to synoptic-scale frontal cyclones, which frequently track over or near the island in winter, bringing heavy snowfalls to higher ground especially. These are essentially responsible for the building of the snowcover. In contrast, when the Arctic Front lies to the north of the study region, cyclonic storms are relatively more frequent, with incursions of mild air and occasional rain leading to winter thaw events, owing to the passage of the cyclonic warm sectors over the area (Banfield, 1983; Banfield, 1991).

Banfield (1981) refers to the overall climate of the island of Newfoundland as modified continental, because of the significant but not overwhelming effect of the sea. The climate of Gros Morne National Park can be classified as Dfc in the Köppen classification system (Banfield, 1983), which refers to a "subarctic microthermal" climate with cool summers and cold winter. This climate category has between one and four months with mean temperature above 10°C, the coldest month below 0°C and precipitation year round (Christopherson, 1997). In Thornthwaite's classification scheme, the Gros Morne area can be classified as "perhumid", i.e. where precipitation greatly exceeds evaporation (Banfield, 1983).

Precipitation on the coastal plain can be deduced from available Meteorological Service of Canada (MSC) mean values for the coastal plain climate stations. Banfield and Jacobs (1998b) evaluated the annual precipitation to be approximately 1200-1450 mm for the coastal plain and they estimated well over 1700 mm/yr for the mountains, derived from stream runoff data. On the coastal lowlands, approximately one third of the annual precipitation falls as snow, on average, and these low elevations usually have a complete snowcover from early December until early April (Banfield and Jacobs, 1998b).

On the Big Level summit plateau the vegetation is a very good indicator of its local climate. The alpine tundra found on the plateau can be attributed to strong winds and abrasion by blowing snow, low winter and summer temperatures and a short frost-free growing season, among other factors. Such proxy indicators could be of importance in the absence of a lengthy meteorological record for high ground of the Long Range Mountains.

The fairly lengthy climate record of the coastal plain permits an analysis of most climate variables. Some important variables for this study, other than temperature, precipitation, winds and snow water equivalent (SWE) or snow depth, which are covered in other chapters, are briefly examined in the following paragraphs.

The most fundamental component of the climate system is solar radiation and, in most cases, it represents a major portion of the energy absorbed at the surface (Male and Gray, 1981). In Gros Morne National Park, the solar elevation at noon varies from 17° at the winter solstice to 64° at the summer solstice, which determines the seasonal variation in amount of global solar radiation that can reach the surface. Although global solar radiation has never been measured at climate stations operating on the coastal plain at GMNP, values have been derived with the input of cloud cover records (Hay, 1977). Two stations located north and south of the park near the coast at low altitude have such cloud cover records (Daniel's Harbour and Stephenville, respectively). From data published by Hay (1977), Banfield (1990) reported that over long periods the monthly mean values of global solar radiation vary little between both these stations outside the Park. Therefore, the derived solar record for Daniel's Harbour, which is much closer to the Park than Stephenville, was considered representative of the coastal plain. Mean daily global solar radiation values for each month of the year are presented in Figure 2.3. The mean derived global solar radiation values vary considerably from season to season with a maximum of 18 MJ $m^{-2}$ $d^{-1}$ in June and a minimum of 2 MJ $m^{-2}$ $d^{-1}$ in December (Environment Canada, 1985). Values for the Long Range Mountains would be lower because of increased cloudiness (Banfield, 1990) in spite of the fact that solar radiation values increase with altitude when cloudiness is held constant, because of decreasing atmospheric turbidity with height (Barry, 1992).

The incident global radiation amounts are well below the maximum possible because the area experiences considerable cloudiness. To illustrate this, the record of bright sunshine hours from the coastal plain inside the Park is examined (from the Rocky Harbour station

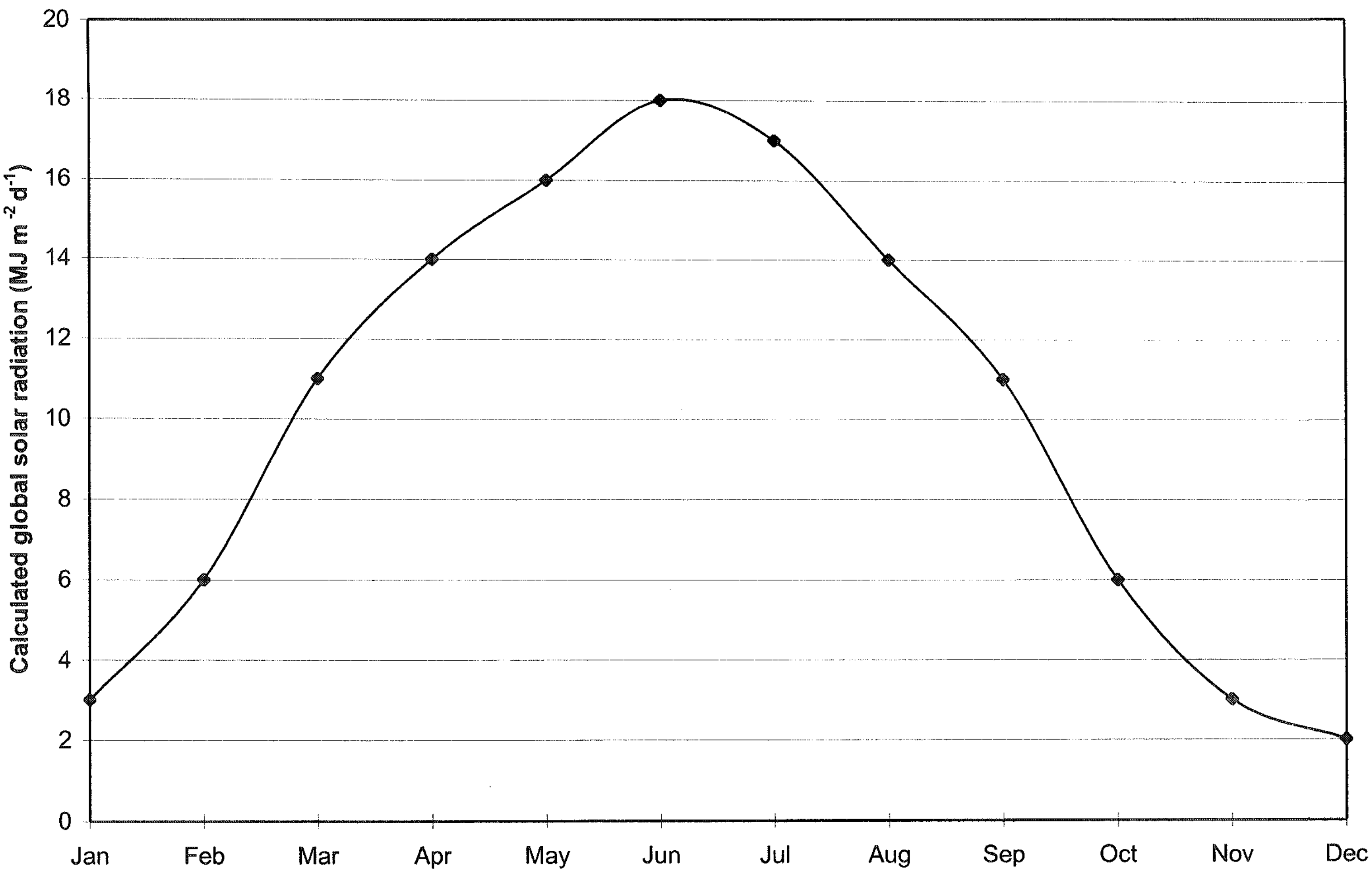


Figure 2.3 **GMNP coastal plain mean global solar radiation.** Data from Daniel's Harbour (1953-1980). Source: Environment Canada (1985).

record). The most sunshine that one can expect to see on average is in July, when the average duration is about 40% of the maximum possible (Figure 2.4) (Banfield, 1990). The warmer season generally has a greater fraction of possible sunshine than winter. Thus, in December and January, one can expect to have less than 10% of the maximum possible amount of bright sunshine; this represents an average of less than one hour per day (Banfield, 1990). It is obvious from Figure 2.4 that cloud cover is frequently considerable during daylight hours. Furthermore, higher ground will experience even more cloudiness due to additional condensation brought on by the mechanical uplift and adiabatic cooling of air (Banfield, 1990; Barry 1992).

The proximity of the Gulf of St. Lawrence, compounded with the large presence of cloud, is an indication of the high relative humidity experienced in the Park. In fact, the monthly average relative humidity is above 80% throughout the year (Figure 2.5) (Environment Canada, 1998). The record of vapour pressure, a measure of absolute humidity, shows clearly that more water vapour is present during the warmer months, as expected. The average relative humidity in the mountains would be higher than on the coastal plain due to lower temperatures and the presence of fog caused by low cloud ceilings (Banfield, 1990). However, the absolute humidity amount would be roughly the same or perhaps even a little lower as absolute humidity decreases with temperature and therefore altitude (Barry, 1992).

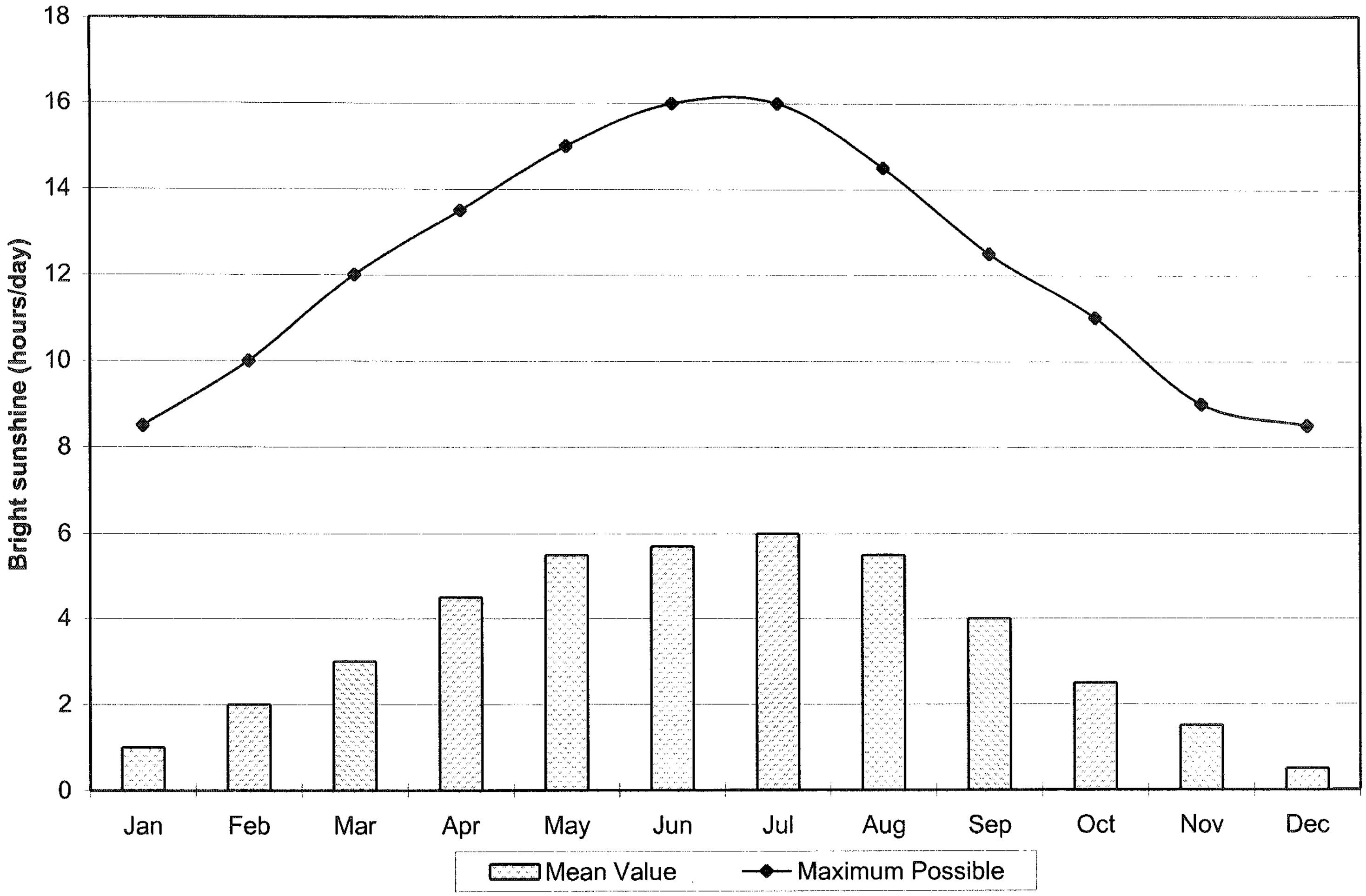


Figure 2.4 **GMNP coastal plain bright sunshine.** Data from Rocky Harbour (1976-1987). Source: Banfield (1990).

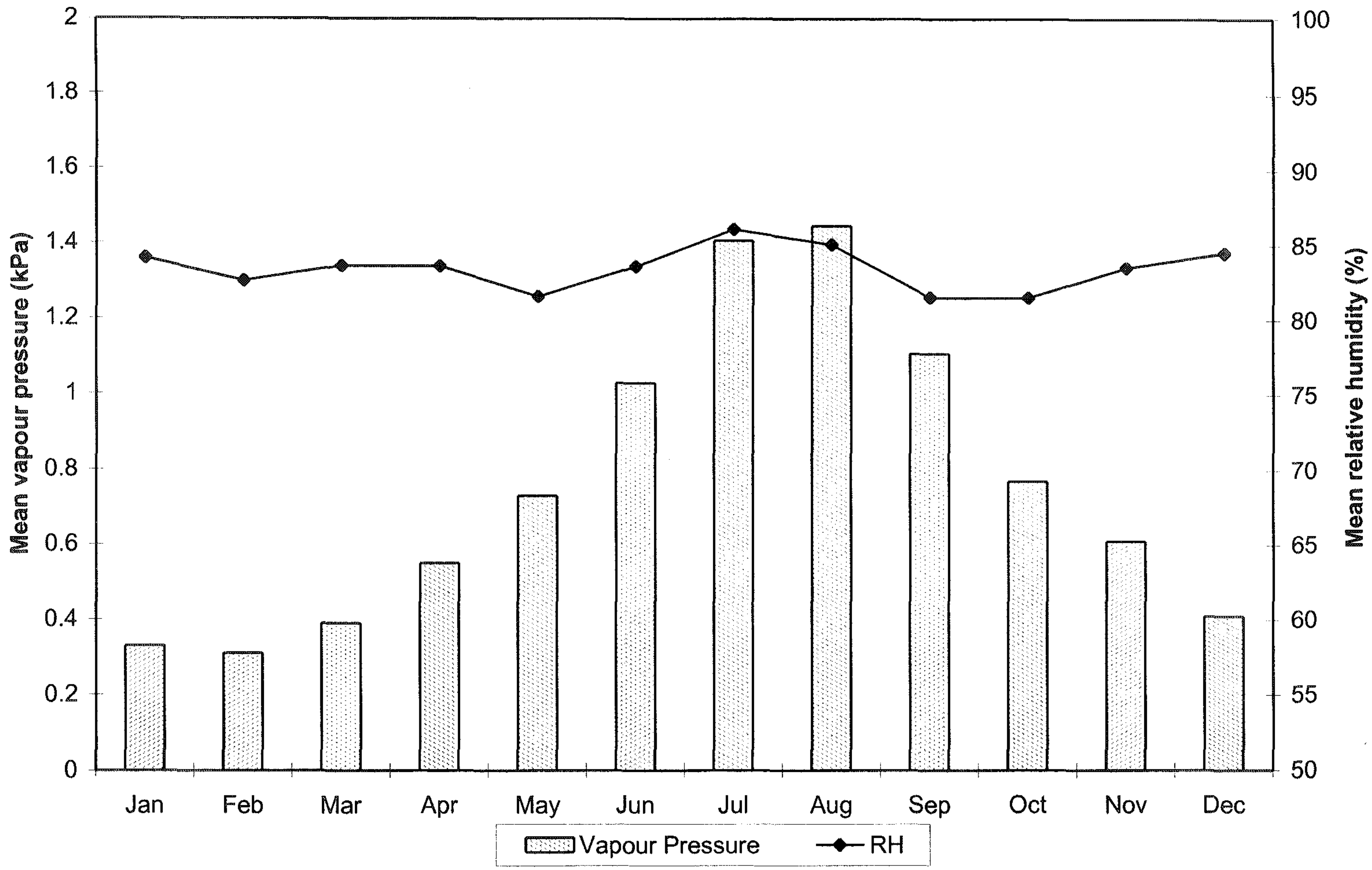


Figure 2.5 **GMNP coastal plain humidity.** Data from Daniel's Harbour (1961-1990). Source: Environment Canada (1998).

The climatic generalizations presented above apply only to decadal scale time periods. Individual seasons can (and do) depart significantly from these climatically "normal" conditions, as will be demonstrated by this research for temperature, winds, precipitation and snowcover.

## 2.2 The Climate Stations

### 2.2.1 The Big Level Summit Plateau Autostation

The Big Level Summit Plateau Autostation is located on the alpine plateau with the same name at 49° 41.8' N, 57° 46.6' W, 778 m a.s.l., 13 km inland to the east of the coast (Figure 2.2). It was set up by Parks Canada and the Wildlife Division of the Newfoundland Department of Tourism and Culture and it is operated by Parks Canada and the Gros Morne climate studies research group. In operation since 3 September (Julian Day [JD] 246) 1993, the station measures several climate elements. Shade air temperature and relative humidity at 1.8 m above the surface are measured from a temperature and relative humidity probe by *Vaisala* inside a Gill multi-plate radiation shield. From the station's single mast is also attached an *R. M. Young* wind monitor measuring both wind speed and direction at a height of 3 m. Rainfall was measured with a *Texas* tipping bucket precipitation gauge at ground level during the snow free season from September 1993 up to October 1997. The station was compiling data on a daily basis up to 6 August (JD 219) 1996 when an hourly recording program was added. At this time, a Fisher and Porter weighing type precipitation gauge with a free swinging Alter shield, collecting both rainfall and snowfall, and a *Licor* pyranometer measuring global solar radiation (K↓) were installed. The data collected by all the instruments are stored in a data logger and downloaded into a portable computer during visits to the station occurring usually each season. Specific details on the instruments, their units of measurement and their accuracy are provided in Table 2.1.

Although the station operated from 1993 up to and beyond the end of this project's study period in June of 1999, its data record is not complete. The harshness of the plateau's climate compounded with the logistical difficulties in visiting and servicing the station has resulted in several periods of missing data. High winds, freezing precipitation, blowing snow and even a probable lightning strike, along with station repairs, have shut down the station completely during eleven intervals lasting a total of 406 days out of the total of 2146 days from September 1993 to June 1999 (19% of total days). Furthermore, individual sensors have had difficulties at different points in time.

The meteorological element with the most complete record is air temperature. In addition to the total station "blackouts", only one missing temperature data event exists and it is only a two day period in September 1993 (JD 257-258). Overall, 82% of the temperature measurements are available for the study period. The anemometer encountered far more difficulties as expected in the very windy and icy environment of the BL station.

Table 2.1 **Meteorological variables, instruments and their uncertainties at the Big Level Autostation.**

| Meteorological variable | Instrument used | Uncertainty of measurement |
|---|---|---|
| Daily temperature ($T_{max}$, $T_{min}$, $T_{mean}$) (in °C) | Platinum resistance temperature detector in the HMP45C Temperature and Relative Humidity Probe by Vaisala | ± 0.4 °C within -30 to +40 °C[1] |
| Relative humidity (in %) | HUMICAP® 180 in the HMP45C Temperature and Relative Humidity Probe by Vaisala | ± 7%[2] |
| Global solar radiation ($Wm^{-2}$) | LI-200SZ Pyronometer by LI-COR | greater of ± 5% or ± 0.02 $MJ^{-2}$ $h^{-1}$ [3] |
| Wind speed (m/s) | R.M. Young model 05103 wind monitor | ± 0.5 m/s [4] |
| Wind direction (degrees) | R.M. Young model 05103 wind monitor | ± 5° [4] |
| Rainfall (mm) | Texas TE525 tipping bucket rain gauge | greater of ± 1 mm or ± 5%[5] |
| Total Precipitation (mm) | Fisher and Porter Weighing Precipitation Gauge with Alter shield | Rainfall = greater of ± 1 mm or ± 5%[6]. For snowfall see section 5.1 |

*Notes:*

1 *Uncertainty recommended by the manufacturer. This value is the largest within the specified temperature range. Furthermore, the AES Guidelines for Co-Operative Climatological Autostations version 2.0 (1992) lists an expected uncertainty for similar sensors of ± 0.3°C.*

2 *Uncertainty specified by the AES Guidelines for Co-Operative Climatological Autostations version 2.0 (1992) for the same sensor. This value is preferred to the manufacturer's, which is smaller and more complex to calculate, because of the harshness of Big Level winter conditions and the long periods between cleaning of the radiation shield. The manufacturer's uncertainty is as follows:*

*± 2% (0 to 90% rh)*
*± 3% (90 to 100% rh)*
*These apply at 20°C, add ± 0.05% per °C difference with this temperature.*

3 *Uncertainty specified by the AES Guidelines for Co-Operative Climatological Autostations version 2.0 (1992) for the same sensor. The value is preferred to the manufacturer's because of the harshness of Big Level winter conditions and the long periods between cleaning of the radiation shield. The manufacturer's uncertainty value is of a maximum of ± 5%, which would be inaccurate for dawn, dusk and small winter values. The effect of snow accumulation, dust and dirt in between clean-ups is not considered, but the data should be used with caution.*

4 *Uncertainties stated by the WMO Guide to Meteorological Instruments and Methods of Observation (1997). The values were stated as typical operational performances determined by the Commission for Instruments and Methods of Observation in 1993. These values are slightly higher than those specified by the manufacturer:*

*wind speed: ± 0.3 m/s*
*wind direction: ± 5°*

*However, the AES Guidelines for Co-Operative Climatological Autostations version 2.0 (1992) lists larger maximum uncertainties for the same sensor.*

5 *The values given in the AES Guidelines for Co-Operative Climatological Autostations. In version 2.0 (1992), the value is listed as greater of ± 0.2 mm or ± 2%. However, in version 1.0, the largest uncertainty given is the greater of ± 3.0 mm or ± 5%. Considering the exposed location of the Big Level Autostation and the high winds experienced there, a large value should apply. However, since the resolution of the gauge is ± 0.1 mm, the largest uncertainty is too large for small precipitation amounts. Furthermore, there is no need to assume a smaller daily uncertainty than ± 1 mm for the purposes of this study.*

6 *See comment no. 5 for rainfall uncertainty values. However, in this case the minimum is ± 1 mm because it is also the resolution of the gauge. Values for snowfall are dependent on wind speed, and are presented in the text.*

Only 43% of the days since the station inauguration date have valid wind speed data, with most of the problems in addition to the total station blackouts occurring during winter. Precipitation measurements also experienced heightened difficulties during winter. In the months of November to April from 1993 to 1999, only 180 days of reliable precipitation data exist. The Fisher/Porter gauge was on occasion filled to capacity, while capping of the gauge orifice seems to have been an important problem during winter storms, as deduced from comparing the gauge collection behavior in comparison with the coastal plain stations. In the longer study period of the months from October to June, a total of 461 days are available, with the tipping bucket rain gauge accounting for 19% of those measurements. Overall, 813 days of precipitation data are available at BL from 1993-1999, which is 38% of the total operation period.

### 2.2.2 Local and Regional Climate Stations

There are currently two climate stations with more than a decade of record operating in Gros Morne National Park. Both are operated by the Meteorological Service of Canada (MSC), formerly AES, and are located on the narrow coastal plain at Rocky Harbour and Cow Head (Figure 2.1). Two other stations are also of considerable importance to the study: the Daniel's Harbour synoptic reporting station with its much lengthier record extending back to the 1940s, and the Stephenville A (airport) upper air sounding station, also operated by the MSC (Figure 2.1). Table 2.2 provides a summary of information concerning these four stations and Table 2.3 indicates the uncertainty values related to the data.

The closest station to the alpine plateau of particular interest is the Rocky Harbour (RH) station, situated 14 km to the SSW of the Big Level station (BLS), in hilly terrain but with generally good exposure to the Gulf to the west. It has been measuring temperature, precipitation (rainfall and snowfall), and snow on the ground (SOG) for almost three decades. It has also been recording bright sunshine hours since 1976. However, a perturbation in its record occurred in 1993 when the station location was changed. The first location was at the Parks Canada compound with several buildings in the vicinity. It was moved by approximately two kilometers to the east from its original location, which resulted in a 32 m increase in its elevation. Due to the small period of overlap between the original, manned, climatological station and the new autostation (four months), no correction was carried out to remove the differences in the data between the two stations. This change in site was therefore overlooked in practical terms and the data set was merged, but used with caution. Furthermore, the autostation has an hourly program which also measures wind speed and direction, relative humidity, as well as snow depth with a snow depth sensor. Unfortunately, this has also resulted in separate measurements of rainfall and snowfall being no longer distinguished, being replaced by total precipitation from a Fisher and Porter recording precipitation.

The other station within the boundaries of the park is the Cow Head (CH) station, sited adjacent to a house immediately above the shoreline in a community. As a manned Climatological Station it measures daily maximum and minimum temperature, daily rainfall

Table 2.2 **Meteorological Service of Canada stations used in this study.**

| Station Name | ID. Number | Lat. and Long. | Altitude (m) | Period of Record | Comments |
|---|---|---|---|---|---|
| Cow Head | 8401335 | 49° 54' N 57° 48' W | 15 | 12/1982 - present | Manned climatological station |
| Daniel's Harbour | 8401400<br>8401410 | 50° 14' N 57° 35' W | 19 | 11/1946 - present | Synoptic station until 1996, converted to an autostation in 02/1998 |
| Rocky Harbour<br>Rocky Harbour CS | 8403096<br>8403097 | 49° 35' N 57° 54' W<br>49° 34' N 57° 53' W | 68<br>100 | 07/1972 - present | Manned climatological station, converted to an autostation and changed site in 01/1993 |
| Stephenville A | 8403800 | 48° 32' N 58° 33' W | 60 | 06/1957 - present | Upper air sounding station, missing data 03/1994 - 09/1997 |

Table 2.3 **Uncertainty values of the meteorological data.** Data obtained from the Meteorological Service of Canada (Environment Canada).

| Meteorological variable | Uncertainty of measurement |
|---|---|
| Air temperature (°C) | ± 0.2°C |
| Relative humidity (%) | ± 5% |
| Atmospheric pressure (mb) | ± 0.3 mb |
| Cloud amount (in eights) | ± 1/8 |
| Wind speed (m/s) | ± 0.5 m/s |
| Wind direction (degrees) | ± 5° |
| Precipitation amount (mm) | ± 5% |
| Depth of snow (cm) | ± 1 cm for values ≤ 20 cm<br>± 5% for values > 20 cm |
| Sunshine Duration (hours) | ± 2% |

*Note:*
*These uncertainty values were obtained from the WMO Guide to Meteorological Instruments and Methods of Observation (1997). The values were stated as typical operational performances determined by the Commission for Instruments and Methods of Observation in 1993. Depth of snow is an exception; the uncertainty was listed as the recommended accuracy for general operational use. As a member of the WMO, Environment Canada must adhere to these standards in order for its data to be included in the global climate archive. Canadian data (MSC) is included in this archive.*

with a Type B rain gauge and daily snowfall and SOG with a ruler. It is located 22 km north of the BLS and its record extends back to 1982.

To the north of the Park, Daniel's Harbour (DH) is of considerable importance because of its long data set representative of the coastal plain. At 61 km to the NNE of the BLS, it offers the nearest meteorological record that extends back prior to 1950. It is located on flat terrain in a community with low residential housing. There are no houses within at least a hundred meters of the station however, and no topographic obstacles within at least 500 m. As a principal synoptic station it measured values of hourly temperature, precipitation, winds, atmospheric pressure, relative humidity and SOG, to mention the main variables (Environment Canada, 1985). At this station a Nipher precipitation gauge has been used since 1962, permitting a more reliable measurement of snowfall. However, in 1996 it was downgraded to a regular, i.e. non-synoptic, hourly climate station and there are considerable gaps of missing data beyond this year. In 1999, measurements were still not taken consistently on a daily basis.

Finally, the Stephenville A station provides the only upper air sounding in western Newfoundland. It is located 140 km to the SSW of the BLS with twice daily soundings at 0 and 12 Z. Height, pressure, temperature, winds, and relative humidity are recorded at all standard and significant levels since before the 1950s. Digital copies of these data are available since June 1957, although a recent gap in the record exists from 1994 to 1997. The uncertainties of the upper air data are presented in Table 2.4.

Table 2.4 **Uncertainty values of the upper air data.** Data obtained from the Meteorological Service of Canada (Environment Canada).

| Meteorological variable | Uncertainty of measurement |
|---|---|
| Atmospheric pressure (mb) | ± 2.0 mb |
| Air temperature (°C) | ± 0.5°C |
| Relative humidity (%) | ± 5% when T > 0°C<br>± 10% when T ≤ 0°C |
| Wind direction (degrees) | ± 5° |
| Wind speed (m/s) | * ± 1 m/s |
| Geopotential Height (m) | ± 1% |

**According to Clark (1969) wind speed uncertainties are dependant on height and wind speed and a substantially complex calculation method is offered. The value shown is the one in the WMO (1997) and is simplified.*

*The uncertainty values (std errors) were retrieved from Clark (1969) and the WMO (1997). The larger values were used to assure the compliance of the older upper air data with the uncertainties. In most cases, the difference between the two sources are negligible. The current radiosonde being used at the Stephenville station is the Vaisala RS80-15LH. It has been in use for 2 years. Unfortunately, no accuracy information on this model is offered by the manufacturer or by Environment Canada.*

# 3 HISTORICAL BIG LEVEL TEMPERATURES

## 3.1 Background

Snow temperature is the single most important factor determining the state of a snowcover. Snow can only exist at temperatures at or below the freezing point; when the snowpack reaches 0°C, it begins to melt. Temperatures well below the freezing point keep light snow crystals loosely bonded after a snowfall, enabling them to be easily transported by wind (McKay and Gray, 1981). Air temperature above a snowpack is often used alone to determine snowmelt, despite the fact that it also depends on other factors such as solar radiation, albedo and humidity content of the air to name a few. However, Zuzel and Cox (1975) affirmed that air temperature is the best single meteorological variable to determine snowmelt.

A degree-day approach can serve as a first approximation for the amount of snowmelt. For example, monthly degree-days above 0°C, which represent the accumulation of heat energy above 0°C for a given month, provide information on whether any significant melt events have occurred over that period and their total magnitude.

The seasonal temperature regime of a given location broadly defines its snowcover season, which is mainly a winter phenomenon at lower elevations of mid-latitudes. If one defines winter as the season in which mean daily maximum temperatures are below 0°C, then winter occurs from December to March on the coastal plain of GMNP (Figure 3.1).

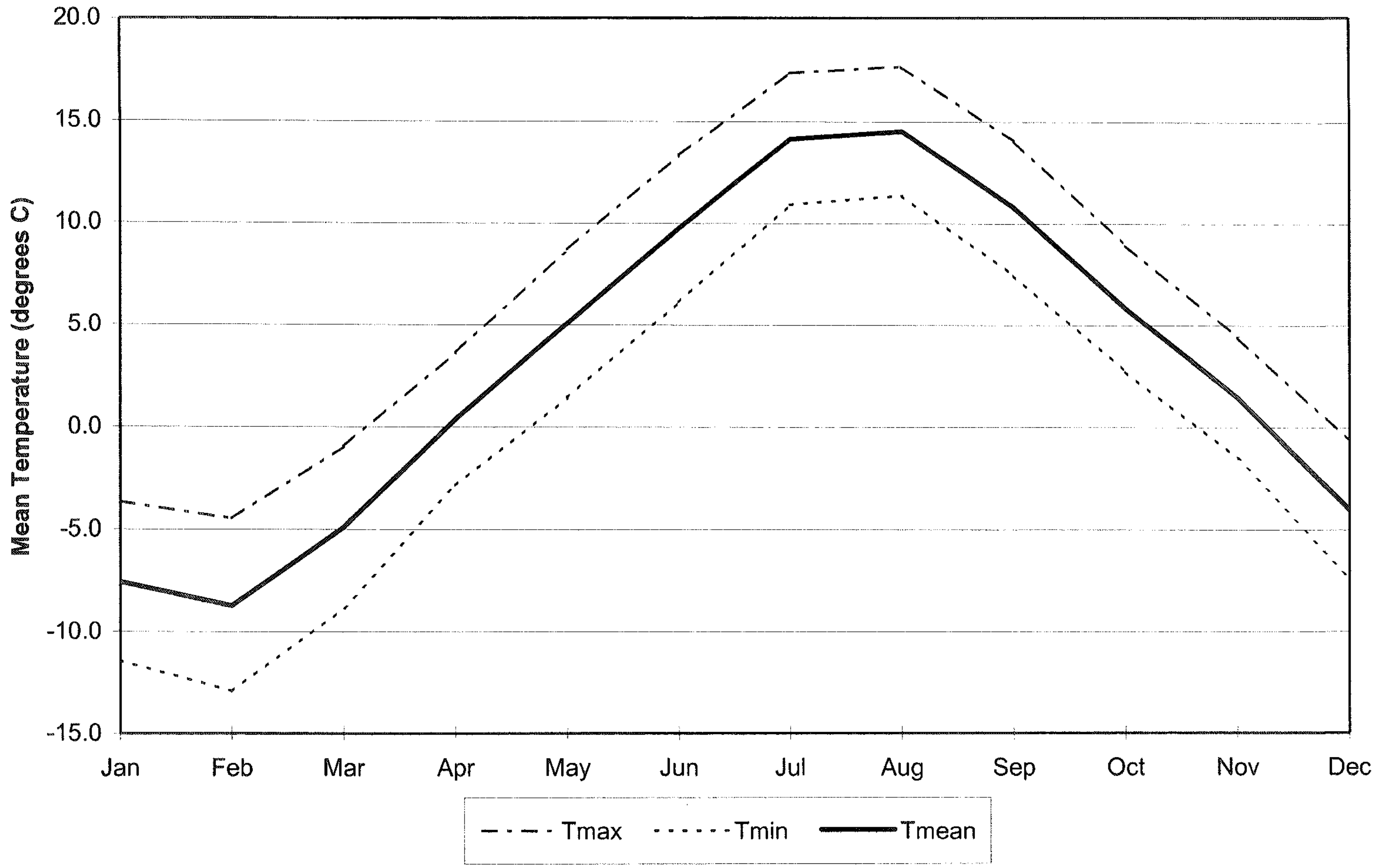


Figure 3.1 **Daniel's Harbour historical monthly mean temperatures**. Period from November 1946 to June 1999 with 2-8 missing years per month. Tmax = daily maximum temperature, Tmin = daily minimum temperature, Tmean = daily mean temperature.

Furthermore, the coldest month of the year in this region is February, two months after the winter solstice, while the warmest time of the year is late July - early August on the coastal plain. Spring on the lowlands is relatively cool for this latitude, because of the effects of cold water in the Gulf and the melting snowpack early in the season (Banfield, 1990). Fall on the other hand is relatively warm due to the thermal inertia (lag) effects associated with the Gulf. The first month with mean daily minimum temperature below 0°C is November, preventing a deep snowcover from appearing before this month in most years on the coastal plain.

Twentieth century temperature trends in Atlantic Canada and more specifically in western Newfoundland have differed from those of the country as a whole (Berry, 1991; Banfield, 1991; Skinner and Gullet, 1993). For the forty-year period 1950-1989, Skinner and Gullet (1993) found that while there was a non-significant increase in mean annual temperature of 0.4°C for Canada as a whole, there was a significant decrease of 0.6°C in Atlantic Canada. This is quite different from the longer 1895-1989 period during which an increase in mean annual temperature was found for both Canada as a whole and the Atlantic Canada region, albeit the increase being much larger in Canada as a whole: a significant 1.1°C versus a non-significant 0.4°C (Skinner and Gullet, 1993). Morgan *et al.* (1993) reported that the warmest conditions in the Atlantic region of the country, as of 1990, occurred in the 1950s, which does not conform with other areas of the country. The 1960s were rather variable with considerable cooling occurring quite abruptly in the early 1970s, reaching a low at this time. Generally cool temperatures from the early 1970s up to 1990 was interrupted only in the warm early 1980s. Banfield and Jacobs (1998a) found that this recent cooling trend continued until 1995 for Newfoundland.

The amplitude of interannual temperature variability has not been distributed uniformly throughout the seasons. The winter season has been much more variable than the other seasons to the extent that its signal dominates the pattern on an annual scale (Banfield, 1983; Morgan *et al.*, 1993). In fact, Banfield and Jacobs (1998a) found little variability in Newfoundland's summer temperatures during the past century especially on the west coast of the Island (Stephenville), whereas the province's winter temperatures (Dec-Mar) of the last century (1895-1995) could be divided into five distinct epochs:

1) 1895-1919: near period average (1895-1995)
2) 1920-1935: cold
3) 1936-1950: near period average
4) 1951-1971: warm
5) 1972-1995: cold.

The relatively large historical variability in winter temperatures and the presence of the above five epochs are due to large temperature differences of the varying winter air masses and the change in observed frequencies of these air masses that can cover large portions the island during this season (Banfield, 1983; Banfield and Jacobs, 1998a). The temperature records from Daniel's Harbour, (DH) (1946-1999) and Rocky Harbour (RH)(1972-1999) support the presence of these epochs for the coastal plain of GMNP during the period of its meteorological record. Figure 3.2 shows coastal plain temperatures for the snow season, defined here as October-May, months which can experience significant snowfalls or snow

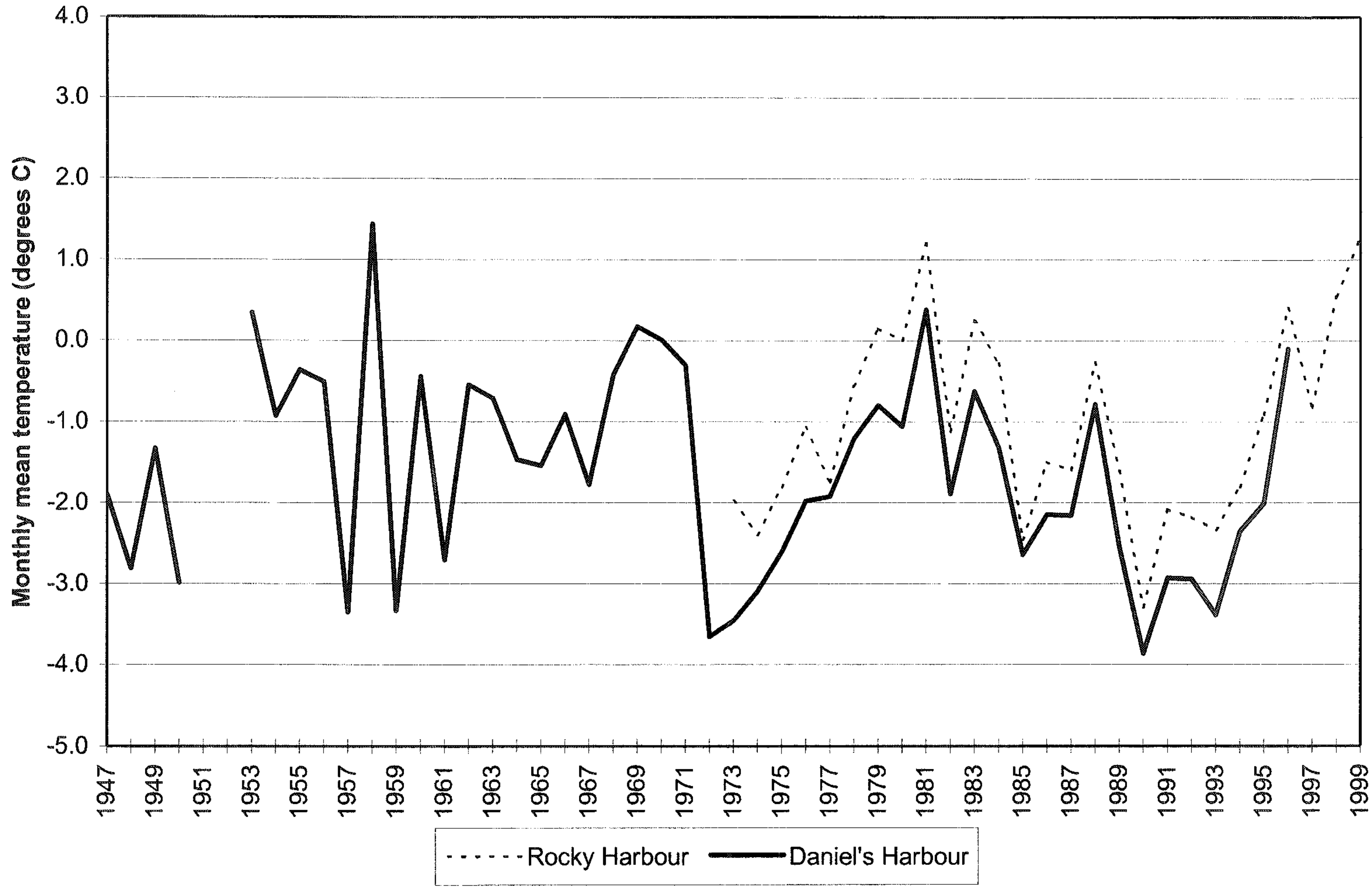


Figure 3.2 **Coastal plain mean snow season temperature**. October (previous year) - May mean temperature for Rocky Harbour and Daniel's Harbour stations.

on the ground ($\geq$5 cm). The years 1951 and 1952 are absent because of too much missing data. The record from Corner Brook (90 km to the south of BL) suggests that both missing snow seasons were very warm for the area, being between the extremes of 1953 and 1958, with 1951 being warmer than 1952. Sharp interannual variability marked the period from 1957-1962. Subsequent winters until 1971 were generally warm. The relatively lower temperatures that marked the last epoch ended in 1995, as confirmed by the much milder conditions of the snow seasons from 1996 to 1999. The temperatures of the last years of the record are comparable to the higher winter values of the 1950s.

There is a relationship between coastal plain and Big Level temperatures. It is well recognized that the free air decrease of temperature with height, or environmental lapse rate, averages 6.5°C/km over large spatial scales (Tabony, 1985; Barry and Chorley, 1998). The largest possible positive lapse rate value for an ascending air parcel is 9.8°C/km, which is known as the dry adiabatic lapse rate (Barry, 1992). Higher positive environmental lapse rate values are possible in very unstable air mass conditions. In contrast, environmental lapse rates can be negative (increase in temperature with height), known as inversions, which are caused by very stable conditions, such as due to rapid nocturnal surface radiational cooling or possibly summer daytime coastal sea breezes. An analysis of radiosonde data from Stephenville by Banfield (1983) reported an average annual lapse rate of 4°C/km for the first km above sea level, which was considered to be representative of most of the west coast of Newfoundland. The relatively reduced lapse rate could be in part the result of generally low sunshine amounts which tend to relatively decrease the temperature near the surface

(Harding, 1979). Furthermore, cool sea temperatures during the summer and sea ice during the late winter may also be causal factors, which also lower the air temperatures near the surface (Tabony, 1985; Banfield, 1990). Hence, seasonal differences are notable, with greater lapse rates observed in the Gros Morne area during the fall and early winter, when the relatively warm water from the Gulf increases lapse rates (Banfield, 1990). Differing air mass characteristics are also a key to determining lapse rates, especially when considering short time periods.

The general consensus is that mountain summits encounter temperatures very close to those in the free air nearby (e.g. Tabony, 1985). On the other hand, more extensive highlands (plateaus) modify the low level air structure, which by and large leads to higher summer temperatures and lower winter temperatures due to radiative processes (Tabony, 1985). A comparison by Banfield (1995) of the temperatures between Big Level summit autostation (BL, 780 m a.s.l.) and RH (100 m) for the September 1993 to March 1994 period found that the mean temperature difference between the two stations was 4.5°C. This results in a lapse rate of 6.4°C/km, which is very close to the average environmental lapse rate of 6.5°C/km. On individual days the maximum temperature at BL was as much as 9-12°C lower than at RH; such days occurred mostly in late winter, perhaps partly due to snowcover contrasts between the coast and the summit (Banfield, 1995). Snowcover with its high albedo (i.e. whiter), tends to depress air temperatures above it (Langham, 1981). A fresher, whiter snowcover on BL could have been a factor in the large late winter daily maximum temperature difference observed in Banfield's (1995) study. Furthermore in spring, when the lowlands are mostly snow free while BL is still under a deep snowpack, the melting of the snow could depress the air temperatures on the plateau even further due to the large amount of energy consumed in the process (Tabony, 1985).

In summary, factors affecting the temperature lapse rate between the coastal plain and BL, such as ground cover which controls albedo (i.e. the presence of snow), the sea temperature along with the presence of sea ice, and the types of air masses occupying the region and their control over cloud cover, are important in determining the plateau-coastal plain temperature difference (Barry, 1992). These factors highlight the need to break down the year into smaller time periods, or seasons, with more homogenous meteorological conditions when attempting to determine highland temperatures from values at lowland stations. However, many authors have warned that this general type of statistical procedure could be flawed due to changes in the structure of the atmosphere in a changing climate (Beniston, 1997; Pepin *et al.*, 1997). Beniston (1997) reported that positive (i.e. warm) temperature anomalies increase with altitude in the Alps during periods of high pressure. During these events temperatures increased by 0.5°C at 500 m height and by 3°C at 2500 m compared to values at low altitude. Moreover, Pepin *et al.* (1997) suggest through empirical evidence in the British Isles that a warmer and more humid future climate would lead to a reduction in observed lapse rates. However, knowledge in this area is still premature. In addition to this fact, given the relatively low altitude of BL and the absence of a generalized warming trend in the historical coastal plain record, this study will assume that average lapse rates have been constant throughout the historical period in GMNP.

## 3.2 Methodology

Since the Big Level Summit Plateau autostation (BLS) has been measuring daily maximum ($T_{max}$) , minimum ($T_{min}$) and mean air temperature ($T_{mean}$) for only 6 years (as of 1999 - see section 2.2), the data collected can hardly be representative of the climate variability over an annual or decadal time scale. In the pursuit of assessing the snowcover conditions on the plateau for the latter part of the $20^{th}$ century, the development of a reliable historical temperature record is of utmost importance. The presence of several climate stations near BL provides the means to accomplish this goal by developing and applying the statistical relationships that exist between them. The geographical proximity of the Rocky Harbour (RH), Cow Head (CH), Daniel's Harbour (DH) and Stephenville A upper air (SUA) stations to BL offers the interesting opportunity of modeling their temperature relationships through linear regression analysis (see, for example, Clark and Hosking, 1986).

The task of estimating Big Level temperatures from coastal plain measurements will involve the effects of altitude. Given that there is a height difference of 678 to 763 m between the coastal stations and the BLS, a lapse rate can be computed for this elevation range and would be expected to vary from season to season (see section 3.1). In an effort to take these factors into account, the free air lapse rate from the surface to 780 m was calculated from the SUA for every sounding of the period of record. The 780 m values were determined by using linear interpolation between the levels immediately above and below this height. The twice daily lapse rates were then averaged into daily values in three different ways to account for the time of day and the meteorological controls which affect the daily maximum, minimum and mean temperatures. For maximum temperature, which normally occurs in early afternoon, as supported by the times of the daily maximums at BL, a daily average of the 830 NST (12 Z) and 1830 NST (0 Z the next day) lapse rates was used. Concerning minimum temperature, the average of the previous day's 1830 NST (0 Z present day) value and the current day's 830 NST value was deemed more appropriate because daily minimum temperatures, as verified by an analysis of BL's temperature record, normally occur at night or early in the morning. Finally, an average lapse rate using three soundings (1830 NST previous day, 830 NST and 1830 NST) was calculated to properly represent the entire climatological day (day begins at 230 NST and ends at 230 NST the following day). In order to reduce missing values, a daily lapse rate was calculated from only one sounding if it was the only one available for that day. The daily lapse rates thus calculated from the Stephenville record were then applied to the daily temperature records of the coastal plain stations in GMNP.

A record of BL temperatures from November 1946, the date of DH's inauguration, to June 1999 was then constructed using multiple linear regression. Daily maximum, minimum and mean temperatures and the calculated lapse rates from the neighboring stations were used as independent variables or predictors. The period which served to build the temperature models, for which data from all five stations were intermittently present, was from September 1993 to June 1999.

Daily maximum, minimum and mean temperature models were prepared separately for each month for a total of 36 models, using *SPSS for Windows (version 9.0).* The predictors for each model, i.e. one for each element for each month (e.g. maximum temperature for January), were selected using stepwise regression. Independent variables were added and removed using the 0.05 significance level threshold. Moreover, each model is composed of predictors from a combination of stations to account for days in the historical period where one or many of the predictor values are missing, or before their existence. Cow Head's record was excluded from this process because of questionable values in its record, but also because of the general redundancy with the other Park station at Rocky Harbour (RH). Furthermore, the lapse rate (LR) could not be used as a sole predictor since a LR is only useful if a temperature value is available as a reference point. Therefore, each monthly model can include a maximum of 6 equations, one for each possible scenario of available station data:

1) RH, DH, LR
2) RH, DH
3) RH, LR
4) DH, LR
5) RH
6) DH.

Each equation was ordered within each model by using the standard error of the estimate from the regression. The lower standard error equations were preferentially used before the higher ones, according to which independent variables were available.

The equations were carefully inspected with graphical methods to ensure each one met the assumptions of normality and constant variance of the variables. The linearity of the equations was verified by randomly plotting independent variables versus their respective dependant ones. All the equations were tested for first order serial correlation (or autocorrelation) using the Durbin-Watson test, which is the standard one for this problem (Clark and Hoskings, 1986). A detailed analysis of serial correlation and the Durbin-Watson tests are available in many econometrics texts such as Maddala (1977). In the case where the resulting d value fell below the likelihood of no autocorrelation ($D_U$) threshold at the 5% significance level, the null hypothesis of no serial correlation was rejected, as suggested by Clark and Hosking (1986), and a correction technique was used. The $D_U$ thresholds were determined from tables calculated by Savin and White (1977). They vary according to the number of independent variables and to the sample size.

The Generalized Least Squares method was used to correct for serial correlation as presented in Studenmund (1997). In this method, the "d" value of the Durbin-Watson test serves as the basis for a correction coefficient applied to the independent and dependent variables and the regression is subsequently redone. The new results are deemed to be free of serial correlation, and the statistics of the regression: the correlation coefficients, the standard error and the significance level are considered accurate. The resulting equation is easily reconverted to accept the initial variables, in the place of the modified ones used in the regression.

The equations were combined into single *Microsoft Excel* expressions and applied inside this spreadsheet program with all the independent variable values listed in columns. The results of the models are spread over three columns with the estimated value in one, the number of the equation used in the second and the uncertainty of the estimate in the third. Two standard errors (2 SE) of the estimate for the applied equation represents the 95% confidence interval for the value predicted by the model and was thus used as the uncertainty of the estimated temperature values. When means for longer periods were calculated, the daily uncertainties were averaged and divided by the square root of the number of values considered in the calculation. When sums are calculated, as would be the case for cumulative degree-days, the uncertainty would be the sum of the daily uncertainties divided by the square root of the number of values. Finally, monthly temperature and degree-days above 0°C values were compiled for the entire historical period and averages for the 1946-1999 period were also prepared.

## 3.3 Results and Discussion

### 3.3.1 Regional Free Air Lapse Rates

Monthly average lapse rates from the Stephenville upper air station are presented in Figure 3.3 for the period from 1957 to 1999. The lapse rates used for the determination of the monthly averages were those compiled for daily maximum temperature, i.e. the average

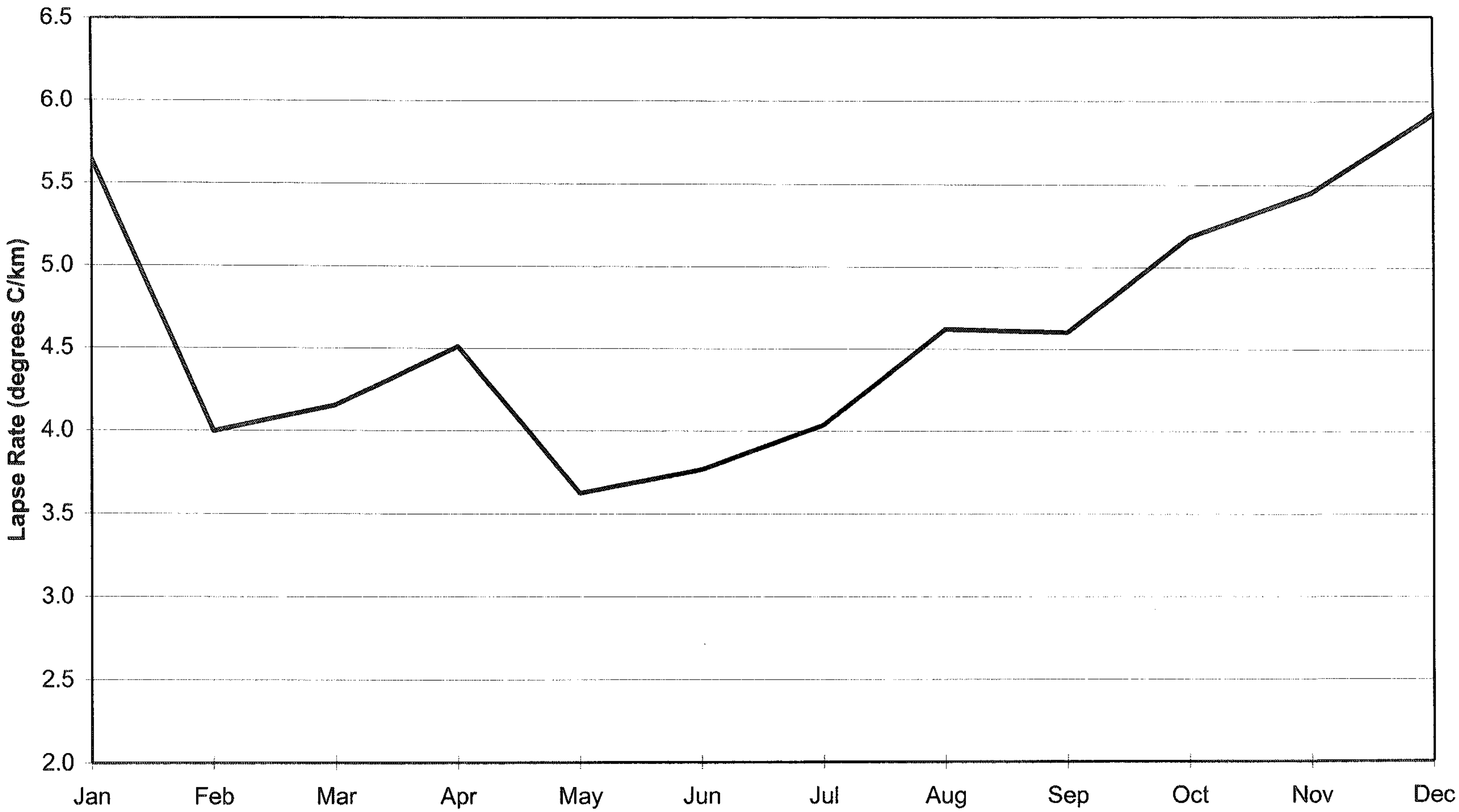


Figure 3.3 **Mean monthly western Newfoundland free air lapse rate between 60 m and 780 m.** Data from Stephenville A (1957-1999), 15-18 years missing per month.

of 830 NST + 2030 NST, for every climatological day. The mean annual value of 4.6°C/km is slightly higher than the value given by Banfield (1983) (4°C/km). However, in this case the lapse rates were calculated for the first 780 m and not the first kilometer. The differences could be partially compensated by a reduced lapse rate from 780 m to 1 km, but this topic is beyond the scope of this thesis.

The highest average monthly lapse rate values occur in fall and early winter (Oct. to Jan.) which appears to confirm the effects the relatively warm Gulf of St. Lawrence on the lowest air layer during this period. The lapse rate value drops sharply in February, when a combination of sea ice and more stable air masses are probably introduced to the region. The highest average value occurs for December (5.9°C/km) while the lowest is for May (3.6°C/km). Lapse rates in late spring through mid-summer are reduced, which is probably the result of the waters in the Gulf being relatively cool during this period. These differing monthly lapse rates will have an impact on the BL-coastal plain temperature relationships to be derived. However, since the temperature models are constructed on a monthly basis, a portion of these seasonal effects would be inherently included in the models' equations.

### 3.3.2 The Temperature Models

A total of 124 equations estimating BL temperatures were found to be statistically significant (0.01 level) out of a possible 216 (6 equations * 3 temperature elements * 12 months) and were used in the 36 temperature models. See Table 3.1 for details concerning the statistics of the regression equations.

The column named "Rank" presents the order in which the equations were used depending on which independent variables were available. The models have on average only 3.4 equations, mainly because 21 out of the 36 models do not use the lapse rate as a predictor. Only one model uses all six possible equations (December mean temperature), while several models from June to September use only two: one with RH as a predictor and the other with DH as a predictor.

The "Equation number" column gives the order in which the regressions were carried out with an "a" indicating the one in question failed the initial Durbin-Watson test and was corrected using the generalized least squares technique. A total of 95 equations, or 77% of the total, failed the Durbin-Watson test at the 5% significance level. After the corrections were applied, all but two passed at the 5% level, while the two others passed at the 1% level. Both were equations that used DH only as a predictor, and it was necessary to retain them to permit the extension of the BL record to the beginning of the study period. Furthermore, the equations were allowed a maximum of 7% of outliers outside the two standard error limit in order to permit their use in the models. However, seven equations had slightly more than 7% of outliers but were still retained. These equations had their standard error increased to reduce the number of outliers to under 7%. In contrast, the vast majority of equations had the standard value of approximately 5% of outliers.

Another necessary clarification to Table 3.1 is the column entitled "adjusted r

Table 3.1 **Statistics of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Equation number | Independant variables | 2 std err (deg. C) | r | r squared (adjusted) | n | Durbin-W | Comment |
|---|---|---|---|---|---|---|---|---|---|---|
| 1 | Tmax | 1 | 6 | TmeanRH, TmaxRH, TmaxDH | 5.8 | 0.893 | 0.793 | 122 | 1.973 | |
| 1 | Tmax | 2 | 5 | TmaxDH, TmaxRH | 6.0 | 0.885 | 0.779 | 122 | 1.981 | |
| 1 | Tmax | 3 | 4 | TmeanRH, TmaxRH | 6.5 | 0.863 | 0.741 | 155 | 2.011 | |
| 1 | Tmax | 4 | 3a | TmaxDH | 7.8 | 0.777 | 0.601 | 116 | 1.934 | 1 std err increased from 3.85 |
| 1 | Tmin | 1 | 10a | TminRH,TminDH, MeanLR | 4.8 | 0.868 | 0.740 | 56 | 1.700 | |
| 1 | Tmin | 2 | 4a | TmeanRH, TminDH | 5.1 | 0.833 | 0.688 | 116 | 1.834 | |
| 1 | Tmin | 3 | 9 | TminRH, MeanLR | 5.4 | 0.865 | 0.740 | 62 | 1.867 | |
| 1 | Tmin | 4 | 3a | TminRH, TmeanRH | 5.7 | 0.801 | 0.636 | 150 | 1.852 | |
| 1 | Tmin | 5 | 7a | TminDH | 5.8 | 0.735 | 0.537 | 116 | 2.006 | |
| 1 | Tmean | 1 | 2a | TmeanRH, TmeanDH, MaxLR | 3.7 | 0.960 | 0.917 | 56 | 2.038 | 1 std err increased from 1.8 |
| 1 | Tmean | 2 | 7 | TmeanRH, TmeanDH | 3.9 | 0.938 | 0.879 | 122 | 2.246 | |
| 1 | Tmean | 3 | 4a | TmeanRH | 4.3 | 0.933 | 0.870 | 150 | 1.983 | |
| 1 | Tmean | 4 | 5a | TmeanDH | 6.0 | 0.811 | 0.654 | 116 | 1.988 | |
| 2 | Tmax | 1 | 9a | TmeanRH, TmaxDH, MeanLR | 4.7 | 0.965 | 0.921 | 27 | 1.972 | |
| 2 | Tmax | 2 | 10 | TmaxRH, TmeanRH, TmaxDH | 5.0 | 0.937 | 0.875 | 110 | 1.861 | |
| 2 | Tmax | 3 | 6 | TmaxDH, meanLR | 5.3 | 0.933 | 0.86 | 28 | 2.119 | |
| 2 | Tmax | 4 | 5 | TmaxDH | 5.8 | 0.912 | 0.83 | 110 | 1.845 | |
| 2 | Tmax | 5 | 4a | TmaxRH, TmeanRH | 5.9 | 0.897 | 0.801 | 127 | 1.933 | |
| 2 | Tmin | 1 | 7a | TmeanRH, TminDH | 4.1 | 0.922 | 0.847 | 105 | 1.769 | |
| 2 | Tmin | 2 | 5a | TminDH, MeanLR | 3.9 | 0.927 | 0.848 | 27 | 1.952 | |
| 2 | Tmin | 3 | 4a | TminDH | 4.8 | 0.846 | 0.713 | 104 | 1.900 | |
| 2 | Tmin | 4 | 2a | TmeanRH | 4.9 | 0.906 | 0.819 | 127 | 1.967 | |
| 2 | Tmean | 1 | 6 | TmeanRH, TmaxDH | 3.6 | 0.962 | 0.924 | 110 | 1.938 | |
| 2 | Tmean | 2 | 8a | TmeanRH, MeanLR | 4.0 | 0.966 | 0.930 | 45 | 1.992 | |
| 2 | Tmean | 3 | 3 | TmeanRH | 4.4 | 0.942 | 0.887 | 133 | 1.888 | |
| 2 | Tmean | 4 | 2a | TmaxDH, TminDH | 4.5 | 0.907 | 0.819 | 104 | 1.908 | |
| 3 | Tmax | 1 | 11a | TmaxRH, TmaxDH | 5.4 | 0.885 | 0.78 | 108 | 2.022 | |
| 3 | Tmax | 2 | 9 | TmeanRH, MeanLR | 5.7 | 0.909 | 0.82 | 58 | 1.834 | 1 std err increased from 2.615 |
| 3 | Tmax | 3 | 5a | TmaxRH, TmeanRH | 5.9 | 0.861 | 0.738 | 165 | 1.965 | |
| 3 | Tmax | 4 | 2a | TmaxDH | 6.2 | 0.823 | 0.675 | 108 | 2.042 | |

n = number of cases used to build the equation

An equation number followed by an "a" indicates that the initial regression failed the Durbin-Watson test.

Table 3.1 (Cont.) **Statistics of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Equation number | Independant variables | 2 std err (deg. C) | r | r squared (adjusted) | n | Durbin-W | Comment |
|---|---|---|---|---|---|---|---|---|---|---|
| 3 | Tmin | 1 | 4a | TmeanRH, TminDH | 3.5 | 0.944 | 0.889 | 110 | 2.124 | |
| 3 | Tmin | 2 | 2a | TmeanRH | 4.5 | 0.900 | 0.808 | 165 | 1.940 | |
| 3 | Tmin | 3 | 3a | TminDH, TmeanDH | 4.7 | 0.884 | 0.777 | 109 | 2.214 | |
| 3 | Tmean | 1 | 5a | TmeanRH, MinLR | 3.4 | 0.946 | 0.891 | 57 | 2.049 | |
| 3 | Tmean | 2 | 4 | TmeanRH, TmeanDH | 3.6 | 0.951 | 0.902 | 117 | 1.803 | |
| 3 | Tmean | 3 | 2 | TmeanRH | 3.9 | 0.943 | 0.888 | 173 | 1.787 | |
| 3 | Tmean | 4 | 3a | TmaxDH, TminDH | 4.7 | 0.885 | 0.780 | 108 | 2.146 | |
| 4 | Tmax | 1 | 8 | TmaxRH, MeanLR | 5.3 | 0.767 | 0.574 | 59 | 1.959 | |
| 4 | Tmax | 2 | 3a | TmaxRH, TmeanRH | 5.6 | 0.683 | 0.459 | 143 | 1.999 | |
| 4 | Tmax | 3 | 2a | TmaxDH | 6.1 | 0.609 | 0.363 | 83 | 2.033 | |
| 4 | Tmin | 1 | 6a | TminDH, TmeanRH | 3.0 | 0.862 | 0.737 | 85 | 1.711 | |
| 4 | Tmin | 2 | 8a | TminDH, MinLR | 2.9 | 0.874 | 0.746 | 29 | 2.043 | |
| 4 | Tmin | 3 | 7a | TminRH, TmeanRH, MinLR | 3.4 | 0.825 | 0.663 | 56 | 1.978 | |
| 4 | Tmin | 4 | 2a | TminRH, TmeanRH | 3.7 | 0.773 | 0.592 | 143 | 1.765 | |
| 4 | Tmin | 5 | 3a | TminDH | 4.0 | 0.82 | 0.669 | 85 | 1.740 | 1 std err increased from 1.740 |
| 4 | Tmean | 1 | 8 | TmeanRH, TminDH, MeanLR | 2.2 | 0.939 | 0.868 | 30 | 1.884 | |
| 4 | Tmean | 2 | 4 | TmeanRH, MeanLR | 2.7 | 0.906 | 0.815 | 60 | 1.823 | |
| 4 | Tmean | 3 | 7a | TmeanRH,TMinDH | 2.9 | 0.814 | 0.655 | 85 | 1.771 | |
| 4 | Tmean | 4 | 3a | TmeanRH | 3.0 | 0.800 | 0.637 | 145 | 1.882 | |
| 4 | Tmean | 5 | 2a | TmaxDH, TminDH | 3.1 | 0.769 | 0.581 | 83 | 1.850 | |
| 5 | Tmax | 1 | 2a | TmaxRH | 5.5 | 0.734 | 0.536 | 150 | 1.934 | |
| 5 | Tmax | 2 | 3 | TmeanDH | 6.4 | 0.693 | 0.474 | 84 | 1.685 | |
| 5 | Tmin | 1 | 10a | TmeanRH,TminDH | 3.0 | 0.826 | 0.674 | 81 | 1.961 | |
| 5 | Tmin | 2 | 2a | TminDH, TmeanDH | 3.1 | 0.833 | 0.686 | 79 | 1.831 | |
| 5 | Tmin | 3 | 4a | TmeanRH, Mean LR | 3.2 | 0.819 | 0.659 | 60 | 1.930 | |
| 5 | Tmin | 4 | 3a | TmeanRH | 3.4 | 0.755 | 0.567 | 150 | 1.850 | |
| 5 | Tmean | 1 | 6a | TminRH, TmeanRH, TmeanDH | 3.3 | 0.861 | 0.730 | 79 | 1.974 | |
| 5 | Tmean | 2 | 4a | TminRH, TmeanRH | 3.5 | 0.806 | 0.645 | 150 | 1.890 | |
| 5 | Tmean | 3 | 2a | TmeanDH | 3.8 | 0.803 | 0.640 | 79 | 1.836 | |
| 6 | Tmax | 1 | 2a | TmaxRH, TMeanRH | 5.3 | 0.751 | 0.556 | 113 | 1.777 | |
| 6 | Tmax | 2 | 3a | TmeanDH | 6.2 | 0.609 | 0.363 | 75 | 1.818 | |

Table 3.1 (Cont.) **Statistics of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Equation number | Independant variables | 2 std err (deg. C) | r | r squared (adjusted) | n | Durbin-W | Comment |
|---|---|---|---|---|---|---|---|---|---|---|
| 6 | Tmin | 1 | 4a | TmeanRH, TminDH | 4.2 | 0.802 | 0.633 | 75 | 2.106 | |
| 6 | Tmin | 2 | 3a | TminDH | 4.4 | 0.767 | 0.583 | 75 | 2.173 | |
| 6 | Tmin | 3 | 2a | TmeanRH | 5.0 | 0.722 | 0.517 | 116 | 1.957 | |
| 6 | Tmean | 1 | 2a | TmeanRH, TmaxRH | 4.1 | 0.812 | 0.654 | 122 | 1.845 | |
| 6 | Tmean | 2 | 3a | TmeanDH | 4.5 | 0.735 | 0.533 | 68 | 2.135 | |
| 7 | Tmax | 1 | 3a | TmaxRH | 4.1 | 0.663 | 0.434 | 90 | 1.902 | |
| 7 | Tmax | 2 | 4a | TmaxDH | 4.9 | 0.507 | 0.248 | 88 | 1.851 | |
| 7 | Tmin | 1 | 2a | TmeanRH, TmeanDH | 3.8 | 0.746 | 0.546 | 88 | 2.039 | |
| 7 | Tmin | 2 | 3a | TmeanRH | 4.0 | 0.703 | 0.488 | 90 | 2.021 | |
| 7 | Tmin | 3 | 4 | TmeanDH | 4.5 | 0.739 | 0.542 | 92 | 1.823 | |
| 7 | Tmean | 1 | 6a | TmeanRH, TmaxDH | 2.6 | 0.840 | 0.696 | 60 | 2.088 | |
| 7 | Tmean | 2 | 3 | TmaxDH, TminDH | 2.9 | 0.891 | 0.787 | 62 | 1.732 | |
| 7 | Tmean | 3 | 4a | TmeanRH | 3.0 | 0.792 | 0.621 | 60 | 2.069 | |
| 8 | Tmax | 1 | 4a | TmaxRH, TMeanRH, TmaxDH | 3.2 | 0.887 | 0.778 | 84 | 1.917 | |
| 8 | Tmax | 2 | 2a | TmaxRH, TmeanRH | 3.5 | 0.855 | 0.724 | 84 | 1.940 | |
| 8 | Tmax | 3 | 3a | TmaxDH | 3.8 | 0.824 | 0.674 | 84 | 1.842 | |
| 8 | Tmin | 1 | 4a | TmeanRH, TminDH | 2.6 | 0.867 | 0.745 | 84 | 2.005 | |
| 8 | Tmin | 2 | 2a | TminDH, TmeanDH | 2.9 | 0.825 | 0.672 | 84 | 2.073 | |
| 8 | Tmin | 3 | 3a | TmeanRH | 3.3 | 0.789 | 0.618 | 84 | 1.880 | |
| 8 | Tmean | 1 | 4a | TmaxRH, TmeanRH, TmaxDH, TminDH | 2.3 | 0.903 | 0.806 | 84 | 1.808 | |
| 8 | Tmean | 2 | 2a | TmaxRH, TmeanRH | 2.6 | 0.875 | 0.76 | 84 | 1.813 | |
| 8 | Tmean | 3 | 3a | TmaxDH, TminDH | 2.8 | 0.846 | 0.709 | 84 | 1.784 | |
| 9 | Tmax | 1 | 9a | TmaxRH, TmaxDH | 4.8 | 0.746 | 0.549 | 109 | 1.989 | |
| 9 | Tmax | 2 | 2a | TmaxRH, TmeanRH | 5.0 | 0.719 | 0.508 | 111 | 1.936 | |
| 9 | Tmax | 3 | 5a | TmaxDH | 5.1 | 0.700 | 0.485 | 109 | 1.962 | |
| 9 | Tmin | 1 | 4a | TmeanRH, TmeanDH | 3.5 | 0.789 | 0.615 | 109 | 1.873 | |
| 9 | Tmin | 2 | 2a | TmeanRH | 3.6 | 0.767 | 0.585 | 111 | 1.898 | |
| 9 | Tmin | 3 | 3a | TmeanDH | 3.9 | 0.715 | 0.507 | 109 | 1.844 | |
| 9 | Tmean | 1 | 6a | TminRH, TmeanRH, TmaxDH | 3.3 | 0.837 | 0.692 | 109 | 1.986 | |
| 9 | Tmean | 2 | 7a | TminRH, TmeanRH | 3.6 | 0.805 | 0.641 | 111 | 1.970 | |
| 9 | Tmean | 3 | 3a | TmaxDH, TmeanDH | 3.8 | 0.781 | 0.603 | 109 | 1.954 | |

Table 3.1 (Cont.) **Statistics of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Equation number | Independant variables | 2 std err (deg. C) | r | r squared (adjusted) | n | Durbin-W | Comment |
|---|---|---|---|---|---|---|---|---|---|---|
| 10 | Tmax | 1 | 4a | TmaxRH, TmaxDH | 4.5 | 0.842 | 0.703 | 116 | 1.962 | |
| 10 | Tmax | 2 | 2a | TmaxDH | 4.7 | 0.821 | 0.67 | 116 | 2.035 | |
| 10 | Tmax | 3 | 3a | TmaxRH | 5.4 | 0.759 | 0.573 | 137 | 1.924 | |
| 10 | Tmin | 1 | 4a | TmeanRH, TminDH | 2.8 | 0.839 | 0.698 | 116 | 1.836 | |
| 10 | Tmin | 2 | 3a | TminDH, TmeanDH | 3.0 | 0.814 | 0.657 | 116 | 1.901 | |
| 10 | Tmin | 3 | 2a | TmeanRH | 3.2 | 0.821 | 0.672 | 137 | 1.950 | |
| 10 | Tmean | 1 | 5a | TmaxRH, TmeanDH | 2.7 | 0.891 | 0.79 | 116 | 1.919 | |
| 10 | Tmean | 2 | 6a | TmeanDH, Max LR | 2.7 | 0.902 | 0.8 | 30 | 2.051 | |
| 10 | Tmean | 3 | 2a | TminDH, TmeanDH | 3.0 | 0.864 | 0.742 | 116 | 2.016 | |
| 10 | Tmean | 4 | 3a | TminRH, TmeanRH | 3.3 | 0.866 | 0.745 | 137 | 1.931 | |
| 11 | Tmax | 1 | 7a | TmaxRH, TmaxDH | 4.7 | 0.869 | 0.75 | 98 | 2.001 | |
| 11 | Tmax | 2 | 4a | TmaxRH, TmeanRH | 5.4 | 0.811 | 0.652 | 127 | 1.998 | |
| 11 | Tmax | 3 | 5 | TmaxDH, TmeanDH | 5.6 | 0.832 | 0.687 | 101 | 1.726 | |
| 11 | Tmin | 1 | 7 | TmeanRH, TminDH | 2.9 | 0.950 | 0.9 | 101 | 1.747 | |
| 11 | Tmin | 2 | 5a | TminDH, TmeanDH | 3.4 | 0.893 | 0.792 | 91 | 2.396 | passed Dubin-W at 1% |
| 11 | Tmin | 3 | 3 | TminRH, TmeanRH | 3.7 | 0.917 | 0.838 | 132 | 1.845 | |
| 11 | Tmean | 1 | 6 | TmeanRH, Tmean DH | 2.7 | 0.970 | 0.939 | 101 | 1.927 | 1 std err increased from 1.089 |
| 11 | Tmean | 2 | 2 | TmeanRH | 2.9 | 0.944 | 0.89 | 132 | 1.847 | |
| 11 | Tmean | 3 | 4 | TmeanDH | 3.8 | 0.905 | 0.817 | 101 | 1.600 | passed Durbin-W at 1% |
| 12 | Tmax | 1 | 5 | TmaxRH, TmeanRH, Max LR | 4.6 | 0.904 | 0.807 | 61 | 1.914 | |
| 12 | Tmax | 2 | 6 | TmaxRH, Tmean RH, Tmax DH | 4.7 | 0.898 | 0.801 | 120 | 1.853 | |
| 12 | Tmax | 3 | 4 | TmaxRH, TmeanRH | 4.8 | 0.892 | 0.792 | 137 | 1.845 | |
| 12 | Tmax | 4 | 2a | TmeanDH | 6.7 | 0.637 | 0.401 | 112 | 1.933 | |
| 12 | Tmin | 1 | 6a | TmeanRH, TminDH | 3.7 | 0.901 | 0.809 | 112 | 1.848 | |
| 12 | Tmin | 2 | 4a | TminRH, TmeanRH | 4.3 | 0.881 | 0.772 | 131 | 1.762 | |
| 12 | Tmin | 3 | 3a | TminDH, MeanLR | 4.5 | 0.797 | 0.62 | 52 | 1.990 | 1 std err increased from 2.110 |
| 12 | Tmin | 4 | 2a | TminDH | 4.7 | 0.712 | 0.503 | 112 | 1.946 | |
| 12 | Tmean | 1 | 7a | TmeanRH, TminDH, MeanLR | 2.9 | 0.947 | 0.89 | 52 | 2.000 | |
| 12 | Tmean | 2 | 5 | TmeanRH, MeanLR | 3.3 | 0.944 | 0.887 | 62 | 1.797 | |
| 12 | Tmean | 3 | 6a | TmeanRH, TmeanDH | 3.4 | 0.926 | 0.855 | 112 | 1.882 | |
| 12 | Tmean | 4 | 4 | TmeanRH | 3.6 | 0.931 | 0.867 | 137 | 1.794 | |
| 12 | Tmean | 5 | 3a | TmeanDH, MeanLR | 4.6 | 0.820 | 0.659 | 52 | 1.715 | 1 std err increased from 2.155 |
| 12 | Tmean | 6 | 2a | TmeanDH | 5.0 | 0.711 | 0.5 | 112 | 1.878 | |

squared". As with the regular r squared value, it points out the percentage of the variability of the estimated parameter explained by the equation. However, the adjusted value is a more conservative one, taking into account the number of independent variables in the equation and the number of cases used to develop the model. The median adjusted r squared value is 0.709, with a maximum of 0.939 and a minimum of 0.248.

In all, 37 equations were produced for estimating daily maximum temperature, 43 for minimum temperature and 44 for mean temperature. Maximum daily temperature estimations produced the largest uncertainties, with a mean uncertainty (2 SE) of ± 5.3°C, and daily mean temperatures estimations yielded the smallest errors, ± 3.5°C. Minimum temperature estimations were slightly less precise at ± 3.9°C. However, uncertainties for larger periods are smaller, as discussed in the methodology of this section. For example, the average daily uncertainty for the estimation of daily maximum temperature in January of 1947 is ± 7.8°C, because equation ranked number four was used. In contrast, when the average uncertainty is divided by the square root of 31 (days), the uncertainty of the mean daily maximum temperature for the month of January 1947 falls to only ± 1.4°C.

The equations and their coefficients are presented in Table 3.2. Each column from "constant" to "Mean LR" represent terms of the equations. The constant is the y axis intercept while in each following column are located the coefficients to be multiplied by the individual daily temperature or lapse rate values (LR) in question. For example, the first row of the table would be inserted in the model under the following form: -3.147 + 0.374*(Tmax RH) + 0.357*(Tmean RH) + 0.350*(Tmax DH). This equation returns the estimated BL

Table 3.2 **Coefficients of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Independant variables | Constant | RH max | RH min | RH mean | DH max | DH min | DH mean | Max LR | Min LR | Mean LR |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | Tmax | 1 | TmeanRH, TmaxRH, TmaxDH | -3.147 | 0.374 | n/a | 0.357 | 0.350 | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmax | 2 | TmaxDH, TmaxRH | -4.813 | 0.598 | n/a | n/a | 0.399 | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmax | 3 | TmeanRH, TmaxRH | -2.763 | 0.513 | n/a | 0.531 | n/a | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmax | 4 | TmaxDH | -5.296 | n/a | n/a | n/a | 0.803 | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmin | 1 | TminRH,TminDH, MeanLR | -2.367 | n/a | 0.442 | n/a | n/a | 0.382 | n/a | n/a | n/a | -0.618 |
| 1 | Tmin | 2 | TmeanRH, TminDH | -7.346 | n/a | n/a | 0.528 | n/a | 0.372 | n/a | n/a | n/a | n/a |
| 1 | Tmin | 3 | TminRH, MeanLR | -3.250 | n/a | 0.773 | n/a | n/a | n/a | n/a | n/a | n/a | -0.717 |
| 1 | Tmin | 4 | TminRH, TmeanRH | -9.251 | n/a | 0.189 | 0.63 | n/a | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmin | 5 | TminDH | -6.707 | n/a | n/a | n/a | n/a | 0.71 | n/a | n/a | n/a | n/a |
| 1 | Tmean | 1 | TmeanRH, TmeanDH, MaxLR | -2.753 | n/a | n/a | 0.790 | n/a | n/a | 0.244 | -0.281 | n/a | n/a |
| 1 | Tmean | 2 | TmeanRH, TmeanDH | -4.081 | n/a | n/a | 0.811 | n/a | n/a | 0.289 | n/a | n/a | n/a |
| 1 | Tmean | 3 | TmeanRH | -4.641 | n/a | n/a | 1.030 | n/a | n/a | n/a | n/a | n/a | n/a |
| 1 | Tmean | 4 | TmeanDH | -4.670 | n/a | n/a | n/a | n/a | n/a | 0.896 | n/a | n/a | n/a |
| 2 | Tmax | 1 | TmeanRH, TmaxDH, MeanLR | -0.954 | n/a | n/a | 0.370 | 0.615 | n/a | n/a | n/a | n/a | -0.465 |
| 2 | Tmax | 2 | TmaxRH, TmeanRH, TmaxDH | -2.822 | 0.290 | n/a | 0.273 | 0.544 | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmax | 3 | TmaxDH, meanLR | -2.668 | n/a | n/a | n/a | 0.904 | n/a | n/a | n/a | n/a | -0.437 |
| 2 | Tmax | 4 | TmaxDH | -3.842 | n/a | n/a | n/a | 1.001 | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmax | 5 | TmaxRH, TmeanRH | -2.541 | 0.594 | n/a | 0.491 | N/a | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmin | 1 | TmeanRH, TminDH | -6.503 | n/a | n/a | 0.524 | n/a | 0.422 | n/a | n/a | n/a | n/a |
| 2 | Tmin | 2 | TminDH, MeanLR | -3.246 | n/a | n/a | n/a | n/a | 0.828 | n/a | n/a | n/a | -0.494 |
| 2 | Tmin | 3 | TminDH | -6.014 | n/a | n/a | n/a | n/a | 0.776 | n/a | n/a | n/a | n/a |
| 2 | Tmin | 4 | TmeanRH | -9.009 | n/a | n/a | 0.950 | n/a | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmean | 1 | TmeanRH, TmaxDH | -5.461 | n/a | n/a | 0.696 | 0.365 | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmean | 2 | TmeanRH, MeanLR | -3.691 | n/a | n/a | 0.979 | n/a | n/a | n/a | n/a | n/a | -0.321 |
| 2 | Tmean | 3 | TmeanRH | -4.497 | n/a | n/a | 1.033 | n/a | n/a | n/a | n/a | n/a | n/a |
| 2 | Tmean | 4 | TmaxDH, TminDH | -4.739 | n/a | n/a | n/a | 0.544 | 0.413 | n/a | n/a | n/a | n/a |
| 3 | Tmax | 1 | TmaxRH, TmaxDH | -4.305 | 0.509 | n/a | n/a | 0.590 | n/a | n/a | n/a | n/a | n/a |
| 3 | Tmax | 2 | TmeanRH, MeanLR | 1.430 | n/a | n/a | 0.714 | n/a | n/a | n/a | n/a | n/a | -0.690 |
| 3 | Tmax | 3 | TmaxRH, TmeanRH | -3.277 | 0.467 | n/a | 0.508 | n/a | n/a | n/a | n/a | n/a | n/a |
| 3 | Tmax | 4 | TmaxDH | -3.676 | n/a | n/a | n/a | 0.984 | n/a | n/a | n/a | n/a | n/a |

n/a = not applicable

Table 3.2 (Cont.) **Coefficients of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Independant variables | constant | RH max | RH min | RH mean | DH max | DH min | DH mean | Max LR | Min LR | Mean LR |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 3 | Tmin | 1 | TmeanRH, TminDH | -6.091 | n/a | n/a | 0.591 | n/a | 0.432 | n/a | n/a | n/a | n/a |
| 3 | Tmin | 2 | TmeanRH | -9.182 | n/a | n/a | 0.936 | n/a | n/a | n/a | n/a | n/a | n/a |
| 3 | Tmin | 3 | TminDH, TmeanDH | -5.993 | n/a | n/a | n/a | n/a | 0.422 | 0.553 | n/a | n/a | n/a |
| 3 | Tmean | 1 | TmeanRH, MinLR | -4.039 | n/a | n/a | 0.847 | n/a | n/a | n/a | n/a | -0.300 | n/a |
| 3 | Tmean | 2 | TmeanRH, TmeanDH | -4.295 | n/a | n/a | 0.678 | n/a | n/a | 0.327 | n/a | n/a | n/a |
| 3 | Tmean | 3 | TmeanRH | -5.291 | n/a | n/a | 0.944 | n/a | n/a | n/a | n/a | n/a | n/a |
| 3 | Tmean | 4 | TmaxDH, TminDH | -4.232 | n/a | n/a | n/a | 0.578 | 0.430 | n/a | n/a | n/a | n/a |
| 4 | Tmax | 1 | TmaxRH, MeanLR | -0.335 | 0.588 | n/a | n/a | n/a | n/a | n/a | n/a | n/a | -0.631 |
| 4 | Tmax | 2 | TmaxRH, TmeanRH | -3.601 | 0.457 | n/a | 0.406 | n/a | n/a | n/a | n/a | n/a | n/a |
| 4 | Tmax | 3 | TmaxDH | -2.689 | n/a | n/a | n/a | 0.685 | n/a | n/a | n/a | n/a | n/a |
| 4 | Tmin | 1 | TminDH, TmeanRH | -5.917 | n/a | n/a | 0.386 | n/a | 0.656 | n/a | n/a | n/a | n/a |
| 4 | Tmin | 2 | TminDH, MinLR | -2.713 | n/a | n/a | n/a | n/a | 0.903 | n/a | n/a | -0.373 | n/a |
| 4 | Tmin | 3 | TminRH, TmeanRH, MinLR | -4.269 | n/a | 0.365 | 0.382 | n/a | n/a | n/a | n/a | -0.377 | n/a |
| 4 | Tmin | 4 | TminRH, TmeanRH | -7.561 | n/a | 0.202 | 0.624 | n/a | n/a | n/a | n/a | n/a | n/a |
| 4 | Tmin | 5 | TminDH | -4.519 | n/a | n/a | n/a | n/a | 0.984 | n/a | n/a | n/a | n/a |
| 4 | Tmean | 1 | TmeanRH, TminDH, MeanLR | -1.013 | n/a | n/a | 0.462 | n/a | 0.395 | n/a | n/a | n/a | -0.495 |
| 4 | Tmean | 2 | TmeanRH, MeanLR | -2.600 | n/a | n/a | 0.760 | n/a | n/a | n/a | n/a | n/a | -0.441 |
| 4 | Tmean | 3 | TmeanRH,TMinDH | -4.028 | n/a | n/a | 0.621 | n/a | 0.250 | n/a | n/a | n/a | n/a |
| 4 | Tmean | 4 | TmeanRH | -4.793 | n/a | n/a | 0.785 | n/a | n/a | n/a | n/a | n/a | n/a |
| 4 | Tmean | 5 | TmaxDH, TminDH | -3.856 | n/a | n/a | n/a | 0.355 | 0.426 | n/a | n/a | n/a | n/a |
| 5 | Tmax | 1 | TmaxRH | -3.252 | 0.676 | n/a | n/a | n/a | n/a | n/a | n/a | n/a | n/a |
| 5 | Tmax | 2 | TmeanDH | -0.910 | n/a | n/a | n/a | n/a | n/a | 1.025 | n/a | n/a | n/a |
| 5 | Tmin | 1 | TmeanRH,TminDH | -5.022 | n/a | n/a | 0.384 | n/a | 0.572 | n/a | n/a | n/a | n/a |
| 5 | Tmin | 2 | TminDH, TmeanDH | -4.809 | n/a | n/a | n/a | n/a | 0.552 | 0.453 | n/a | n/a | n/a |
| 5 | Tmin | 3 | TmeanRH, Mean LR | -5.295 | n/a | n/a | 0.670 | n/a | n/a | n/a | n/a | n/a | -0.210 |
| 5 | Tmin | 4 | TmeanRH | -5.976 | n/a | n/a | 0.695 | n/a | n/a | n/a | n/a | n/a | n/a |
| 5 | Tmean | 1 | TminRH, TmeanRH, TmeanDH | -5.298 | n/a | -0.224 | 0.663 | n/a | n/a | 0.479 | n/a | n/a | n/a |
| 5 | Tmean | 2 | TminRH, TmeanRH | -4.793 | n/a | -0.277 | 1.003 | n/a | n/a | n/a | n/a | n/a | n/a |
| 5 | Tmean | 3 | TmeanDH | -3.689 | n/a | n/a | n/a | n/a | n/a | 0.954 | n/a | n/a | n/a |
| 6 | Tmax | 1 | TmaxRH, TMeanRH | -0.950 | 0.428 | n/a | 0.457 | n/a | n/a | n/a | n/a | n/a | n/a |
| 6 | Tmax | 2 | TmeanDH | 2.550 | n/a | n/a | n/a | n/a | n/a | 0.950 | n/a | n/a | n/a |

Table 3.2 (Cont.) **Coefficients of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Independant variables | constant | RH max | RH min | RH mean | DH max | DH min | DH mean | Max LR | Min LR | Mean LR |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 6 | Tmin | 1 | TmeanRH, TminDH | -4.837 | n/a | n/a | 0.355 | n/a | 0.799 | n/a | n/a | n/a | n/a |
| 6 | Tmin | 2 | TminDH | -2.251 | n/a | n/a | n/a | n/a | 1.079 | n/a | n/a | n/a | n/a |
| 6 | Tmin | 3 | TmeanRH | -5.765 | n/a | n/a | 0.860 | n/a | n/a | n/a | n/a | n/a | n/a |
| 6 | Tmean | 1 | TmeanRH, TmaxRH | -4.027 | 0.281 | n/a | 0.623 | n/a | n/a | n/a | n/a | n/a | n/a |
| 6 | Tmean | 2 | TmeanDH | -1.842 | n/a | n/a | n/a | n/a | n/a | 1.008 | n/a | n/a | n/a |
| 7 | Tmax | 1 | TmaxRH | 4.076 | 0.547 | n/a | n/a | n/a | n/a | n/a | n/a | n/a | n/a |
| 7 | Tmax | 2 | TmaxDH | 7.105 | n/a | n/a | n/a | 0.485 | n/a | n/a | n/a | n/a | n/a |
| 7 | Tmin | 1 | TmeanRH, TmeanDH | -4.438 | n/a | n/a | 0.563 | n/a | n/a | 0.305 | n/a | n/a | n/a |
| 7 | Tmin | 2 | TmeanRH | -3.220 | n/a | n/a | 0.763 | n/a | n/a | n/a | n/a | n/a | n/a |
| 7 | Tmin | 3 | TmeanDH | -3.614 | n/a | n/a | n/a | n/a | n/a | 0.865 | n/a | n/a | n/a |
| 7 | Tmean | 1 | TmeanRH, TmaxDH | -2.371 | n/a | n/a | 0.575 | 0.320 | n/a | n/a | n/a | n/a | n/a |
| 7 | Tmean | 2 | TmaxDH, TminDH | -1.974 | n/a | n/a | n/a | 0.502 | 0.491 | n/a | n/a | n/a | n/a |
| 7 | Tmean | 3 | TmeanRH | -0.025 | n/a | n/a | 0.784 | n/a | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmax | 1 | TmaxRH, TMeanRH, TmaxDH | -5.197 | 0.256 | n/a | 0.304 | 0.495 | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmax | 2 | TmaxRH, TmeanRH | -2.601 | 0.501 | n/a | 0.382 | n/a | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmax | 3 | TmaxDH | -4.034 | n/a | n/a | n/a | 0.998 | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmin | 1 | TmeanRH, TminDH | -3.252 | n/a | n/a | 0.371 | n/a | 0.450 | n/a | n/a | n/a | n/a |
| 8 | Tmin | 2 | TminDH, TmeanDH | -2.495 | n/a | n/a | n/a | n/a | 0.372 | 0.405 | n/a | n/a | n/a |
| 8 | Tmin | 3 | TmeanRH | -3.725 | n/a | n/a | 0.729 | n/a | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmean | 1 | TmaxRH, TmeanRH, TmaxDH, TminDH | -5.003 | 0.147 | n/a | 0.336 | 0.305 | 0.156 | n/a | n/a | n/a | n/a |
| 8 | Tmean | 2 | TmaxRH, TmeanRH | -3.556 | 0.319 | n/a | 0.481 | n/a | n/a | n/a | n/a | n/a | n/a |
| 8 | Tmean | 3 | TmaxDH, TminDH | -3.457 | n/a | n/a | n/a | 0.580 | 0.323 | n/a | n/a | n/a | n/a |
| 9 | Tmax | 1 | TmaxRH, TmaxDH | -4.422 | 0.431 | n/a | n/a | 0.554 | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmax | 2 | TmaxRH, TmeanRH | -3.252 | 0.617 | n/a | 0.310 | n/a | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmax | 3 | TmaxDH | -2.585 | n/a | n/a | n/a | 0.936 | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmin | 1 | TmeanRH, TmeanDH | -5.890 | n/a | n/a | 0.536 | n/a | n/a | 0.346 | n/a | n/a | n/a |
| 9 | Tmin | 2 | TmeanRH | -5.047 | n/a | n/a | 0.781 | n/a | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmin | 3 | TmeanDH | -4.698 | n/a | n/a | n/a | n/a | n/a | 0.812 | n/a | n/a | n/a |
| 9 | Tmean | 1 | TminRH, TmeanRH, TmaxDH | -6.715 | n/a | -0.187 | 0.776 | 0.428 | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmean | 2 | TminRH, TmeanRH | -5.612 | n/a | -0.397 | 1.318 | n/a | n/a | n/a | n/a | n/a | n/a |
| 9 | Tmean | 3 | TmaxDH, TmeanDH | -5.144 | n/a | n/a | n/a | 0.633 | n/a | 0.309 | n/a | n/a | n/a |

Table 3.2 (Cont.) **Coefficients of the regression equations used in the temperature models.**

| Month | BL element estimated | Rank | Independant variables | constant | RH max | RH min | RH mean | DH max | DH min | DH mean | Max LR | Min LR | Mean LR |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 10 | Tmax | 1 | TmaxRH, TmaxDH | -4.720 | 0.280 | n/a | n/a | 0.708 | n/a | n/a | n/a | n/a | n/a |
| 10 | Tmax | 2 | TmaxDH | -4.197 | n/a | n/a | n/a | 0.946 | n/a | n/a | n/a | n/a | n/a |
| 10 | Tmax | 3 | TmaxRH | -3.905 | 0.857 | n/a | n/a | n/a | n/a | n/a | n/a | n/a | n/a |
| 10 | Tmin | 1 | TmeanRH, TminDH | -4.555 | n/a | n/a | 0.430 | n/a | 0.369 | n/a | n/a | n/a | n/a |
| 10 | Tmin | 2 | TminDH, TmeanDH | -4.739 | n/a | n/a | n/a | n/a | 0.228 | 0.508 | n/a | n/a | n/a |
| 10 | Tmin | 3 | TmeanRH | -5.518 | n/a | n/a | 0.772 | n/a | n/a | n/a | n/a | n/a | n/a |
| 10 | Tmean | 1 | TmaxRH, TmeanDH | -5.284 | 0.334 | n/a | n/a | n/a | n/a | 0.589 | n/a | n/a | n/a |
| 10 | Tmean | 2 | TmeanDH, Max LR | -1.898 | n/a | n/a | n/a | n/a | n/a | 0.847 | -0.302 | n/a | n/a |
| 10 | Tmean | 3 | TminDH, TmeanDH | -4.851 | n/a | n/a | n/a | n/a | -0.405 | 1.243 | n/a | n/a | n/a |
| 10 | Tmean | 4 | TminRH, TmeanRH | -5.100 | n/a | -0.339 | 1.237 | n/a | n/a | n/a | n/a | n/a | n/a |
| 11 | Tmax | 1 | TmaxRH, TmaxDH | -5.200 | 0.493 | n/a | n/a | 0.497 | n/a | n/a | n/a | n/a | n/a |
| 11 | Tmax | 2 | TmaxRH, TmeanRH | -3.179 | 0.375 | n/a | 0.592 | n/a | n/a | n/a | n/a | n/a | n/a |
| 11 | Tmax | 3 | TmaxDH, TmeanDH | -3.646 | n/a | n/a | n/a | 0.448 | n/a | 0.469 | n/a | n/a | n/a |
| 11 | Tmin | 1 | TmeanRH, TminDH | -7.036 | n/a | n/a | 0.620 | n/a | 0.456 | n/a | n/a | n/a | n/a |
| 11 | Tmin | 2 | TminDH, TmeanDH | -6.149 | n/a | n/a | n/a | n/a | 0.624 | 0.375 | n/a | n/a | n/a |
| 11 | Tmin | 3 | TminRH, TmeanRH | -7.811 | n/a | 0.182 | 0.799 | n/a | n/a | n/a | n/a | n/a | n/a |
| 11 | Tmean | 1 | TmeanRH, Tmean DH | -5.547 | n/a | n/a | 0.640 | n/a | n/a | 0.407 | n/a | n/a | n/a |
| 11 | Tmean | 2 | TmeanRH | -5.412 | n/a | n/a | 0.977 | n/a | n/a | n/a | n/a | n/a | n/a |
| 11 | Tmean | 3 | TmeanDH | -5.425 | n/a | n/a | n/a | n/a | n/a | 0.932 | n/a | n/a | n/a |
| 12 | Tmax | 1 | TmaxRH, TmeanRH, Max LR | -1.711 | 0.347 | n/a | 0.547 | n/a | n/a | n/a | -0.323 | n/a | n/a |
| 12 | Tmax | 2 | TmaxRH, Tmean RH, Tmax DH | -3.876 | 0.412 | n/a | 0.490 | 0.134 | n/a | n/a | n/a | n/a | n/a |
| 12 | Tmax | 3 | TmaxRH, TmeanRH | -3.778 | 0.466 | n/a | 0.521 | n/a | n/a | n/a | n/a | n/a | n/a |
| 12 | Tmax | 4 | TmeanDH | -1.704 | n/a | n/a | n/a | n/a | n/a | 0.840 | n/a | n/a | n/a |
| 12 | Tmin | 1 | TmeanRH, TminDH | -6.068 | n/a | n/a | 0.652 | n/a | 0.386 | n/a | n/a | n/a | n/a |
| 12 | Tmin | 2 | TminRH, TmeanRH | -7.093 | n/a | 0.264 | 0.692 | n/a | n/a | n/a | n/a | n/a | n/a |
| 12 | Tmin | 3 | TminDH, MeanLR | -2.197 | n/a | n/a | n/a | n/a | 0.723 | n/a | n/a | n/a | -0.553 |
| 12 | Tmin | 4 | TminDH | -5.560 | n/a | n/a | n/a | n/a | 0.707 | n/a | n/a | n/a | n/a |
| 12 | Tmean | 1 | TmeanRH, TminDH, MeanLR | -0.904 | n/a | n/a | 0.622 | n/a | 0.295 | n/a | n/a | n/a | -0.528 |
| 12 | Tmean | 2 | TmeanRH, MeanLR | -3.392 | n/a | n/a | 0.901 | n/a | n/a | n/a | n/a | n/a | -0.341 |
| 12 | Tmean | 3 | TmeanRH, TmeanDH | -4.774 | n/a | n/a | 0.836 | n/a | n/a | 0.213 | n/a | n/a | n/a |
| 12 | Tmean | 4 | TmeanRH | -5.159 | n/a | n/a | 1.000 | n/a | n/a | n/a | n/a | n/a | n/a |
| 12 | Tmean | 5 | TmeanDH, MeanLR | -1.077 | n/a | n/a | n/a | n/a | n/a | 0.703 | n/a | n/a | -0.717 |
| 12 | Tmean | 6 | TmeanDH | -5.025 | n/a | n/a | n/a | n/a | n/a | 0.772 | n/a | n/a | n/a |

daily maximum temperature for January in degrees C with the associated uncertainty.

### 3.3.3 Big Level Historical Temperature Record

The monthly temperature means for 1946-1999 were compiled from the estimated record of daily maximum, minimum and mean temperatures for the Big Level Summit Plateau (BL). Table 3.3 and Figure 3.4 show the results. The plateau has an estimated mean annual temperature of -0.7°C for the period between November 1946 and June 1999, which is 3.8°C colder on average than at DH. At BL the months of November through April have a mean daily maximum temperature lower than 0°C, the period defined earlier as "winter", which is two months longer than on the coastal plain. In contrast, only two months of the year have a mean temperature above 10°C on BL, namely July and August. Distinguishable from the coastal plain, the warmest month on BL is clearly July, exhibiting the more continental position of the plateau due to its horizontal and vertical distance from the Gulf of St. Lawrence. The months of June and September are also relatively warm on BL, with the mean daily minimum temperature being above 0°C and the daily maximum above 10°C, which is not the case from October through May. Finally, the months of May and October are shoulder seasons, characterized by frequent freeze-thaw cycles, with their mean maximum and minimum temperatures above and below 0°C respectively.

The monthly mean temperature differences between DH and BL, which were

Table 3.3 **Big Level temperature and thawing degree-day monthly mean values** Period from November 1946 to June 1999

| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Daily Maximum Temperature (Tmax** | | | | | | | | | | | | | |
| Mean | -8.1 | -8.5 | -4.6 | -0.1 | 4.3 | 11.6 | 15.2 | 13.4 | 10.3 | 4.3 | -0.8 | -5.3 | 2.6 |
| Maximum monthly | -3.2 | -2.2 | 0.3 | 3.5 | 9.0 | 15.7 | 16.8 | 16.0 | 12.2 | 6.4 | 1.8 | -0.5 | |
| Minimum monthly | -14.0 | -14.2 | -10.1 | -2.5 | 0.9 | 8.8 | 12.8 | 11.0 | 7.3 | 1.8 | -4.2 | -11.3 | |
| Median | -7.9 | -9.1 | -4.2 | -0.4 | 4.2 | 11.5 | 15.3 | 13.2 | 10.2 | 4.2 | -0.9 | -5.3 | |
| Standard deviation | 2.2 | 3.1 | 2.3 | 1.5 | 1.6 | 1.2 | 0.8 | 1.1 | 1.0 | 1.1 | 1.5 | 2.0 | |
| Number of years available | 51 | 51 | 53 | 53 | 51 | 53 | 49 | 48 | 49 | 52 | 49 | 50 | |
| **Daily Minimum Temperature (Tmin** | | | | | | | | | | | | | |
| Mean | -14.9 | -16.1 | -12.3 | -7.2 | -1.6 | 4.3 | 8.7 | 7.7 | 4.1 | -1.0 | -6.6 | -11.0 | -3.8 |
| Maximum monthly | -8.2 | -9.6 | -6.9 | -2.7 | 1.7 | 6.9 | 10.6 | 10.1 | 5.6 | 0.7 | -3.7 | -6.5 | |
| Minimum monthly | -21.5 | -23.6 | -18.8 | -11.2 | -5.2 | 1.3 | 5.6 | 5.6 | 1.6 | -2.8 | -10.3 | -18.1 | |
| Median | -14.8 | -16.3 | -11.9 | -7.5 | -1.8 | 4.2 | 8.7 | 7.6 | 4.1 | -1.0 | -6.4 | -10.5 | |
| Standard deviation | 2.6 | 3.5 | 3.0 | 1.9 | 1.4 | 1.2 | 1.1 | 0.9 | 0.8 | 0.8 | 1.5 | 2.1 | |
| Number of years available | 51 | 51 | 53 | 53 | 51 | 53 | 49 | 48 | 49 | 51 | 49 | 51 | |
| **Daily Mean Temperature (Tmean** | | | | | | | | | | | | | |
| Mean | -11.5 | -12.5 | -8.6 | -3.7 | 1.2 | 7.8 | 12.2 | 10.4 | 7.0 | 1.4 | -3.9 | -8.3 | -0.7 |
| Maximum monthly | -4.5 | -6.1 | -3.7 | 0.4 | 5.3 | 11.2 | 14.3 | 13.0 | 8.8 | 3.3 | -1.4 | -4.1 | |
| Minimum monthly | -18.4 | -20.0 | -14.8 | -6.5 | -2.0 | 5.3 | 9.0 | 8.3 | 4.0 | -0.8 | -7.4 | -14.6 | |
| Median | -11.1 | -12.9 | -8.4 | -3.9 | 1.0 | 7.7 | 12.2 | 10.4 | 6.8 | 1.3 | -4.0 | -8.2 | |
| Standard deviation | 2.7 | 3.6 | 2.6 | 1.6 | 1.5 | 1.1 | 1.1 | 1.0 | 1.0 | 1.0 | 1.4 | 2.0 | |
| Number of years available | 51 | 51 | 53 | 53 | 51 | 53 | 49 | 48 | 49 | 51 | 49 | 50 | |
| **Degree-Days above 0 degree C (using Tmax** | | | | | | | | | | | | | |
| Mean | 6 | 9 | 19 | 40 | 135 | 347 | 472 | 416 | 310 | 137 | 39 | 10 | 1940 |
| Maximum monthly | 51 | 42 | 86 | 119 | 283 | 471 | 521 | 496 | 367 | 200 | 102 | 47 | |
| Minimum monthly | 0 | 0 | 0 | 2 | 56 | 264 | 397 | 340 | 220 | 69 | 1 | 0 | |
| Median | 3 | 4 | 13 | 39 | 130 | 347 | 474 | 409 | 308 | 135 | 41 | 6 | |
| Standard deviation | 9 | 12 | 17 | 30 | 46 | 36 | 26 | 35 | 31 | 30 | 23 | 13 | |
| Number of years available | 50 | 51 | 51 | 50 | 48 | 52 | 48 | 47 | 46 | 48 | 47 | 50 | |
| **Degree-Days above 0 degree C (using Tmean** | | | | | | | | | | | | | |
| Mean | 1 | 1 | 3 | 8 | 59 | 236 | 376 | 322 | 209 | 62 | 12 | 3 | 1292 |
| Maximum monthly | 23 | 15 | 19 | 65 | 177 | 337 | 444 | 403 | 264 | 104 | 44 | 27 | |
| Minimum monthly | 0 | 0 | 0 | 0 | 5 | 162 | 280 | 257 | 119 | 16 | 0 | 0 | |
| Median | 0 | 0 | 1 | 4 | 49 | 232 | 373 | 319 | 206 | 60 | 9 | 0 | |
| Standard deviation | 4 | 4 | 4 | 13 | 35 | 34 | 34 | 31 | 30 | 22 | 11 | 6 | |
| Number of years available | 50 | 51 | 51 | 50 | 48 | 52 | 47 | 47 | 45 | 48 | 47 | 50 | |

All temperatures and related indices are in degrees Celsius

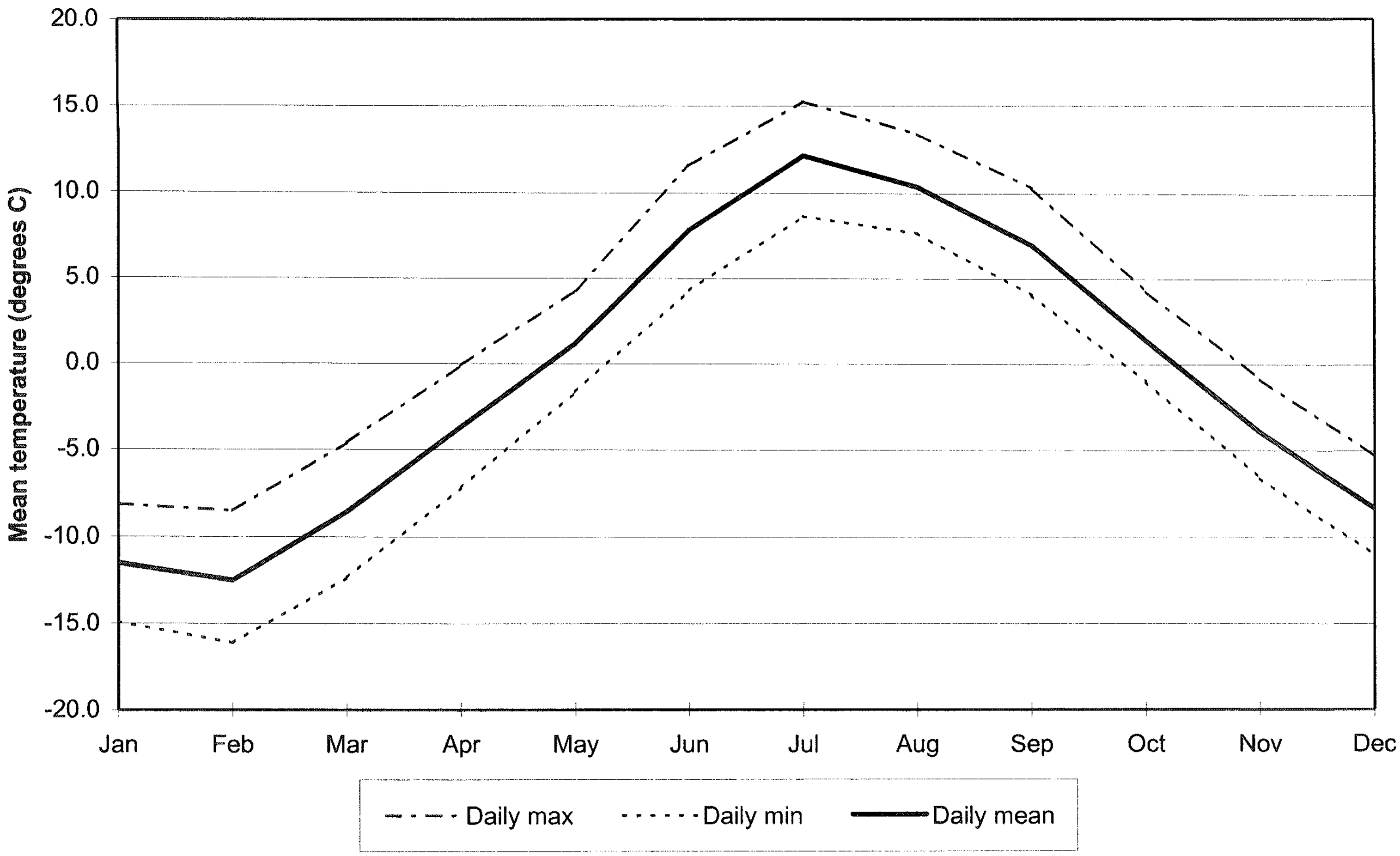


Figure 3.4 **Big Level (Summit Plateau) estimated historical monthly mean temperatures.** Period from November 1946 to June 1999 with 0-4 missing years per month.

calculated from the data presented in Figures 3.1 and 3.4, vary from 2.0°C in June and July to a maximum of 5.4°C in November and are similar to the free air lapse rates discussed in section 3.3.1. The months of August through December have larger average temperature differences owing in part to the more limited effect of the Gulf of St. Lawrence inland, while from mid-winter until mid-summer have lower values due to ice and cold waters in the Gulf. In fact, because of the more continental position and higher altitude of BL with respect to the Gulf, the plateau is less exposed to its moderating effect than DH. November and May have particularly high mean temperature difference values in the context of their seasons, 5.4°C and 3.9°C respectively, which is considered to be the result of a snowcovered (cool) plateau and a contrasting snow free coastal plain normally occurring during these months.

From this daily temperature record were also derived the 1946-1999 thawing degree-day median values presented in Table 3.3 and Figure 3.5. The data in the figure are based on daily maximum temperature; the median monthly values were used instead of the mean because most of the months with snow on the ground have degree-day values near zero. The central tendency is better represented in this case by the median, without the influence of the large values of a few abnormal years.

As would be expected from the temperature mean values, the monthly median thawing degree-day values for the months of December through March (based on Tmax) are no greater than 20. April and November also have relatively low values, their medians being under 45. May and October, which are the seasonal transition months, have approximately 135 degree-days each. It is clear from Figure 3.5 that the period of June to September

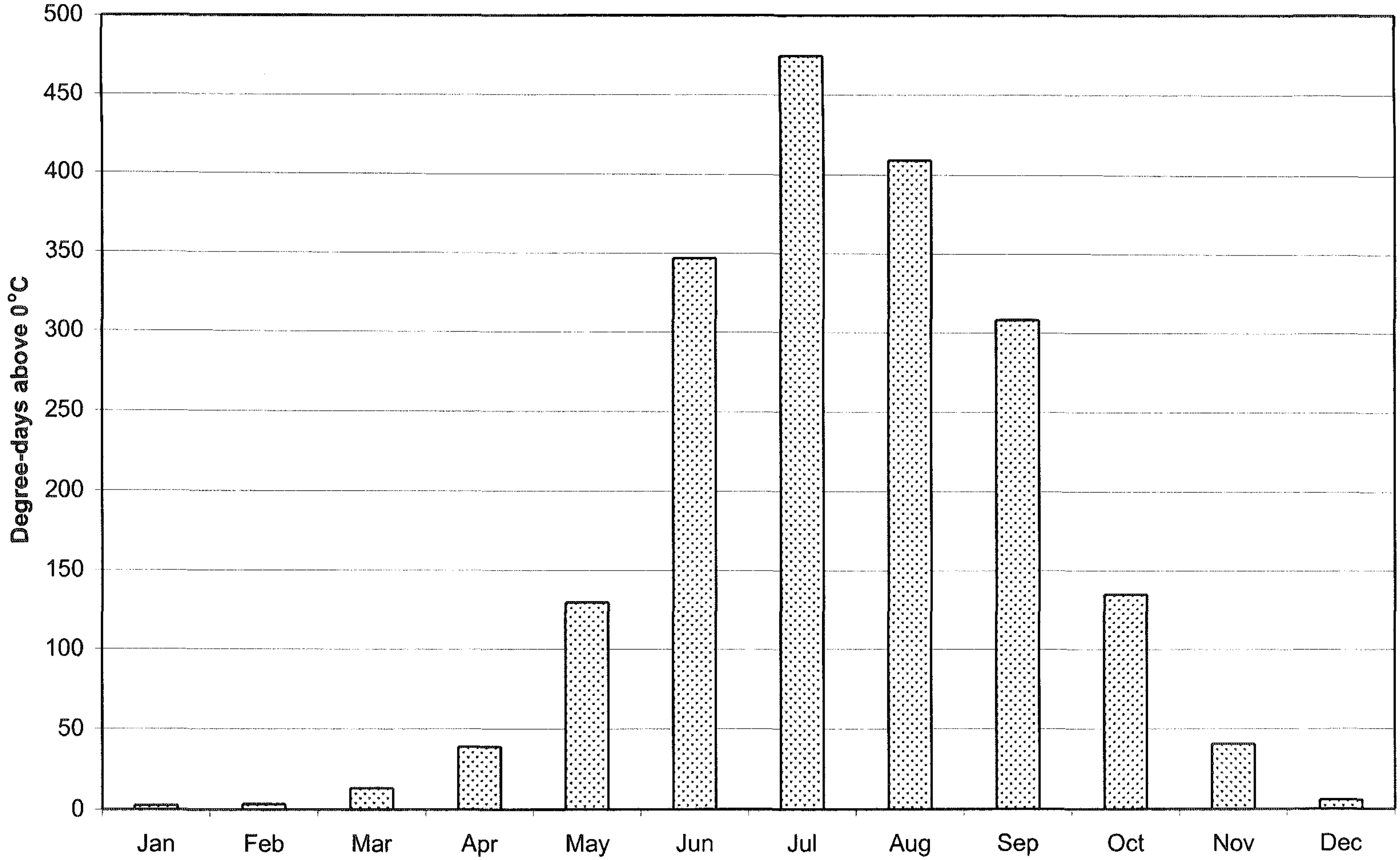


Figure 3.5 **Big Level median thawing degree-days using daily maximum temperature.** Period from November 1946 to June 1999 with 2-5 missing years per month.

constitutes the warm period, with more than 300 degree days per month.

The analysis of the monthly temperature mean values and thawing degree-day median values permits the division of the year into snow seasons for subsequent analysis of the Big Level climate. The period from June to September is clearly the "summer" season with any snow on the ground quickly melting due to the relatively high maximum temperatures and the virtual absence of frost at night from mid June to mid September. The onset of the snowcover season may be as early as mid to late October during colder autumns, with the mean temperature for this month being slightly above 0°C. The latest conceivable time for the appearance of a permanent (seasonal) snowcover on BL would be late November, with temperatures for this month normally being below freezing. December through March are without a doubt all winter months on Big Level. April would also most likely contribute to the building of the snowpack with its temperatures below freezing most of the time. Therefore, the period November through April is defined as the snow accumulation season for BL, the second period of the snow year. October will be studied separately to identify years with early snow season starts. The snow accumulation season is then followed by the snowmelt season.

The snowmelt season includes May and June. The latter month is included within the snowmelt season, despite having summer like temperatures, because May is probably not warm enough during most years to fully melt a thick snowcover. Since late lying snowbeds are specifically under study, the months of July, August and September are only marginally important in the context of this thesis, i.e. during exceptionally early or late snow seasons.

Figure 3.6 shows the estimated October mean temperatures (Tmean) for the entire period of the constructed record of BL and for the coastal plain (DH and RH). The mean October temperature for BL is 1.4 ± 0.1°C, with no significant warming or cooling trend throughout the period. October 1956 had too many missing days to include, using the criterion of 3 consecutive days or 5 total days missing (Environment Canada, 1994). Furthermore, the DH average temperature is 5.8°C, 4.4°C higher than at BL. Temperature values outside of one standard deviation around the mean were considered above or below average, being warm or cold respectively, the others considered normal. Eight Octobers are thus considered cold (1969, 1972, 1974, 1982, 1984-1986 and 1993), with four in the early to mid-1980s. These cooler months had mean temperature values close to 0°C or lower; therefore, early starts to the snowcover season, i.e. in the second half of October, could have occurred at these times. In contrast, of 10 Octobers that were warm, seven were in the 1960s and 1990s (1961, 1963, 1967, 1968, 1970, 1983, 1987, 1991, 1994 and 1995). Moreover, no such warm Octobers were encountered prior to 1961, only one occurred in the 1970s and only two in the 1980s. These autumns would have not seen early starts to the snow season.

The snow accumulation season defined as the period from November to April occupies half the year. Figure 3.7 shows the estimated mean temperature for each season from 1947 to 1999, with November and December of each accumulation season being part of the previous calendar year. This must be taken in consideration when linking accumulation season temperatures to October temperatures. Four years (1948, 1949, 1950

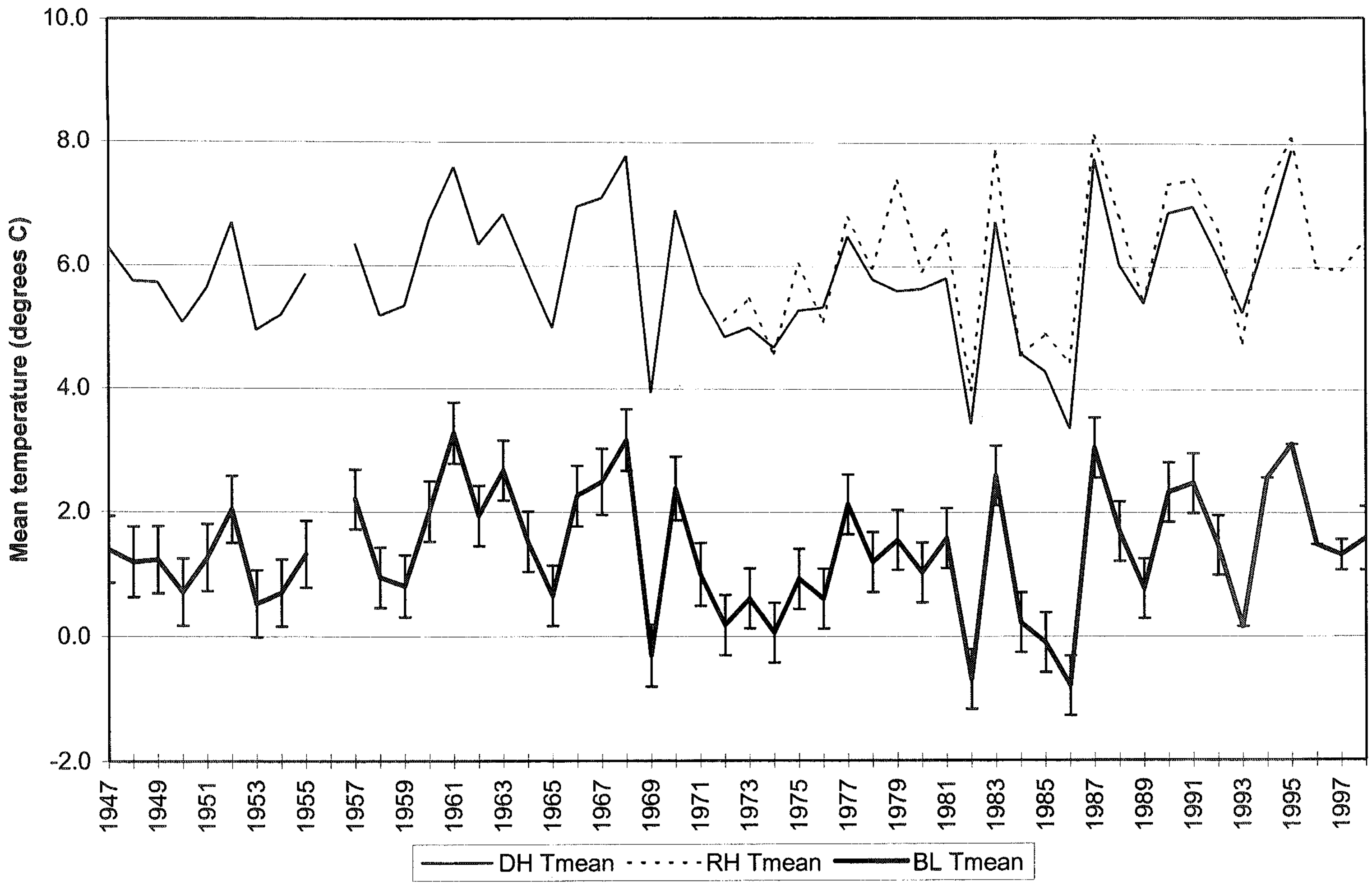


Figure 3.6 **Big Level and coastal plain October mean temperatures.**
Period from 1947 to 1998 with 1956 missing.

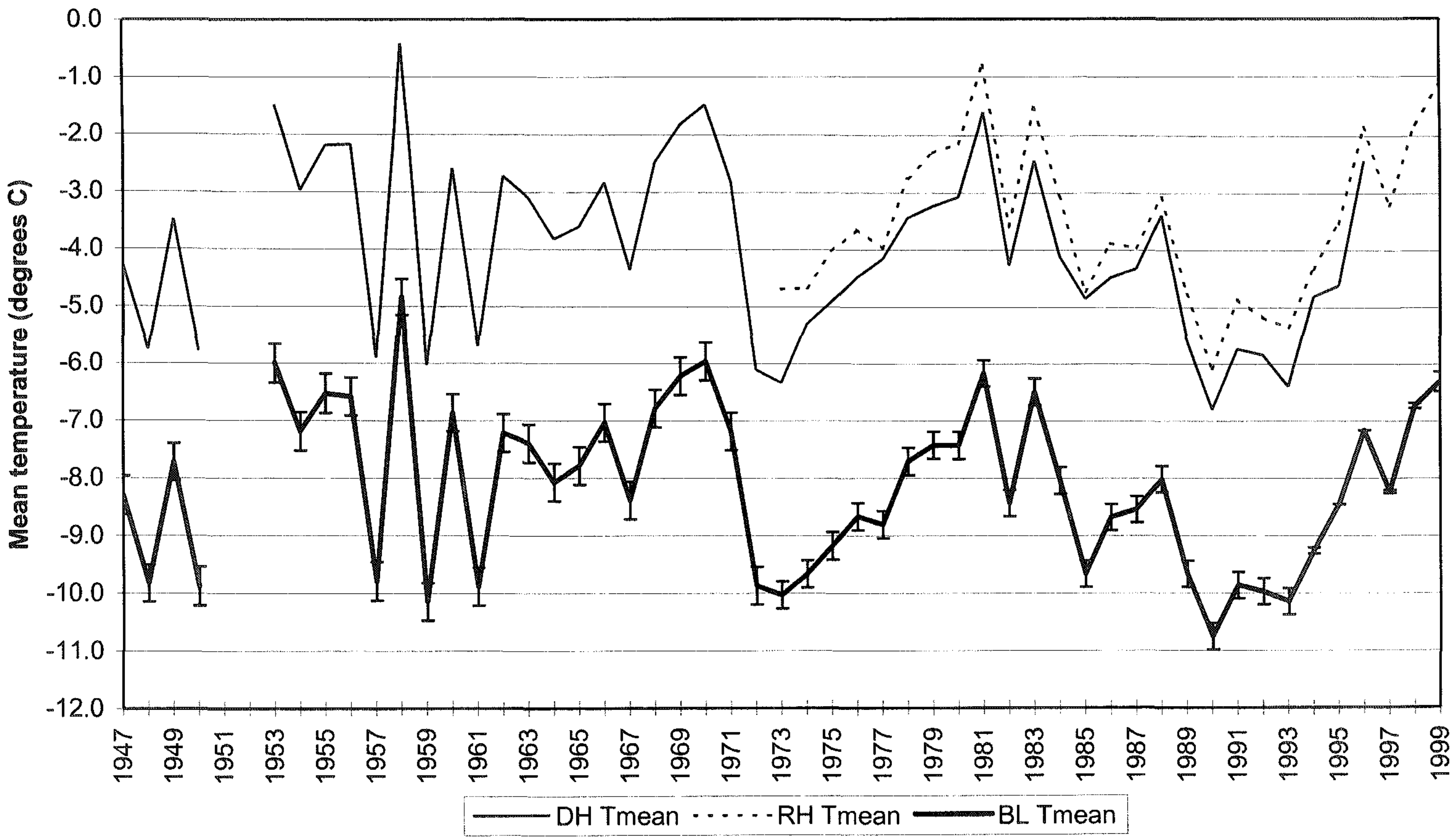


Figure 3.7 **Big level and coastal plain snow accumulation season mean temperatures.** Period from 1947 to 1999 with 1951 and 1952 missing.

and 1958) had one monthly temperature value missing, which was replaced by the mean monthly value for the whole period. The years 1951 and 1952 had two missing months each and were not included. However, their values should be above normal, i.e. near the 1953 value, as determined from the Daniel's Harbour December to March average temperatures for those years.

Thus, the estimated mean temperature for the snow accumulation season from 1947 to 1999 is -8.1 ± 0.1°C for BL, while the measured value for DH is -3.9°C, i.e. 4.2°C warmer than on the plateau. To be classified as below or above average, individual snow accumulation season temperatures had to be outside of one standard deviation around the mean. Following this methodology, the 1953, 1955, 1956, 1958, 1969, 1970, 1981, 1983, 1998 and 1999 seasons had above average temperatures, while the 1948, 1950, 1957, 1959, 1961, 1972-1974, 1985, 1989 and 1990-1993 seasons had below average temperatures. There is no significant trend for the seasonal mean temperature during the period of study. However, there are distinct epochs of alternating cooler and warmer periods, which coincide with those identified for Newfoundland and Labrador by Banfield and Jacobs (1998a). According to Banfield and Jacobs (1998a) 1947-1950 would have marked the end of a near average ("stationary") epoch that began in the mid 1930s, although in this study area, 1948 and 1950 were below average when considering the longer November to April period. An unparalleled lengthy warm temperature anomaly would have begun in 1951 and lasting up to 1958, which was the warmest season on the estimated record on BL at -4.8 ±0.3°C. However, the cold season of 1957 can also serve as a boundary, starting a 6 year epoch of striking interannual variability. In fact, the 1957 to 1961 seasons (excluding 1960) alternate from below to above average temperatures. The next notable warm period, 1968-1971, had two above average snow accumulation seasons, 1969-1970, and was followed by three seasons of below average temperatures for 1972-1974. The warm early 1980s had two above average snow accumulation season temperatures, 1981 and 1983, while the cold late 1980s and early 1990s had five consecutive below average years from 1989-1993, with the 1990 season having the coldest estimated mean temperature value for BL at -10.8 ± 0.2°C. The period of study ends with two above average years in 1998 and 1999.

Thawing degree-days using daily maximum temperature (the accumulation of day-degrees above 0°C) offer an opportunity to determine the amount of melting occurring during each season. Figure 3.8 shows the thawing degree-days for the snow accumulation season for the 1947-1999 period. Conspicuous are the five missing seasonal values for 1951, 1952, 1955, 1958 and 1999. Unlike the temperature values, degree-days are initially compiled as monthly totals and therefore the 3/5 rule for missing data does not apply. Only months with two missing values at most were corrected by replacing the blanks with the average monthly temperature. Missing values in months with average temperatures below 0°C were replaced by 0 degree-day values. The additive nature of degree-days results in more missing months than for temperatures and thus more missing seasonal values. However, snow accumulation seasons with only one monthly value missing were attributed a degree-day value. The years 1948, 1949 and 1950 were assigned values by replacing the missing monthly value with the median value for the whole period.

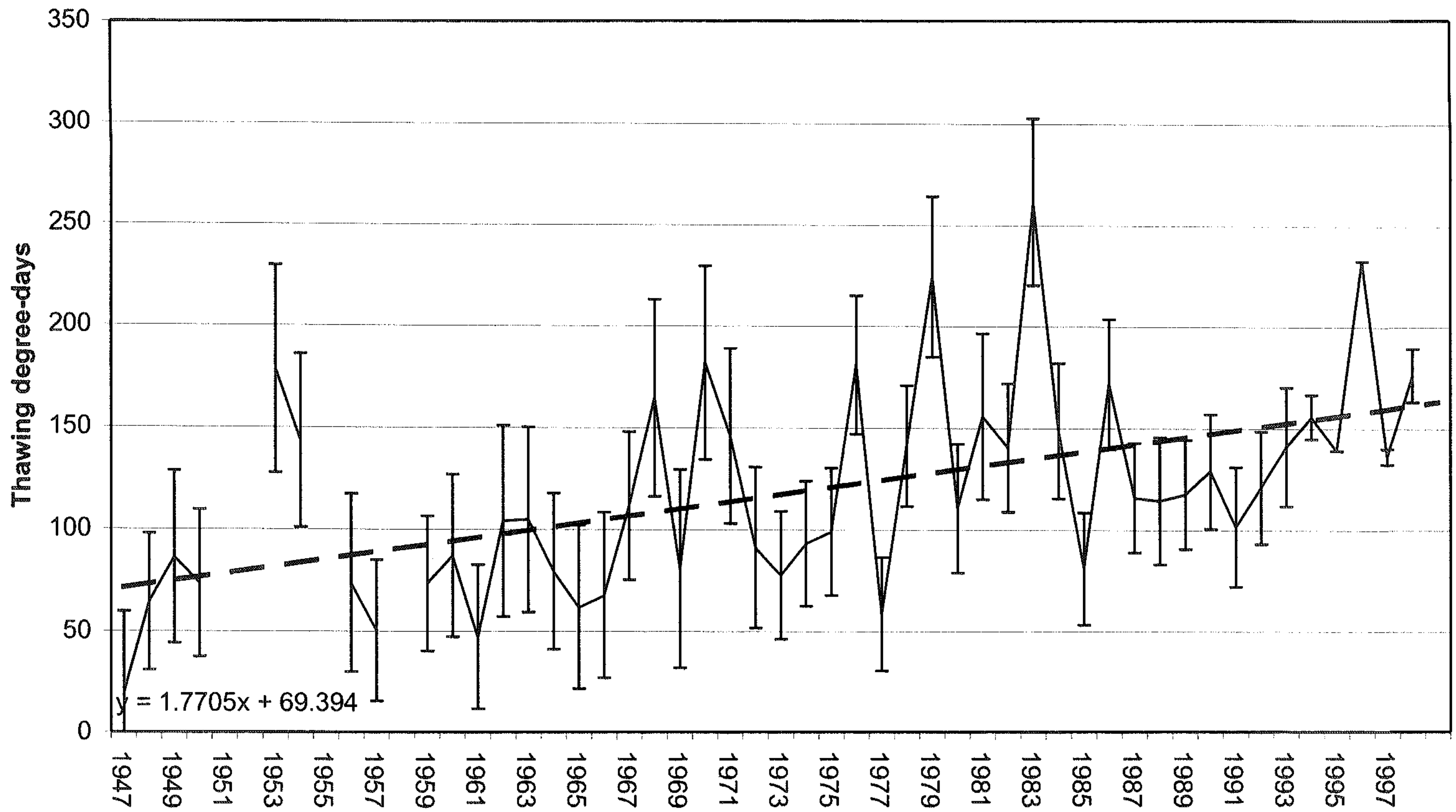


Figure 3.8 **Big Level snow accumulation season thawing degree-days using daily maximum temperature.** Period from 1947 to 1999 with 1951,1952,1955,1958 and 1999 missing. Straight line shows the linear trend for the entire period.

From the above analysis the average estimated thawing degree-day value for the snow accumulation season on BL is very low at 119 ±5°C (i.e. less than the value for the single months of October or May), with above and below average values defined using the same method as for temperature. The 1953, 1970, 1976, 1979, 1983, 1986, 1996 and 1998 seasons had above average thawing degree-day totals while the 1947, 1948, 1950, 1956, 1957, 1959, 1961, 1965, 1966, 1977 seasons had below average totals. The most surprising result is that the snow accumulation season temperature trends are not equivalent (or opposite to) to the degree-day trends. Although a slightly negative but non-significant cooling trend was observed for temperature in this period, a significant (95% level) positive trend of 1.8 thawing degree-days per year was calculated for degree-days using the available data. Even when the warm missing seasons of 1951, 1952 and 1958 are replaced with the 1953 value, there is still a significant positive trend of 1.3 thawing degree-days per year. This shows that in spite of a generally steady overall mean snow accumulation season temperature, the intra-monthly variability in temperatures must have increased. Furthermore, by observing the above and below average degree-day values listed above, it becomes apparent that an above average degree-day season is not necessarily an above average temperature season. This is evidenced by the 1976, 1979, 1986 and 1996 seasons. The inverse is also true in two different years. These particularities demonstrate that seasonal or even monthly temperatures do not always adequately characterize the amount of melt occurring in that time period, especially when temperatures are near the freezing point. It is more relevant to know the length and magnitude of particular melting events in this case.

Higher winter thawing degree-day accumulations at BL translate into more melting events and perhaps less snow accumulation events, although days in which accumulation could occur (below 0°C) greatly outnumber those in which melting occurs, as shown in Figure 3.9. The missing years do not completely coincide because snow accumulation seasons with less than 12 total days missing were kept in the record. These missing days were assumed to have a maximum temperature above 0°C, but minimum and mean temperatures below 0°C. However, missing days are rare, especially after 1972 when the Rocky Harbour station (RH) station commenced. An average of 37 days during the snow accumulation season have a daily maximum temperature above 0°C, which is far below the total number of days in the season (181). These days of above freezing temperatures are spread over an average of 16 events a season on BL, an event being defined as one or a string of days that meet the criterion. As with thawing degree-days, days with the maximum temperature above 0°C increase significantly at the 95% level over the study period by a rate of 0.2 d $yr^{-1}$. The 1953, 1958, 1979, 1981, 1983, 1996 and 1998 snow accumulation seasons had an above average number of thawing degree-days, and the 1947, 1950, 1957, 1959, 1961, 1964, 1965, 1967, 1973 and 1977 seasons had a below average number of days above freezing, with these values being selected if they were outside of one standard deviation around the mean.

Strictly, days with daily maximum temperatures above freezing at BL during the snow season would obviously cause at least a little surface melt, with snow accumulation seasons that have more of these days and associated thawing degree-days having a more complex

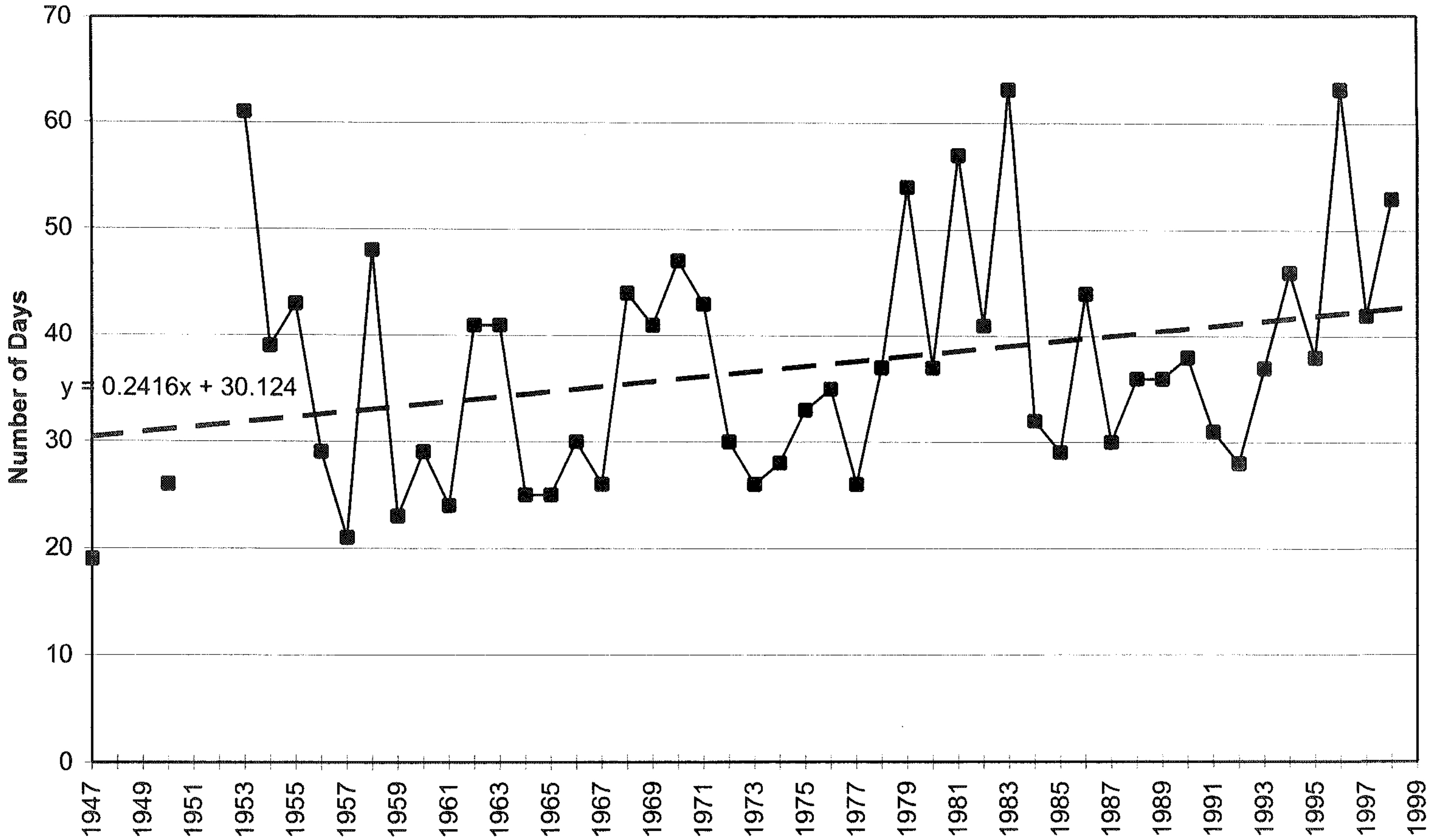


Figure 3.9 **Number of days during the BL snow accumulation season with a daily maximum temperature above 0°C.** Period from 1947 to 1999 with 1948, 1949, 1951, 1952 and 1999 missing. The straight line shows the linear trend for the entire period.

snowpack stratigraphy, with more ice layers. However, with an average of 20% of days from November through April that have a maximum temperature above 0°C, it is evident that possible accumulation days greatly outnumber melt days. Furthermore, if one considers days when the mean temperature and the minimum temperature are also above 0°C, the estimated percentage of days drop to an average of 7% (Tmean above 0°C) and 2% (for Tmin), with most of these more intense thawing days taking place in November and April. Consequently, days with major melt are infrequent on the plateau, even during the warmest of winters. However, there is a significant trend showing that the frequency of these days has been slowly on the rise since 1947.

The snowmelt season (May and June) marks the final period of the snow year. It is a key time in which the snowpack that was formed during the snow accumulation season gradually melts to leave only snowbeds in drift areas by the time it ends. Figure 3.10 shows the mean temperature for this season throughout the study period, with 1950 and 1955 excluded due to missing data. The estimated mean value for BL is 4.5 ± 0.1°C, which is 3°C lower than at DH, and it is generally an uneventful data set, with no strong trends or patterns. Thawing-degree days were also plotted, but the results are similar to temperature, with the curve revealing similar spikes and no significant trend for the period of record. However, the exception to this, concerning temperature and thawing degree-days, is the 1998 and 1999 seasons which are both well above average in an unprecedented fashion. The highest mean temperature value on record for May and June is the 1999 season, being 1.3°C above the value of any other spring season, followed closely by 1998, which is also most

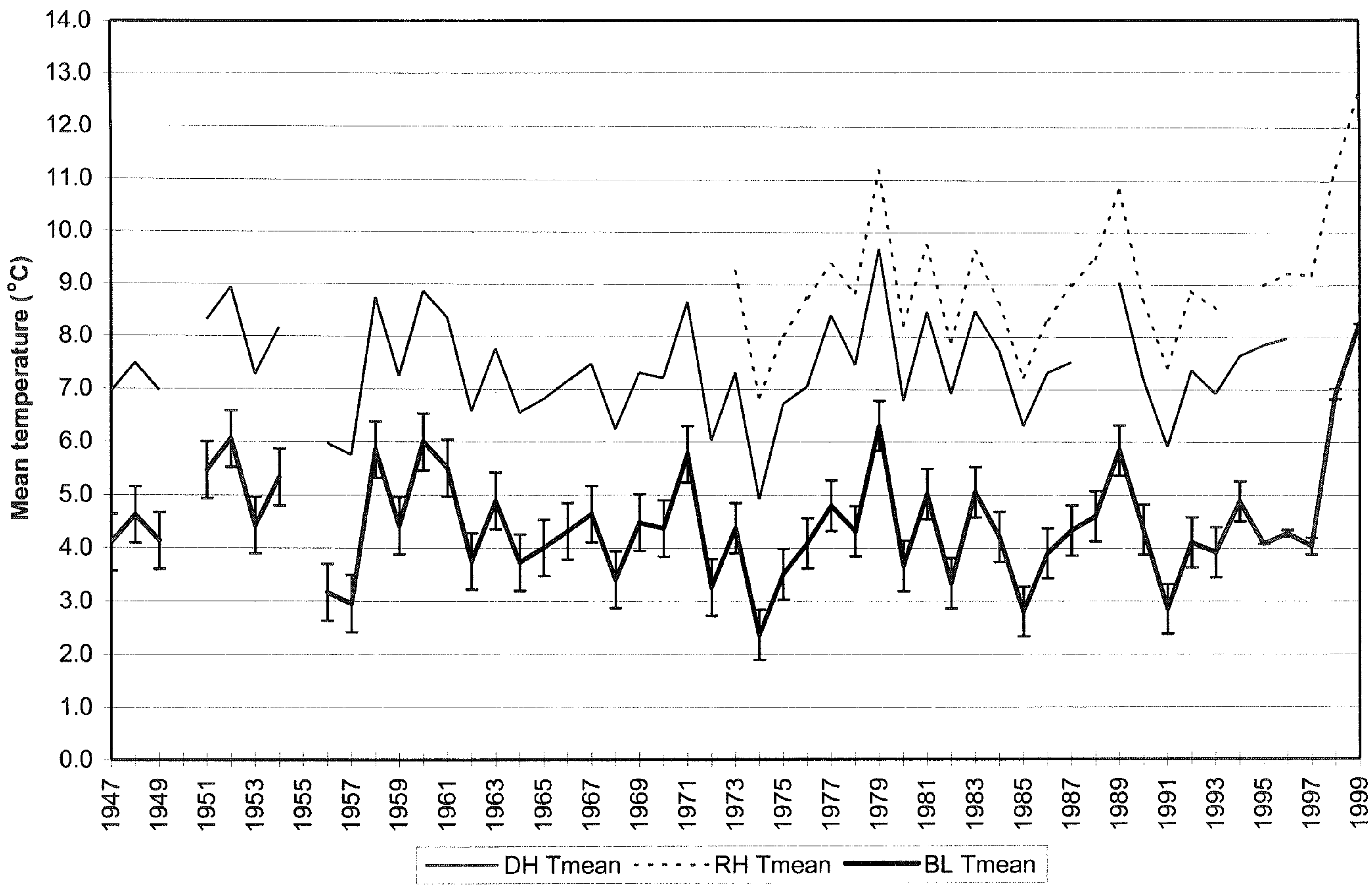


Figure 3.10 **Big Level and coastal plain snowmelt season mean temperatures.**
Period from 1947 to 1999 with 1950 and 1955 missing.

likely a few tenths of degrees Celsius above previously experienced values in the past 50 years on BL. The fact that two such seasons have occurred in succession, compounded with the fact that these two springs have been the warmest since the beginning meteorological records in this area could be a sign of climate change in progress.

There are generally weak relationships between above and below average temperatures within a snow season (October-June) when comparing their temperatures for October (autumn), snow accumulation and snowmelt season temperatures. Only one season had an above average temperature for both the autumn and the snow accumulation season, which was 1969 (1968-1969). Moreover, only one below average October was associated with a below average snow accumulation season, in 1985. The year 1958 had above average temperatures for both the snow accumulation season and snowmelt season, while 1957, 1972, 1974, 1985 and 1991 had below average temperatures for both these seasons. Finally, the only snow year having a link between autumn and subsequent snowmelt season temperatures was 1971, when both periods had above average temperature values. Not a single snow year had consistently above or below average temperatures throughout the three sub-seasons.

# 4 HISTORICAL BIG LEVEL WIND SPEED

## 4.1 Background

Wind has an immense impact on snowcovers in open windswept areas. The most visible effect of wind is the relocation of snow crystals resulting in relatively denser drifts that can be several times deeper than its initial material (Kind, 1981; McKay and Gray, 1981). This mechanism tends to erode the snowcover in more exposed areas and deposit snow in more protected areas causing a smoothing and leveling of the winter landscape (Adams, 1981; Tabler *et al.*, 1990). On flat and open surfaces in the Canadian Prairies for example, Pomeroy and Gray (1995) suggest that winds can remove up to 75% of annual snowfall. Furthermore, winds generally harden and compact a snowcover resulting in the development of significant resistance to transport after several hours of exposure (Verge and Williams, 1981; Gray and Prowse, 1993). In addition to these effects, very large sublimation rates can result from in-transit snow particles, being up to100-1000 times larger than the rates than can occur directly from the snowpack (Schmidt, 1982; Morris, 1989; Tabler *et al.*, 1990). Thus instead of being negligible, sublimation in windy areas is an essential and perhaps even dominant component of the mass balance of the snowcover (Tabler *et al.*, 1990).

It is generally recognized that wind speed (u) increases with height in mid-latitudes, an area under the westerly wind belt (Barry, 1992). Moreover, winds in the planetary boundary layer typically increase exponentially with height up to 500-1000 m above the surface, i.e. the "friction layer". Above this height in the free air, winds are governed by large scale pressure patterns and are largely independent of the topography (Kind, 1981). In the case of an alpine plateau area, such as Big Level (BL), the vertical compression of air flowing over the summit tends to further increase wind speeds, while the friction with the plateau's surface tends to slow down the airflow in comparison with the free air (Barry, 1992). Because of the smooth nature of the alpine surface (treeless, often snow-covered), the first effect might be expected to dominate the second, increasing wind speeds. However, free air wind speeds are said to be comparable with those at mountain summits (Barry, 1992). An increase in wind speed by a factor of 1.5 over a convex dome-shaped summit is often assumed in comparison with the speed observed at the base of a hill (Walmsley *et al.*, 1989). For this study area, Banfield (1995) reported average monthly ratios of wind velocity between BL and Rocky Harbour (RH) of 2.0-2.5 from September 1993 to March 1994. These ratios would be expected to be lower when comparing BL to Daniel's Harbour (DH) because of the relatively sheltered location of RH from some wind directions off the Gulf of St. Lawrence.

The standard measurement height for surface wind speed and direction for most climatological studies is 10 m. This means that measurements made at other heights often have to be corrected to be readily comparable with standard observation heights from other locations. The wind profile near the surface is heavily dependent on the surface features. Obstacles like microscale topographic features, trees, shrubs and rocks tend to slow down the wind near the surface. In fact at the surface itself the wind speed is nil and increases with height until it reaches a point where the surface plays no role in its speed or direction (Kind, 1981; Pomeroy and Goodison, 1997). The roughness of the surface represented by $z_0$, the "roughness length", is the key factor determining how quickly the wind speed increases with height. The following equation describes the change in wind speed with height for a surface with only small obstacles (derived from Oke, 1987):

$$u_1 = u_2 * \ln (z_1 / z_0) / \ln (z_2 / z_0) \qquad \text{(m/s)} \qquad \text{Equation 4.1}$$

where:

$u_{1\,(\text{or } 2)}$ = wind speed at height 1(or 2) (m/s)
$z_{1\,(\text{or } 2)}$ = height 1 (or 2) (m)
$z_0$ = roughness length (m)

It assumes a neutral wind profile, which is typical of windy environments where considerable mechanical mixing occurs (Oke, 1987). Goodison *et al.* (1998) recommend a $z_0$ value of 0.01 m during winter for standard (i.e. official) meteorological measuring sites such as DH. During the growing season, $z_0$ would be similar to the winter value proposed by Goodison e*t al.* (1998), as short grass prairie surfaces also have $z_o$ values near 0.01 m (Oke, 1987). By assuming a constant $z_0$ value, the inspection of whether there is snow on the ground becomes unnecessary, even though many snow surfaces may have roughness values under 0.01 m. Furthermore, also for simplicity, the variable height of snow above the surface, which would change the effective height of the wind measurement, is ignored. The micro-scale effects of this assumption are considered to be minor. This is the standard practice for wind measurements from the Meteorological Service of Canada and to attempt corrections for this variable would be too time consuming. Finally, $z_0$ for the BL site can also be assumed as 0.01 m, as it is a relatively extensive and flat surface with short vegetative cover, similar to a short grass prairie.

The climatological wind record at DH is deemed to be representative of the annual regime of the coastal plain and to a lesser degree of GMNP in general (Banfield, 1990). The mean monthly wind speeds at DH for the 1962 to 1999 period are presented in Figure 4.1. The mean monthly speeds vary in a narrow band of values from 21 km/h (April to June) to 27 km/h (January). Wind speeds in fall and winter are generally higher than those in spring and summer, as pressure gradients are larger during the former seasons. The most frequent wind direction for every month has a westerly component, with December and January exhibiting a westerly most frequent direction and all other months exhibiting a southwesterly direction (SSW winds are often recorded in the summer). Although these directions are more frequent, directions can vary considerably over periods of one to three days with the passage of transient synoptic scale pressure systems. Transient low pressure systems outnumber anticyclones by about two to one in winter over the Island (Banfield, 1983). Over northwestern Newfoundland, northwesterly flows are common during winter. In spring, northerly and easterly components to the airflows are relatively more common in the province (Banfield, 1983).

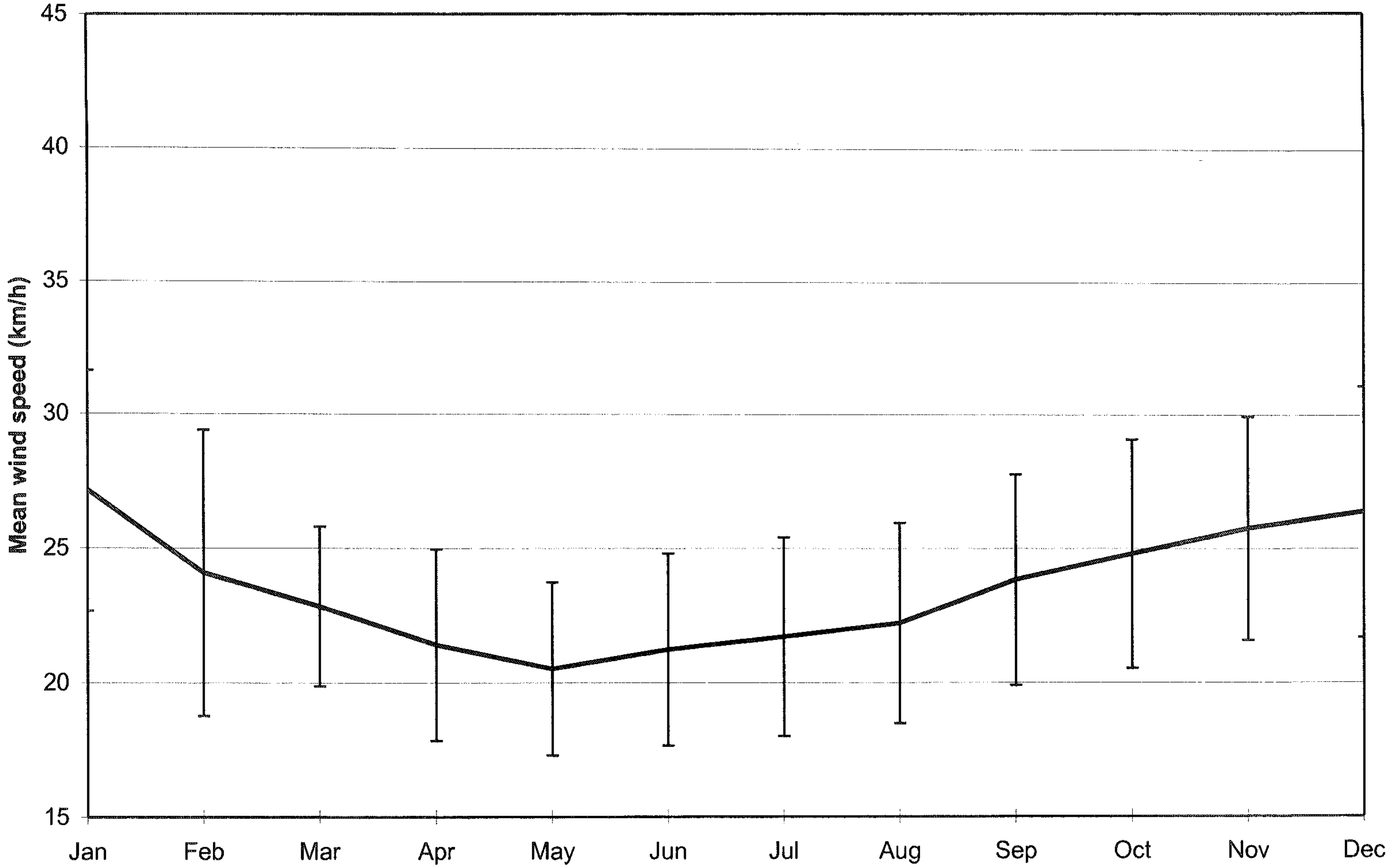


Figure 4.1 **Daniel's Harbour historical monthly mean wind speed (10 m).** Period from January 1962 to June 1999 with 0-3 missing years per month. Error bars show the standard deviation of the monthly mean values.

## 4.2 Methodology

The historical record of mean hourly wind speed was estimated for the Big Level Summit Plateau (BL) for the period between January 1962 and June 1999 by using multiple linear regression. The method used here is analogous to the one used for the estimation of historical BL temperatures (section 3.2). The predictors used to estimate BL hourly wind speed were DH hourly wind speed, SUA 780 m twice daily wind speed (0830 and 2030 NST) and BL average daily values for use before the BL station had an hourly observation program, from 1993 JD 246 to 1996 JD 175.

The first and most important predictor is the hourly wind speed record at DH because of its proximity to BL and its mostly complete wind speed record. However, some modifications to its record were made: 1) prior to October 1963 (JD274), the measuring height of wind speed and direction was 7.4 m instead of the conventional 10 m. Equation 4.1, including the previously selected value of 0.01 m for $z_o$, was used to correct the wind speed values to the standard height of 10 m. 2) Prior to 1965 JD 346, DH did not have a full hourly record, measuring wind speed and direction only three to six times a day between 0830 and 2030 NST. Therefore, the 0830 NST value was used as an estimate of the hourly values from 0230 NST to 0930 NST for this period and the 2030 NST value was used to estimate the 1830 NST to 0130 NST value (next calendar day, same climatological day). The other wind observations between 1030 and 1730 NST were used to estimate the other missing wind speed values during the mid-day period.

The SUA data, in contrast, had a large number of missing values. The 780 m wind speed and direction values for each sounding were calculated using linear interpolation from the levels immediately above and below 780 m. More than 66% of the values at these levels were missing, however. The wind speed values for the 830 sounding were used to estimate the BL hourly wind speed from 0230 NST to 1330 NST, while the 2030 sounding was used to estimate BL hourly speeds from 1430 to 0130 NST. Finally, for the most recent three years (except when malfunctioning), the average daily wind speed (BLD) from 1996 JD 220 was used as a predictor to estimate BL hourly wind speed for the period prior to BL's hourly observation record.

Equations estimating hourly wind speed at BL, incorporating the available predictors, were constructed for each hour of the climatological day for the entire year for the January 1962 to June 1999 period. A total of 24 models were thus constructed with a possible maximum of 6 equations per model: equations using only the Big Level average daily wind speed record (BLD) were not attempted because of its limited use. The Durbin-Watson test was used for serial correlation, with corrections made using the generalized least squares technique when applicable. Section 3.2 provides more details concerning the procedures followed, which are identical to those used for temperature. The equations were prioritized within the models using an ascending order of their standard errors, with the equations having the lower standard error estimates being preferentially used before those with higher ones. Thus, for each given hour of the day the equation used for the prediction of the BL wind speed was then accomplished using the independent variables available for that time.

The reconstructed record of BL wind speed, including estimated and measured values, was corrected to a height of 10 m from its measurement height of 3 m to be directly comparable to data at other stations, specifically DH in this study. Equation 4.1 was used for this purpose. The reconstructed record of hourly BL wind speed was first compiled into average daily values if more than 19 hourly values were present on any given day. Subsequently, average monthly wind speed values were derived using the 3/5 rule (three consecutive days or five days total) for missing data as the criterion to determine if the value is present, with the two standard error of the estimations used as the uncertainties. The estimated BL and the DH monthly wind speeds were then used to calculate the average wind speed values for the three seasons of the BL snow year: 1) October, 2) the snow accumulation season (November - April) and 3) the snowmelt season (May - June).

## 4.3 Results and Discussion

### 4.3.1 The Wind Speed Models

Once the regressions were completed, a total of 96 equations were determined to be significant (>0.001 level) in the 24 models estimating hourly wind speed at BL, four equations per model. Half of these equations were corrected for serial correlation with the generalized least squares technique (section 3.2). The construction of equations that would have simultaneously used BLD and SUA as independent variables was not carried out because of the very limited time period during which these equations could have been used. Therefore four equations per hourly model remained, with generally the same rank for each set of independent variables for every hour. In fact, equations using the same predictors were so similar from one hour to the next that there was no significant difference between them, signaling that they could be combined to simplify the estimation procedure. This fact demonstrates that large scale factors dominate the determination of wind speed for GMNP, because of the small size of the sub-diurnal signal in wind speed. Thus, wind speed patterns depend primarily on the synoptic-scale systems affecting the area, with the time of day, related to local factors, having a very limited effect.

Each group of 24 equations with the same predictors was combined into one expression for a total of four equations . The statistics of these equations are presented in Table 4.1 and the coefficients are shown in Table 4.2. These equations output wind speed at BL in km/hr and require wind speed values in km/hr for the estimators. Additional information on the columns of the tables can be found in section 3.3.2 as they are equivalent to those used for the temperature models (Tables 3.1 and 3.2). The statistics and the coefficients of the wind speed equations were determined by averaging the values for each group of 24 equations. These models were tested against the BL actual record. Equations 2, 3 and 4 each have standard error values (x2) of 13, 15 and 17 km/hr respectively for estimated mean daily wind speed, determined by a comparison of estimated BL daily wind speed values to the daily average measured values at BL. This kind of analysis does not

Table 4.1 **Statistics of the regression equations used in the wind speed models.** The equations are applicable to any given hour of the day.

| Equation rank | Independant variables | 2 std err (km/h) | r | r squared (adjusted) | n | Durbin-W |
|---|---|---|---|---|---|---|
| 1 | DH, BLD | 16 | 0.832 | 0.691 | 338 | 2.038 |
| 2 | DH, UA | 21 | 0.691 | 0.467 | 99 | 1.906 |
| 3 | UA | 21 | 0.525 | 0.273 | 95 | 2.023 |
| 4 | DH | 22 | 0.616 | 0.379 | 337 | 1.969 |

n = number of cases used to build the equation

Table 4.2 **Coefficients of the regression equations used in the wind speed models.** The equations are applicable to any given hour of the day.

| Equation rank | Independant variables | Constant | DH | SUA | BLD |
|---|---|---|---|---|---|
| 1 | DH, BLD | -1.486 | 0.283 | n/a | 0.833 |
| 2 | DH, UA | 4.167 | 0.553 | 0.231 | n/a |
| 3 | UA | 10.255 | n/a | 0.316 | n/a |
| 4 | DH | 9.500 | 0.734 | n/a | n/a |

n/a = not applicable

apply for equation 1 since mean BL temperature (BLD) is one of the predictors. Although the hourly and daily errors of the estimations are large, they are much smaller when considering monthly or seasonal average values (±1-3 km/h).

### 4.3.2 Big Level Historical Wind Speed Record

The estimated BL wind speeds for the 1962-1999 period were initially compiled into monthly means for the period. These results are presented in Table 4.3 and Figure 4.2. The wind speed series at DH (Fig. 4.1) and BL have very similar fluctuations through this period, with speeds at BL 8 km/h higher on average than at DH. From the values adopted for both sites, this yields an average BL/DH wind speed ratio of 1.33, substantially lower than the range advanced by Banfield (1995) based upon comparison of daily mean and peak speeds at RH and BL over a seven month period. Average monthly wind speeds vary from 29 to 36 km/h at BL with the seasonal distribution being equivalent to that at DH.

The effect of having only a few hourly wind speed values per day, at DH, prior to December 1965 was analyzed in order to verify the representativeness of the wind speed estimations form 1962 to 1965. The hours with wind speed values available prior to December 1965 were compared with the full record of hourly observations concerning mean daily wind speed for the year 1968 (randomly picked, no missing days). The biggest flaw in the data prior to December 1965 is the absence of measurements between 2030 and 830 NST. The average of the available hours for dates prior to the full daily record for each day,

Table 4.3 **Big Level estimated monthly mean wind speed (10 m).** Period from January 1962 to June 1999.

| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mean | 36 | 33 | 32 | 30 | 29 | 29 | 30 | 30 | 32 | 33 | 34 | 35 | 32 |
| Maximum monthly | 48 | 47 | 38 | 36 | 34 | 35 | 35 | 38 | 38 | 41 | 42 | 46 | |
| Minimum monthly | 30 | 25 | 27 | 24 | 23 | 24 | 24 | 21 | 26 | 25 | 24 | 27 | |
| Standard deviation | 4 | 5 | 3 | 3 | 3 | 3 | 3 | 4 | 3 | 4 | 4 | 4 | |
| Number of years available | 37 | 38 | 38 | 38 | 37 | 37 | 35 | 36 | 37 | 37 | 37 | 37 | |

All wind speed values and related indices are in km/h

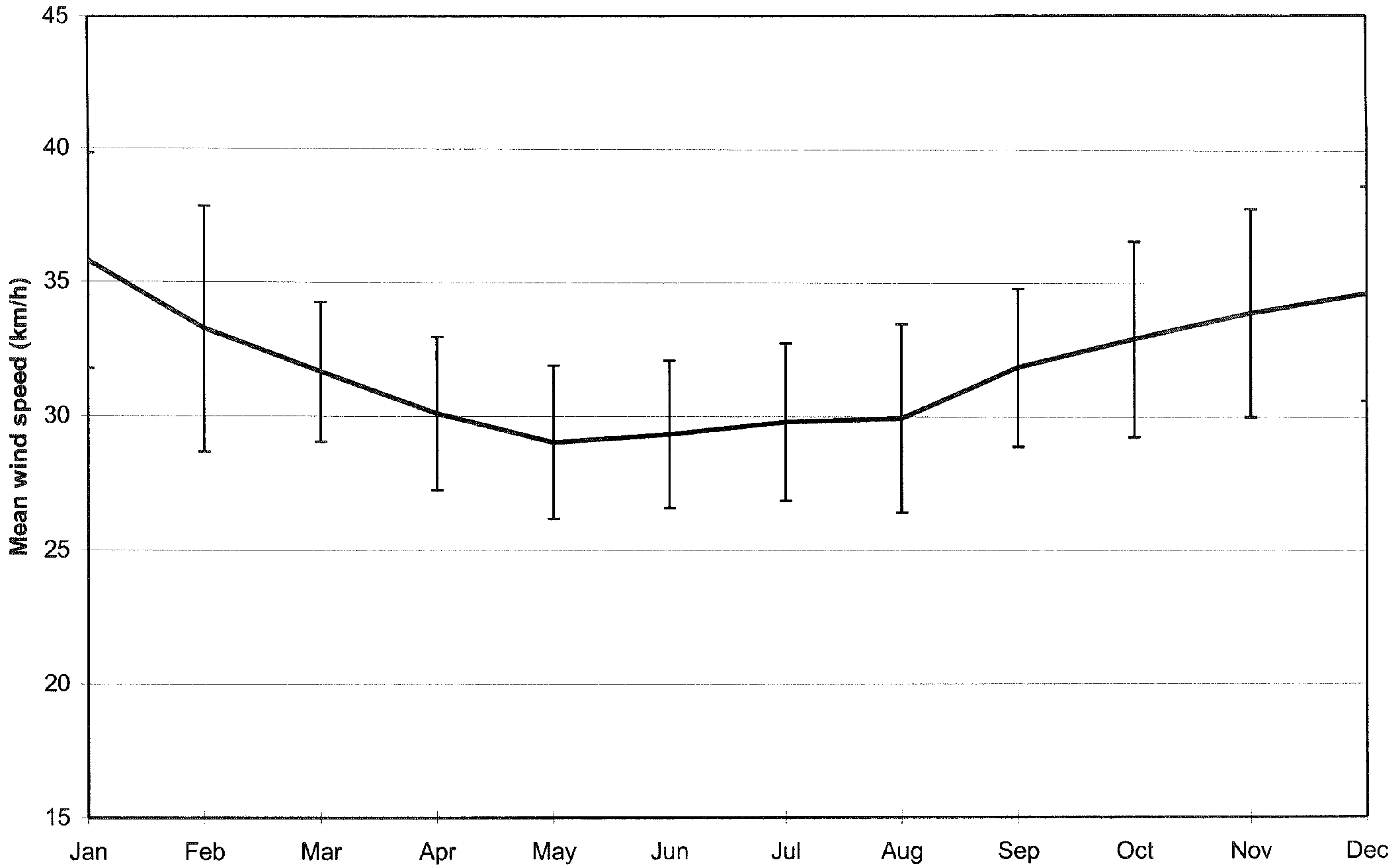


Figure 4.2 **Big Level Summit Plateau estimated historical monthly mean wind speed (10 m).** Period from January 1962 to June 1999 with 0-2 missing years per month. Error bars show the standard deviation of the monthly mean values.

considered to be the daily average wind speed value, was only 0.5 km/h lower than the average obtained using the full record in 1968. This value is negligible when considering that wind speed is only measured to the nearest km/h at DH and that the error on the estimation of BL wind speed far exceeds it. Furthermore, while the effect of a lower height of measurement before October 1963 is not known, an increase of only 2.6 m at that range of height (7.4 to 10 m) should only cause a very small error. Daily mean wind speed values prior to December 1965 are thus considered to be essentially correct.

Consideration of the evolution of monthly and seasonal wind speeds during the study period followed the above analysis. The BL snow season (October - June), is divided into three sub-seasons: 1) October (or autumn), 2) the snow accumulation season (November - April) and 3) the snowmelt season (May - June). Wind speed trends for each of these periods are considered in turn.

October is not very important in terms of blowing snow, but a brief examination of this month is warranted, especially when considering the results. Figure 4.3 shows the historical DH and estimated BL monthly wind speed. Monthly mean wind speed values at BL vary from an estimated low of 25 ±1 km/h in 1996 and 1997 to a high of 41 ±3 km/h in 1966 (uncertainties are the 2 standard error values of the estimates). A significant negative trend at the 95% confidence level exists in the BL series, with a slope of -0.23 km $h^{-1}$ $yr^{-1}$. A very similar trend also exists in the DH data set. All of the above average monthly wind speeds at BL occur prior to 1985, while all the below average values occur after 1981, with above and below average defined as a minimum of one standard deviation outside the overall

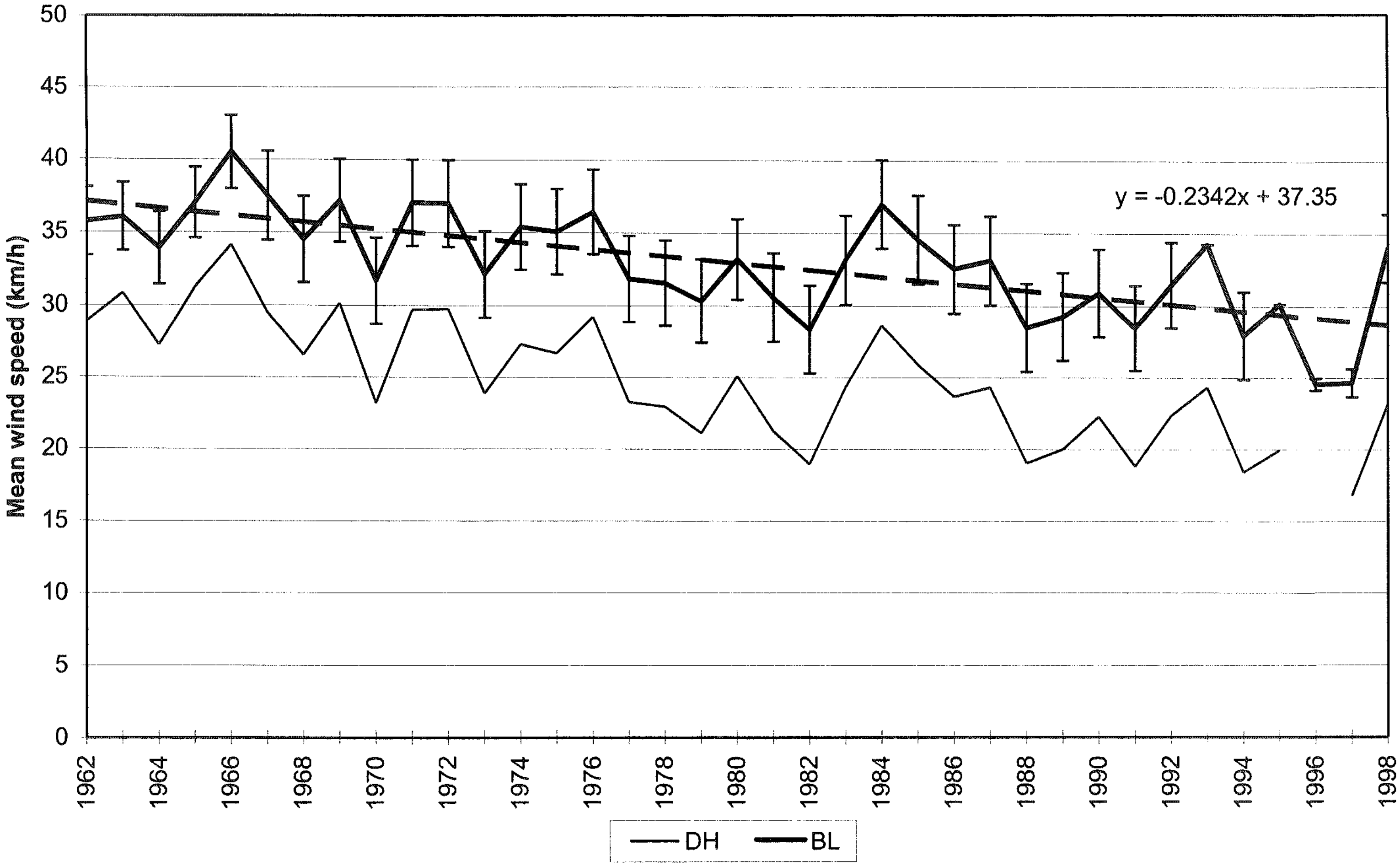


Figure 4.3 **Big Level and Daniel's Harbour October mean wind speed (10 m).** Period from 1962 to 1998. Error bars show the uncertainty of the BL estimated values. Dotted line shows the linear trend for the entire period at BL.

monthly mean value for 1962-1999.

The wind speed record for the snow accumulation season is the most important in terms of snow drifting. The inter-annual variations in wind speed for this season are less pronounced than in October, but are nevertheless present as shown in Figure 4.4. The maximum mean seasonal wind speed for BL was in 1963 and 1984 at 38 ±1 km/h. Furthermore, six seasons are tied for the lowest mean of 30 ±1 km/h, all occurring after 1988. A significant negative trend also exists for the snow accumulation season at BL (and at DH), having a slope of -0.12 km $h^{-1}$ $yr^{-1}$. The 1960s were clearly a windy decade, having four above average snow accumulation season mean speeds. Also, the 1984 and 1985 snow accumulation seasons were relatively windy, especially in compared with the 20 year period around them. In contrast, all of the below average wind speed values were after 1979 and all but two were after 1988, making the 1990s the least windy decade in the four decades studied during for the season in question.

The third period of the year studied, the snowmelt season, also has a significant negative wind speed trend for the period between 1962 and 1999 (see Figure 4.5). The BL record's linear trend slope is -0.15 km $h^{-1}$ $yr^{-1}$. All of the above average melt season wind speeds happened before 1987, whilst all the below average ones occurred after 1974. Two seasons had too much missing data to include in the analysis: 1988 and 1993. Wind speed values for the snowmelt season are generally lower than during October or the snow accumulation season. The greatest mean snowmelt season value on record is 34 ±2 km/h in 1967 and 1994, while the minimum occurred in 1998 at 24 ±1 km/h.

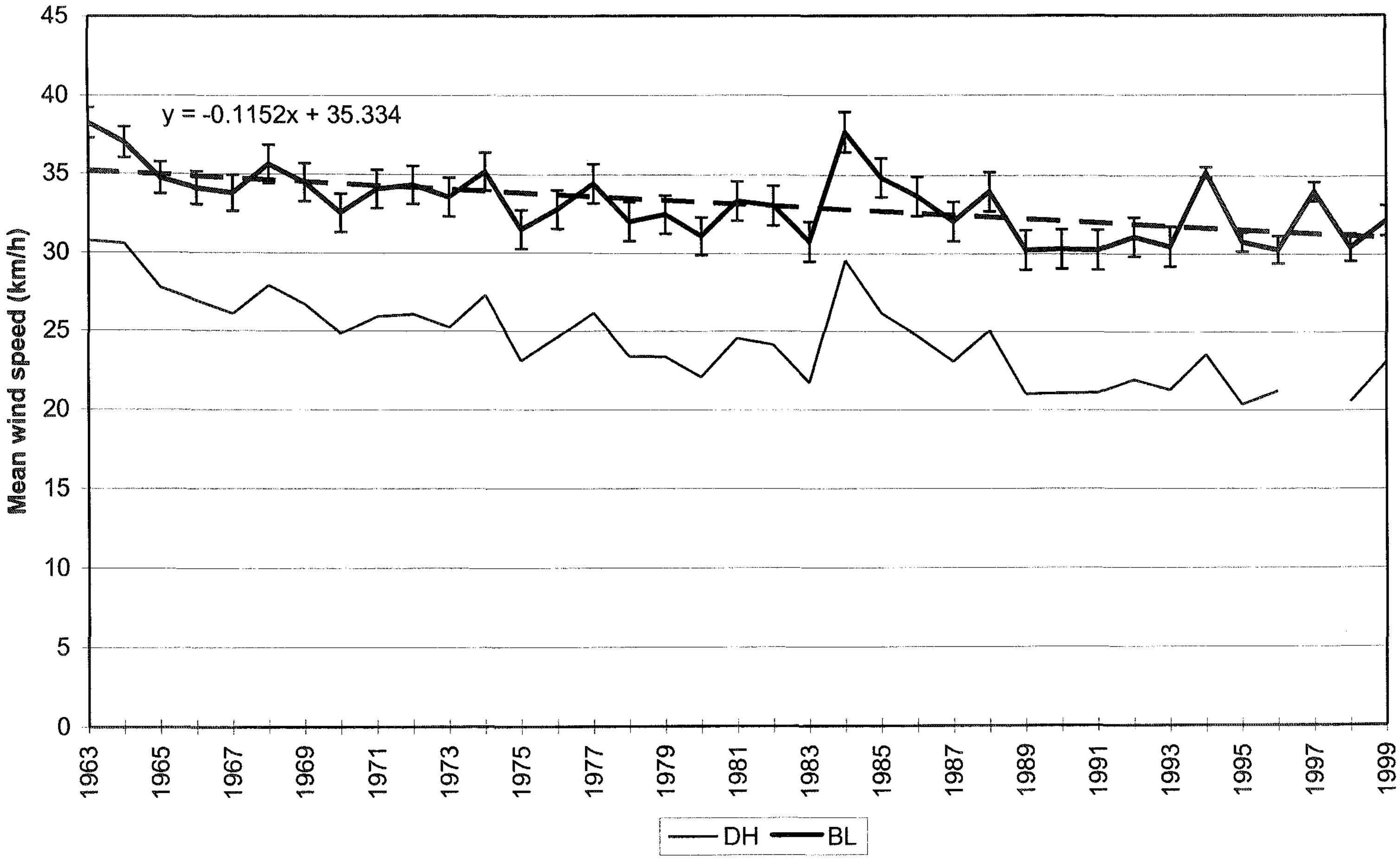


Figure 4.4 **Big Level and Daniel's Harbour mean snow accumulation season wind speed (10 m).** Period from 1963 to 1999. Error bars show the uncertainty of the BL estimated values. Dotted line shows the linear trend for the entire period at BL.

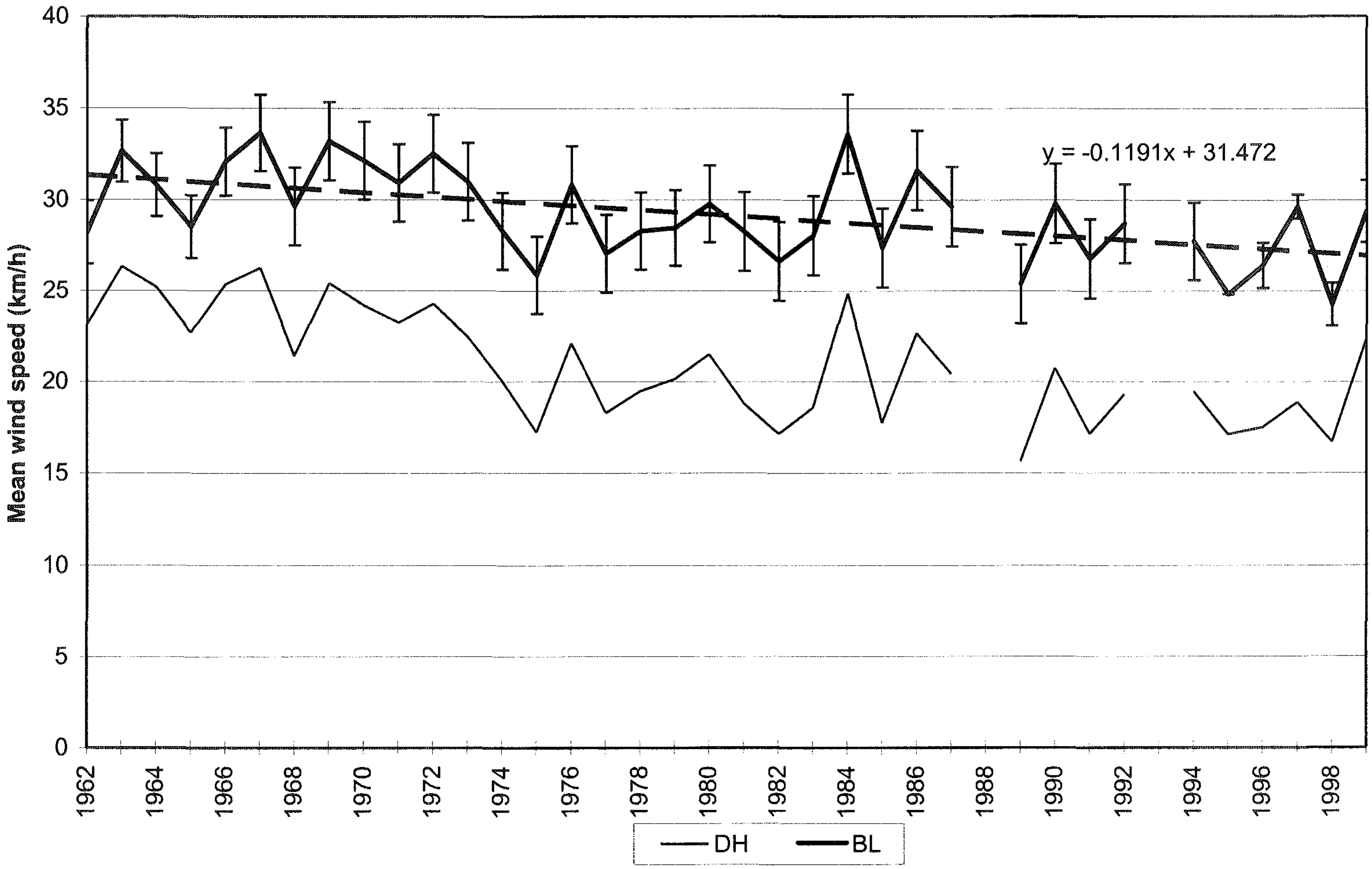


Figure 4.5 **Big Level and Daniel's Harbour mean snowmelt season wind speed (10 m).** Period from 1962 to 1999 with 1988 and 1993 missing. Error bars show the uncertainty of the BL estimated values. Dotted line shows the linear trend at BL.

The wind speed relationships used between the coastal plain (DH) and BL to assist in the estimation of the historical BL wind speed record were built using the least windy decade, the 1990s. The statistical relationships between BL and the coastal plain are accurate only if the relationship between the change in wind speed with height is constant throughout the study period. If the altitudinal variations in wind speed changed over the entire study period, it would lead to less accurate wind speed estimations during the earlier windier years. However, this issue should not cause much concern since the overall linear decreases of wind speed for the entire study period (1962-1999), -8.5 km/h, -5.5 km/h, and -4.4 km/h for the three respective snow season periods are of the same order of magnitude as the year to year wind speed fluctuations and their uncertainties during the calibration period. Although the estimations of BL wind speed values for the windy 1960s may be a little less accurate as a result of the trends, the difference should not be of significant importance.

The trends of decreasing wind speed trends for the entire snow season between 1962 and 1999 are an interesting particularity of the period of study, which would appear to be due to macroscale dynamic factors. In a strict sense, a significant reduction in mean wind speeds implies a causal change or trend to a diminished mean pressure gradient, but it is not known how large an area this phenomenon has affected around GMNP or its relation to the regional weather patterns. However, it is known that the lowest wind speeds occurred during the coldest snow seasons (late 80s to 1995) and may be due to more frequent cold anticyclonic circulations air extending over the Park during these months. On the other hand, one could speculate that a general hemispheric or global warming of the atmosphere, which would occur at greater rate in the northern latitudes, could lead to a reduction in the intensity of the westerlies. This effect would occur predominantly during the cold season, when changes in zonal temperatures would be largest. The long term record of wind speed in western Newfoundland during the snow season is a topic that needs further study.

# 5 HISTORICAL BIG LEVEL PRECIPITATION

## 5.1 Background

Precipitation in the form of snowfall is the building material of a snowcover. Goodison *et al.* (1981) define snowfall as: "the depth of fresh snow which falls during a given recent period." This quantity can be measured directly with a ruler or collected with a gauge and melted to yield a water equivalent (SWE). Rainfall also has an important impact on a snowcover through the changes it causes. The energy brought to a snowpack by rainfall can cause substantial melting. It can also cause large changes to the snow structure and even add to a snowcover water equivalent by freezing within the pack (Male and Gray, 1981).

The coastal plain within and near Gros Morne National Park (GMNP) receives plentiful precipitation amounts year round. This precipitation is caused principally by the passage of frontal cyclones near the area, and to a lesser extent by convective or conditional instability as cool continental airmasses gather moisture over the warmer Gulf of St. Lawrence during fall and early winter (Banfield, 1990; Banfield and Jacobs, 1998a). The measured mean annual total precipitation at Daniel's Harbour (DH) for the study period, from October 1962 to June 1999 is 1152 mm. The monthly regime is shown in Figure 5.1. The most conspicuous characteristic of average total precipitation is the minimum observed in spring (April = 62 mm; May = 78 mm). This relatively drier period is attributed to a larger frequency of easterly and northerly airflows during this time

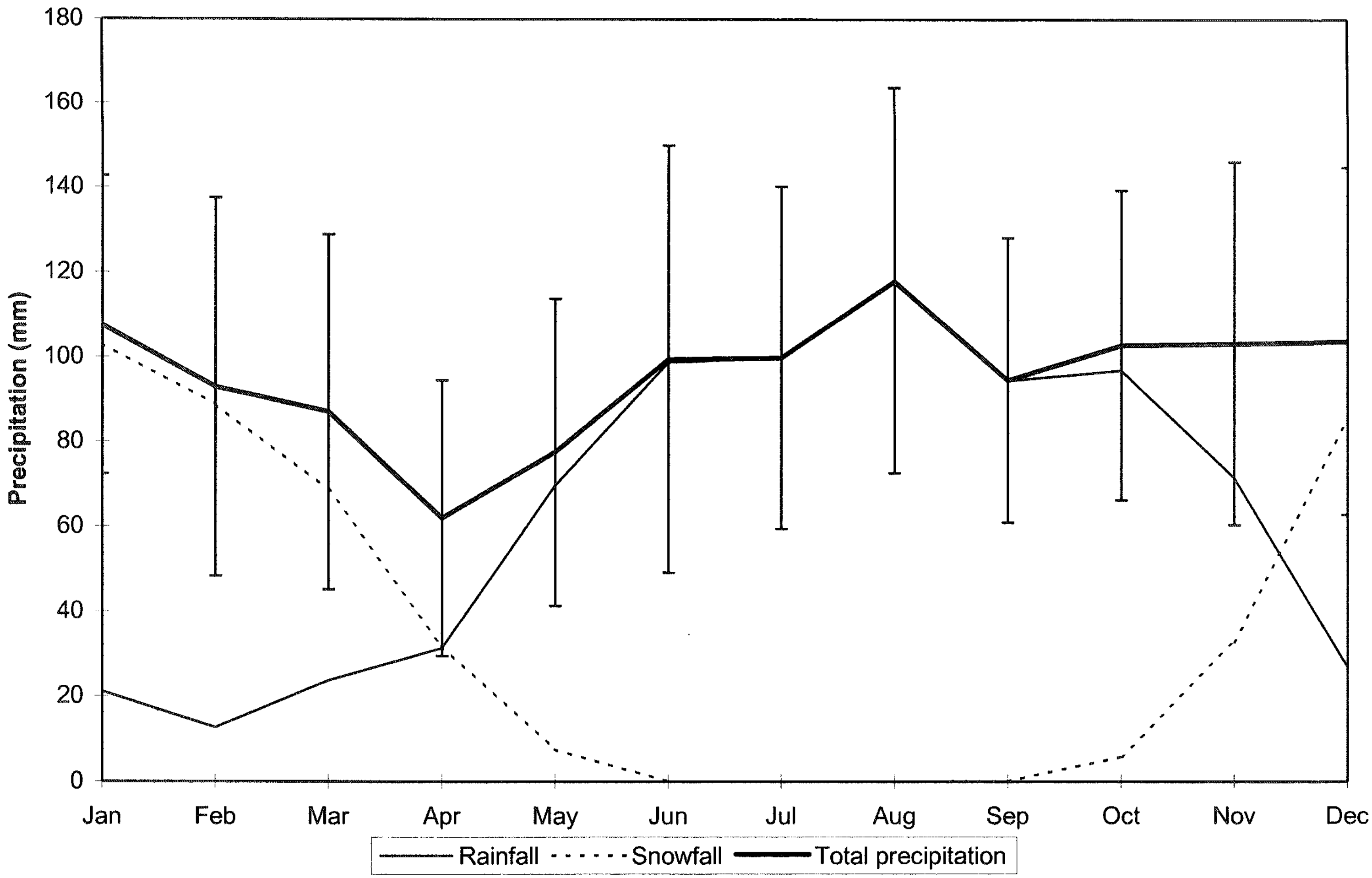


Figure 5.1 **Daniel's Harbour monthly mean precipitation.** Data uncorrected for wind-induced undercatch. Period from October 1962 to June 1999 with 3-5 missing years per month. Bars show the standard deviation of the monthly total precipitation values.

of year, which bring reduced precipitation to the west coast of Newfoundland (Banfield, 1990). In contrast, summer (June-August), fall and early winter (October-January) are wetter, with these periods averaging over 100 mm per month. Furthermore, late winter (February-March) normally sees a drop in precipitation, mainly due to the freezing of significant portions of the Gulf, cutting off this important source of moisture. Banfield (1990) analyzed Rocky Harbour's (RH) record between 1972 and 1987 and he obtained similar mean monthly values for February to September, and significantly larger precipitation values from October to January compared with BL. Given that convective precipitation from the Gulf is important during this period in addition to contributions from low pressure systems, larger amounts are to be expected, and yet they do not figure as prominently in the DH record. This is partially related to the methods used for precipitation measurement, with DH using a Nipher gauge for total precipitation and RH at that time relying on a ruler for the snowfall portion of total precipitation.

The rainfall and snowfall totals, including other forms of liquid and solid precipitation respectively, which together define total precipitation are individually marked by a large seasonality due to temperatures crossing the freezing point. It is clear from Figure 5.1 that total precipitation consists almost exclusively of rainfall from June to September on the coastal plain, while precipitation there falls mostly in the form of snowfall from December to March, on average.

From examining Figure 5.1, one can see that monthly snowfall and rainfall totals during the snow season exceed total precipitation. This is because snowfall is measured with a ruler while total precipitation consists of the SWE of snowfall plus rainfall, measured with a Nipher Gauge. The Nipher Gauge is used at principal stations across Canada and has been used at DH since November 1962 (the gauge was retired on 31 January 1998, marking the end of the usable precipitation record). This instrument gives one of the best winter precipitation estimates in comparison with other gauges due to its inverted bell shield (Nipher shield) which minimizes the turbulence near the gauge's orifice. Therefore, snowfall SWE measured from the gauge at DH is equal to total precipitation minus rainfall, which results in a value of 385 mm SWE per year on average. The ruler method, which is presented in the figure as snowfall, yields a mean annual total of 424 mm SWE, using the commonly used 10 to 1 (0.1) ratio of water content to convert snowfall depth to SWE (Goodison *et al.*, 1981). Ferguson and Pollock (1971) found that an average snow depth/SWE ratio of 0.095 is more appropriate for Newfoundland, which results in an average yearly value of 403 mm SWE for DH, 18 mm greater than the Nipher total. These snowfall totals suggest that approximately 33-35% of the annual precipitation falls as snow on the coastal plain. The snowfall fraction rises to 78-84% of the total if only the months of December through March are considered.

The range of possible snowfall values for DH exposes the inaccuracies of measuring it. In fact, measuring any form of precipitation is an inexact practice. Sumner (1988) described precipitation measurement as being dependant on primitive and imprecise instrumentation and prone to siting and locational unrepresentativeness. In the case of measuring snowfall with a ruler, wind redistribution of snow becomes an important consideration, and much subjectivity may be involved in determining the "true" snowfall by taking a series of depth measurements at different locations (Goodison *et al.*, 1981). In an experiment examining the inaccuracies of rain gauges, Watkins (1955) found a variability of ±5% for monthly rainfall values from nine gauges within a 20 x 20 m grid, a very large uncertainty for such a long reference time period (monthly data) and such a small study area. The overall measuring tendency for precipitation gauge measurements is of systematic under recording with daily rainfall measurement errors being on the order of 5-15% of true precipitation (Sevruk, 1986) and on the order of 20-70% or more for daily snowfall measurements, with wind induced undercatch being the largest problem for both precipitation types, especially in exposed windy environments (Goodison *et al.*, 1998).

It is generally recognized that precipitation is a spatially continuous variable with relationships existing between points not far from each other (Sumner, 1988). Furthermore, Banfield (1990) affirms that Cow Head (CH) and RH have similar precipitation regimes, both being on the coastal plain within GMNP, from observation of their precipitation records. This statement can be broadened to also include DH, also on the coastal plain to the north of the Park. However, in spite of significant correlations between the monthly total precipitation record between these three stations for the period up to June 1999 (over 150 months considered), the monthly total precipitation ratios between them (RH/DH = 1.19, CH/DH = 1.14 and CH/RH = 1.05) have very large standard errors, between 0.42 and 0.55. The largest standard error is, surprisingly, between CH and RH, the two stations located nearest to each other. To illustrate these uncertainties consider the following example: when obtaining an estimate of a monthly precipitation at RH from a 100 mm monthly value from CH, one could only assume that the value lies somewhere between 50 mm and 160 mm with a 68% confidence interval. It is obvious that when precipitation estimates are derived from values from neighboring stations, the usefulness of presenting the statistical uncertainties of the values is highly questionable. However, despite these difficulties, many studies using such estimation methods have been published (see Sumner, 1988).

While the measurement error on rainfall is fairly small and can be left uncorrected, the error on gauge snowfall measurements is too large to be neglected in windy areas such as the coastal plain and BL. The *WMO Solid Precipitation Measurement Intercomparison* (Goodison *et al.*, 1998) reported errors for various gauges are considered and corrective equations were proposed. In the case of the Nipher gauge with Nipher shield and the US NWS precipitation gauge, the latter having an orifice size identical to the Fisher and Porter gauge with Alter shield, only wind speed at gauge height (usually 1-2 m) is considered to be a significant predictor of the undercatch ratio. The equations are:

1) Nipher gauge: $CR = 100 - 1.98*u_p - 0.44*u_p^2$ Equation 5.1

2) US NWS with Alter shield: $CR = EXP( 4.61 - 0.04*u_p^{1.75})$ Equation 5.2

where:

CR = catch ratio ( in %) refers to: catch during wind / catch in calm conditions
$u_p$ = mean precipitation event wind speed (m/s)

The standard errors for these equations are large at 11.05% and 9.77% respectively. These equations show that gauge catch decreases in a quasi-exponential manner with increasing wind speed. Although these equations were designed for precipitation event, or "storm" wind speed, it was found for DH that days with precipitation have on average 18 hours with measurable precipitation, with many large precipitation events undoubtably lasting well over a day (Environment Canada, 1985). Hence, applying this equation to the GMNP area with daily mean wind speed and applying the resulting ratio to daily total snowfall would have only a small distorting affect on these equations. Furthermore, the correction equations were designed for wind speeds up to a maximum of 7.5 m/s. Goodison (2001, personal communication) suggested that wind speed values above 7.5 m/s should not be given higher undercatch ratios, due to the influence ofblowing snow entering the gauge, above these wind speeds. Hence, higher undercatch ratios at higher wind speeds would lead to the overestimation of precipitation values due to the application of the correction factor on precipitation plus blowing snow into the gauge. Instead, the 7.5 m/s wind speed ratio should be maintained when applied to events with wind speed values above this threshold.

Relating precipitation on the coastal plain to that on the Big Level Summit Plateau (BL) must take into account the orographic effect. This phenomenon can be described as a general increase in precipitation amount with altitude, especially encountered in the middle latitudes (Barry, 1992). Sawyer (1956) identified four main factors that contribute to the increase in precipitation with height: 1) Strong winds normal to hills and mountains, 2) a moist air mass with pre-existing cloud layers, 3) a near neutral lapse rate with no inversions and 4) an increase in wind speed with height. Lauscher (1976, in Barry, 1992) proposed a mean orographic increase of 29 mm/100 m per year for the first 1.5 km above sea level for middle latitudes overall. However, in another study set in the windy and moist environment of southern Britain, it was found that the increase in rainfall with height was on the order of 120 - 300 mm/100 m per year on windward facing slopes (Salter, 1918 in Barry, 1992). Barry and Chorley (1998) and Sumner (1988) present other examples of precipitation-height relations. Newfoundland is also considered to be a relatively windy and wet environment. For the island portion of the province Solomon *et al.* (1968), using a statistical approach, related precipitation amounts from most of the island's climate stations to topographic factors using multiple linear regression and found a mean orographic enhancement of 207 mm/100 m per year for the western portion of the island for climate stations under 300 m altitude (there were no stations at higher elevations). Assuming a mean precipitation total of 1200 mm for the coastal plain, this relationship, if extended to highest elevations of the plateau, would translate into average total precipitation of approximately 2700 mm for BL summit. Solomon *et al.* (1968) estimated a somewhat lower yearly average total of 2600 mm for a 100 $km^2$ area including BL, taking in account several topographic and physiographic factors, but this estimate would have been for an average elevation considerably lower than at the BL site considered on its own. This orographic increase is the equivalent of a BL/ coastal plain precipitation ratio of 2.3 : 1. However, caution is advised when applying an annual orographic enhancement factor to periods of less than a year as seasonal changes in airmass characteristics and wind direction lead to events with significantly different precipitation increases with height (Banfield, 1990). Over longer periods however, these different orographic rates of increase would approach a mean value.

The orographic increase in precipitation amounts is due to a combination of increase in precipitation intensity and duration (Barry, 1992). A study in the European Alps by Havlik (1968, in Barry, 1992) showed that days with large precipitation amounts at lower altitudes account for most of the precipitation at higher altitudes, if one includes the orographic effect. Thus, for that area only a very small amount of the excess precipitation occurs on days with little or no precipitation at low altitudes. Given the proximity of the coastal plain to BL (order of 10 km) and the comparatively low altitude of the Long Range mountains versus the Alps, a similar relationship would be expected in our study area.

A mean annual precipitation amount of 2700 mm for BL would yield average snowfall amounts of at least 1000 mm SWE per year, assuming a similar snowfall/rainfall (S/R) ratio to that on the coastal plain (0.33). However, given that this ratio is essentially due to air temperature, one would expect this ratio to be larger at the lower temperatures encountered. Indeed, Davis *et al.* (1999) found that the relationship between the snowfall amounts and temperature is negative for Newfoundland during the entire snow season on average, signifying that lower temperatures lead to higher S/R ratios, if total precipitation is held constant. However, a mean January temperature of -12°C was presented as the lower threshold of negative S/R relationships with temperature. With a mean January-February temperature of approximately -12°C (Table 3.3), BL would near this threshold, possibly resulting in a generally reduced snowfall amount for winters of above average temperature, like the rest of the island. This threshold temperature, i.e. between positive and negative relationships for Snow/Rain ratios versus temperature, may seem low, but it accounts for many precipitation events which occur at temperatures well above the monthly mean. On the other hand, the representativeness of these findings applied specifically for Newfoundland, instead of to Canada as a whole, may be questionable in light of higher winter temperatures on the Island in comparison with most of the mainland.

On time scales of one day or less, snowfall can occur with temperatures above 0°C. However, snowfall usually occurs when screen air temperature is below approximately 1.5°C, although it is possible with temperatures as high as 4°C (Barry, 1992; Barry and Chorley, 1998). However, a better indicator is the freezing level, which is the height above the ground at which the temperature falls to 0°C. Snowfalls are rare when the freezing level is higher than 300 m above the local surface, while a local height of 250 m can be considered as the 50/50 probability threshold between snow and rain (Barry, 1992; Barry and Chorley, 1998). At BL, where the average lapse rate during the snow accumulation season is 5.0°C/km, the surface air temperature of the 50/50 probability between snow and rain would therefore be at 1.25°C, assuming the freezing level to be 250 m above the surface. If a daily mean surface temperature was used to discriminate between snowfall and rainfall, a temperature of 0°C would, however, be more appropriate because such an average temperature would generally still have a large portion of the day with a temperature considerably above this threshold.

## 5.2 Methodology

Unlike the large majority of precipitation data from the coastal plain stations, the winter precipitation data at BL obtained from the Fisher and Porter gauge do not separate snowfall from rainfall. The measurements taken from August 1996 to June 1999 are simply total precipitation. In order to separate the valid BL data into daily snowfall and rainfall, the BL estimated historical air temperature data set and the Stephenville Upper Air (SUA) data were used. Of particular interest are the daily average heights of the freezing level in the free air, as discussed in section 5.1, which were derived from the SUA data. These were accomplished using linear interpolation between the levels where the air temperature crosses the 0°C isotherm. The vertical distance between the base and the top of the layer in which 0°C is encountered was used to determine the height of the freezing level.

The method to distinguish between snowfall and rainfall at BL was therefore as follows:

1) if the mean daily estimated (or measured) air temperature ($T_{mean}$) at BL was above 2°C, the precipitation during that day is assumed to have occurred as rain.

2) If $T_{mean}$ was below -2°C at BL, then the daily precipitation is assumed to have occurred as snow.

3) Between $T_{mean}$ values of -2°C and 2°C, the mean daily SUA freezing level (for the 830 and 2030 NST soundings) was used, with 200, 225, 275 and 300 m above the BL height [or 980, 1005, 1055 and 1080 m a.s.l.] being the thresholds. If the freezing

height was below 980 m a.s.l., daily precipitation was assumed to be only snow, while if it was above the threshold of 1080 m a.s.l., only rainfall amounts were recorded.

4) If the freezing level was between 1005 and 1055 m a.s.l., the daily precipitation amount was considered to be equally partitioned between rain and snow.

5) Furthermore, on days with freezing levels between 980-1005 m a.s.l. and between 1055-1080 m a.s.l., the precipitation was assumed to fall as 75% snowfall and 75% rainfall respectively.

6) On days where the SUA freezing levels were missing, threshold BL $T_{mean}$ values were used instead, with -2°C, -1°C, 1°C and 2°C being the thresholds between the snowfall rainfall proportions equivalent to those used for the freezing levels. The model does not take into account other forms of freezing precipitation, such as freezing rain or drizzle and ice pellets, or hail.

As mentioned in chapter 2 (section 2.2.2), the RH precipitation record experienced a site and instrumentation change on 1 April 1993 (JD 91). The overlap of several years between the CH data with both the RH sites (CH/RH from 1982 to 1993 and CH/RHCS from 1993-1999) offered the possibility of comparing the monthly precipitation ratios between CH and both the RH sites. Through a T-test of the mean monthly ratios of both RH sites with the CH record, it was found that the significance was very low indicating that the two mean ratios are not equal. In other words, the site change with the accompanying change in measurement method (Nipher to Fischer and Porter) had a significant effect on the RH precipitation record. A mean precipitation ratio of 0.97 was found for CH/RH, while a mean ratio of 1.07 was found for CH/RHCS. This translates into a RHCS/RH ratio of 0.91 for this 17 year period, indicating that RHCS precipitation values have to be divided by 0.91 in order for them to be comparable to the original RH series. An increase of the daily RHCS precipitation values by an equivalent of 10% was therefore carried out from 1 April 1993 for it to be comparable to the much longer original site record.

The next step was to estimate wind-induced undercatch for both DH and BL in order to attain more accurate snowfall and total precipitation records for the coastal plain and the summit plateau. The daily total precipitation measured with a Nipher Gauge at DH was corrected for snowfall wind-induced undercatch with equation 5.1. In order to carry out the corrections, the daily 10 m wind speed values were reduced to 2 m above the surface, the gauge's orifice height, assuming a neutral wind profile, using equation 4.1 and the necessary coefficients described in section 4.1/4.2. Because the precipitation corrections only apply to snowfall, the DH Nipher gauge daily snowfall SWE record was isolated by subtracting total rainfall from total precipitation. The undercatch equation (equation 5.1) was utilized up to a daily average wind speed of 7.5 m/s (27 km/h), which corresponds to the 60.4% catch ratio wind speed, the assumed upper limit of the applicability of the equation as discussed in section 5.1. Above this wind speed, the catch ratio of 60.4% was applied instead of the equation output. An analogous method, i.e with 7.5 m/s as the maximum wind speed, was applied to the BL record on days where snowfall is assumed, during periods when the Fisher and Porter gauge was working properly. In this case, the estimated BL daily mean wind speed values, corrected to 2 m, were applied using equation 5.2 to determine the undercatch ratio. The augmented daily DH and BL precipitation records are assumed to be the "true" precipitation values for the coastal plain and the summit plateau respectively.

Total precipitation ratios between the three coastal plain stations and BL were constructed and served as the means to estimate historical BL precipitation. The existing real precipitation record for BL (September 1993 to June 1999), including the tipping bucket rain gauge data for the entire year, was compared to the records from the three coastal plain stations (DH, RH and CH) in different combinations to define total precipitation equations. Due to the different climatological days of the stations involved, the analyses were carried out on an event basis. An event was defined as a string of days in which precipitation occurs at one or more stations considered in each combination. A precipitation event ends when no precipitation is recorded at all the stations in question. Events so defined can vary from one to over ten consecutive days. These strings of days frequently include some days with no precipitation on either BL or the coastal plain. The comparisons were not separated into seasonally or monthly periods due to the lack of valid winter precipitation data at BL (see section 2.2.1) and they were carried out for the entire year. Only events which had total precipitation values above 3.0 mm at all stations concerned were analyzed, to reduce the wide range of ratios that would occur with small precipitation amounts. The different combinations of coastal plain predictors were ranked according to the standard error (std err) of the estimated values.

The predictive equations were then compared with one another using their output of estimated Big Level values for the period between January 1983 to June 1996, a period in which all the combination predictors could be used concurrently. The mean total precipitation value from all the estimated total BL precipitation values was used as the benchmark to harmonize the estimates from different equations. The ratios of each equation were thus increased or decreased in such a manner that the total estimated precipitation values for BL were all equal, no matter which equation was used. Given the fragmented nature of the precipitation data used to construct the predictive equations and the notoriously low accuracy of precipitation estimates, this adjustment procedure was judged necessary in order for the estimated BL precipitation record to be comparable among years which for which different predictive equations were used.

The historical BL precipitation record, i.e daily rainfall and snowfall values, was then reconstructed from October 1962 to June 1999 using the corrected precipitation ratios. The resulting estimated daily historical precipitation values for BL were compiled into monthly and seasonal totals for snowfall, rainfall and total precipitation. Total monthly and seasonal values for the study period were calculated, and the seasonal records (i.e. for October, the snow accumulation season and the snowmelt season) were checked for trends.

## 5.3 Results and Discussion

### 5.3.1 The Precipitation Models

The precipitation ratios between BL and the coastal plain were on a few occasions very large (> 10:1). It was found that the events in question were almost exclusively days on which snowfall should have occurred using the identification method presented in the methodology. Events with BL/coastal plain ratios above 10:1 clearly highlight problems with measuring snowfall at BL and these events were excluded from the analysis. Further quality control checks were carried out on the events of each combination of predictive equations in order to weed out other potentially erroneous data. The distribution of the ratios from all the individual precipitation events was observed to follow a Poisson distribution. An upper boundary control limit was established for each combination of coastal plain stations using the following equation (Ross, 2000):

$$\text{Upper control limit} = \text{mean ratio} + 3(\text{mean ratio})^{0.5} \qquad \text{Equation 5.3}$$

The calculated upper control values varied from 5.78 to 8.74, resulting in a few other events being discarded, clearly all well above the other ratios. Consequently, the initial few months of seemingly valid winter precipitation data are in the final analysis reduced to 70 days, accounting for only approximately 15% of all the events used during the period of precipitation measurement at BL (1993-1999). Snowfall is unavoidably under-represented in the building of the predictive equations, a situation that is the result of the harsh plateau environment hampering the proper measurement of winter precipitation. The preparation of the ratios results yielded four equations that can estimate BL precipitation as shown in Table 5.1. Three particular combinations of coastal plain stations (DH and RH; DH and CH; DH, CH and RH) were found not to add any precision to the prediction and were discarded. The standard errors of the predicted values for the remaining combinations are very large as initially suspected; however the r squared values from a regression analysis between the coastal stations and BL are high (0.5 to 0.8), indicating that a precipitation relationship between the stations does exist on an event basis.

Table 5.1 exhibits the equations used to estimate BL historical precipitation. The average event length used in the construction of the equations was five days. Although these equations were subsequently applied on a daily time scale, the accuracy of the relationships would be better if longer scales were considered, as is ultimately the case in the ensuing historical precipitation analysis. The calculated ratios or empirical predictive equations were ranked according to the standard error of their estimations. The best predictive equation (ranked no. 1) utilizes CH and RH data and can be read from the table as having the form of BL = 0.59(CH daily total precipitation in mm) + 1.29 (RH daily total precipitation in mm). Such an equation has the virtue of estimating BL precipitation even if only one of the coastal stations received measurable precipitation, and accounts for the larger amount of days with precipitation at BL, even though situations in which the estimated BL precipitation value is

Table 5.1 **Ratios used in the precipitation models.** The equations were constructed on an event basis, but they are also applicable to daily and monthly precipitation values.

| Equation rank | Independant variables | DH | CH | RH | Std err (mm) | n (events) | Total days |
|---|---|---|---|---|---|---|---|
| 1 | CH, RH | n/a | 0.59 | 1.29 | 26 | 52 | 272 |
| 2 | RH | n/a | n/a | 1.84 | 27 | 50 | 294 |
| 3 | CH | n/a | 1.97 | n/a | 28 | 81 | 397 |
| 4 | DH | 1.85 | n/a | n/a | 33 | 25 | 106 |

n/a = not applicable

smaller than coastal plain values are possible. With the other equations, which use only one coastal plain predictor, a value of 0 mm of precipitation must be given to BL even though precipitation may have occurred at the other coastal plain stations. This is because the equations are calibrated to include days on which some precipitation occurred at BL but not on the coastal plain (see the definition of an "event" in section 5.2). This excess precipitation is taken into account through a larger precipitation ratio, which applies on days when precipitation does occur at that given station, which is less realistic but adequate for the purposes of this multi-decade study nevertheless. The equation using RH as a predictor is probably slightly superior to the one using CH, followed by the equation utilizing only DH, as would be expected due to the greater distance involved with the last station. The equations were thus used in the order: 1) CH + RH, 2) RH, 3) CH and 4) DH, according to coastal plain data availability. Furthermore, the equations predict that, for the entire year, Big Level receives on average 1.9 times the quantity of precipitation which falls on the coastal plain. This average ratio is lower than the initially expected value of 2.3 : 1 presented in section 5.1 from the Solomon *et al.*(1968) study but it is nevertheless a reasonable estimate given the height difference between BL and the coastal plain and the amount of precipitation received on the lower sites, leading to an annual orographic increase in precipitation of approximately 150 mm / 100 m height.

The comparison of the different equations with their estimated BL precipitation values for the 1983-1996 period resulted in an increase in the ratios of equations ranked no. 1, 2 and 3 by factors of 1.09, 1.01 and 1.03 respectively while equation number 4 was reduced by a factor of 0.90 (DH equation). The equations in Table 5.1 include these adjustments. The BL record based on DH has the most interannual variability. In fact, the standard deviation of the annual total precipitation values is much higher for the record using DH (570 mm) than for the record derived from the other three methods (from RH : 430, from CH : 420 and from CH + RH : 420 mm) . There seems to be a tendency for overestimation of precipitation using the DH method in wetter years and an underestimation during drier years, although some exceptions apply such as in 1993. Furthermore, a large proportion of the precipitation at DH during the BL snow season occurs as snowfall making the wind undercatch correction a significant factor in BL's estimated precipitation record when using DH as a predictor. The amount of snowfall at DH under high wind conditions (> 7.5 m/s) falling during the entire year was much higher prior to 1973 (mean of 200 cm / year) than it was for the calibration period of 1983-1996 used to harmonize the different BL estimation methods (mean of 89 cm / year). This signals that the historical BL precipitation record must be used with caution and the change of method between the record up to 1972 (using DH as a predictor) and from 1973 onward (using mostly RH and RH + CH after 1982) must be kept in mind when interpreting the results.

### 5.3.2 Big Level Historical Precipitation Record

The reconstructed daily precipitation record was compiled into monthly, seasonal (snow accumulation season, snowmelt season) and yearly totals and means. Table 5.2 and Figure 5.2 show the monthly mean values on BL for the October 1962 to June 1999 period. The mean annual precipitation value on BL is determined to be 2610 mm during this period, which is virtually identical to the value of 2600 mm estimated for the Big Level area by the Solomon *et al.* (1968) method, and close to the 2700 mm that would be estimated for the BL summit area using the same equation. Due to the method used, the BL value is approximately 1.9 times as high as the annual coastal plain precipitation totals as expected (DH corrected for wind undercatch data set averages 1310 mm per year (1962-1996), RH averages 1305 mm per year (1972-1999) and CH averages 1200 mm per year (1982-1998)). Of the annual precipitation falling on the plateau, an estimated 53% falls as snow (1380 mm) and the balance of 1230 mm falls as rain. The "wettest" months occur in winter from December to February, when 94% of the precipitation falls as snow on average. Furthermore, the months of November through April are characterized by more snowfall than rainfall, which agrees with the period defined in section 3.3.3 as the snow accumulation season. In contrast, less than 50 mm WE of snow normally falls in the warm months of June to September in an average year; the little snowfall that does occur is normally confined to early June and late September and not a single snowfall event was reconstructed for July and August during the historical period. The standard deviations of monthly total precipitation

Table 5.2 **Big Level estimated monthly mean precipitation values.** There are no missing monthly values in the record, which covers the period from October 1962 to June 1999.

| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Total Precipitation (mm)** | | | | | | | | | | | | | |
| Mean | 290 | 270 | 220 | 140 | 160 | 190 | 190 | 210 | 200 | 220 | 240 | 280 | 2610 |
| Maximum monthly | 600 | 910 | 590 | 570 | 380 | 470 | 390 | 470 | 530 | 410 | 540 | 890 | |
| Minimum monthly | 90 | 60 | 60 | 40 | 30 | 60 | 60 | 90 | 90 | 100 | 80 | 130 | |
| Standard Deviation | 110 | 220 | 130 | 90 | 80 | 90 | 80 | 90 | 80 | 80 | 90 | 160 | |
| **Snowfall (mm WE)** | | | | | | | | | | | | | |
| Mean | 270 | 260 | 200 | 110 | 40 | 0 | 0 | 0 | 10 | 60 | 160 | 260 | 1380 |
| Maximum monthly | 600 | 910 | 590 | 570 | 170 | 50 | 0 | 0 | 120 | 130 | 380 | 890 | |
| Minimum monthly | 90 | 60 | 50 | 10 | 0 | 0 | 0 | 0 | 0 | 0 | 20 | 80 | |
| Standard Deviation | 120 | 220 | 130 | 90 | 40 | 10 | 0 | 0 | 20 | 40 | 70 | 160 | |
| **Rainfall (mm)** | | | | | | | | | | | | | |
| Mean | 20 | 10 | 20 | 30 | 120 | 190 | 190 | 210 | 200 | 160 | 80 | 20 | 1230 |
| Maximum monthly | 130 | 90 | 100 | 180 | 320 | 420 | 390 | 470 | 530 | 320 | 280 | 130 | |
| Minimum monthly | 0 | 0 | 0 | 0 | 10 | 60 | 60 | 90 | 80 | 50 | 0 | 0 | |
| Standard Deviation | 30 | 20 | 20 | 40 | 70 | 80 | 80 | 90 | 80 | 70 | 70 | 30 | |

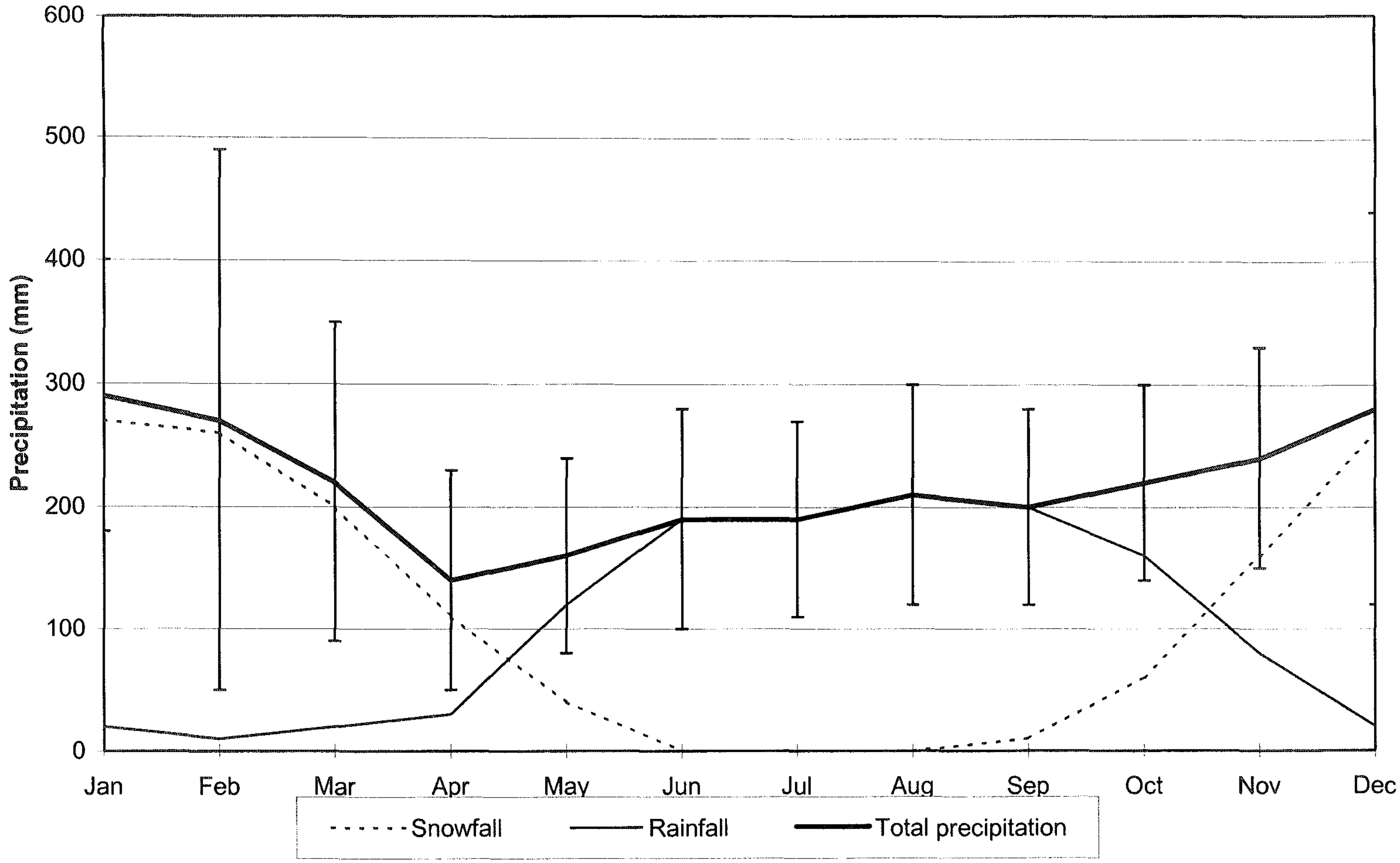


Figure 5.2 **BL estimated mean monthly rainfall, snowfall and total precipitation.** Period from October 1962 to June 1999 with no missing months. Bars show the standard deviation of the montly total precipitation values.

values are approximately 80 to 90 mm from April to November, but they are higher during the December through March period (average standard deviation for the year: 110 mm). A somewhat higher standard deviation may be expected during these months due to higher precipitation amounts, and the largest value in February (standard deviation : 220 mm) could be explained by the variable dates of freezing of the Gulf of St. Lawrence, which happens in mid-winter in most years. The month of February also has the most variable wind speed (Table 4.3) along with generally higher values for the other winter months with suggests that airmasses with very different characteristics can occupy the region: which include cold and relatively calm conditions when arctic high pressure systems dominate, or windy and wet (or snowy) conditions which accompany cyclonic systems.

The analysis of snow season precipitation time series begins with an examination of the month of October from 1962 to 1998. An estimated average of 220 mm of precipitation falls during this month with a standard deviation of 80 mm. Approximately 74% of this precipitation occurs as rainfall on average, signaling that although this month receives notable amounts of snowfall, it is unlikely to lead to large snowcover amounts in most years. Figure 5.3 indicates the estimated snowfall values during the historical period. The monthly average is 60 mm (WE) with a large standard deviation of 35 mm highlighting the substantial year to year variability (CV = 0.6). The largest October snowfall amount is believed to have occurred recently in 1997 (130 mm). Several other years, such as 1993, 1976 and 1969 would have had only slightly lower amounts. However, only three Octobers had more estimated snowfall then rainfall, occurring in 1964, 1969 and 1997. On the other hand, most

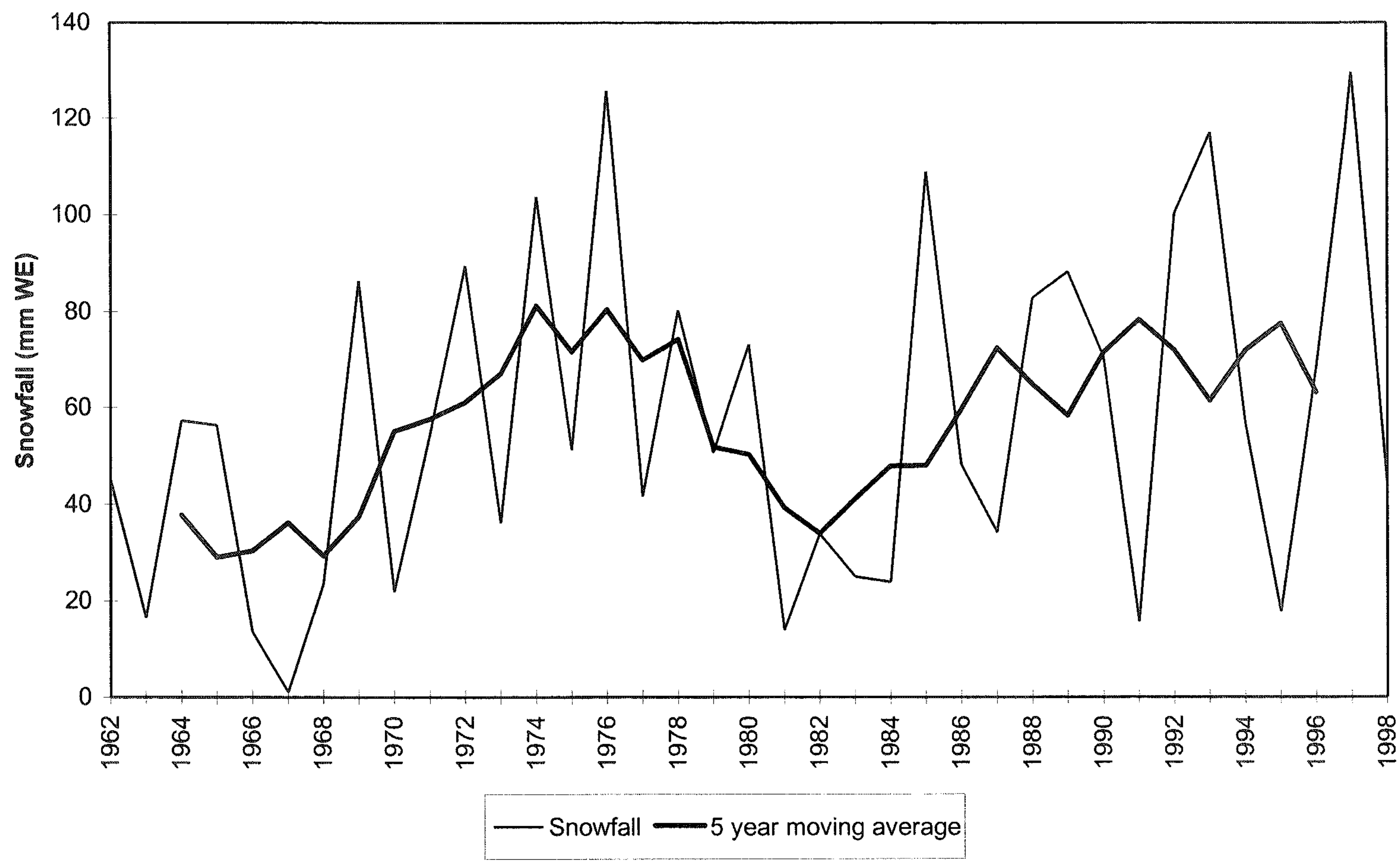


Figure 5.3 **Big Level estimated October snowfall.** Period from 1962 to 1998.

of the 1960s and the early 1980s had relatively little October snowfall (1967 receiving virtually none). Finally, no significant linear precipitation trends were found for this month.

The BL snow accumulation season (SAS) (November - April) is dominated by precipitation falling as snow, with only 13% of its precipitation on average falling as rain. An average of 1440 mm of precipitation falls during the SAS with a standard deviation of 240 mm. This amount exceeds the average precipitation amount at DH by 580 mm. Figure 5.4 illustrates the SAS total precipitation record at BL and at the DH and RH coastal stations. The reconstructed BL record is noteworthy in the 1960s and early 70s and again in the late 1970s and early 80s because of the high values in these periods. The largest seasonal precipitation value was in 1964 with an estimated 1800 mm. A secondary peak occurs in 1982 with an estimated 1750 mm with other higher values in 1967, 1968 and 1971. In contrast, the lowest value of the historical period was in 1975, with 900 mm of precipitation. The interannual variability becomes much less pronounced from the early 1980s and the record has been very stable ever since. Not surprisingly the total precipitation record is dominated by the snowfall record during these months.

The shape of the snowfall curve presented in Figure 5.5 is similar to the total precipitation curve with a mean estimated value of 1260 mm and maximums in 1964, 1967 and 1968. The minimum estimated snow accumulation season (SAS) snowfall value at BL was in 1983, with 660 mm of snow (WE), with other less pronounced lows in 1975 (770 mm) and 1998 (800 mm). In comparison to the total precipitation record, the snowy period from the late 1970s to the early 1980s occurred approximately two years earlier and is less

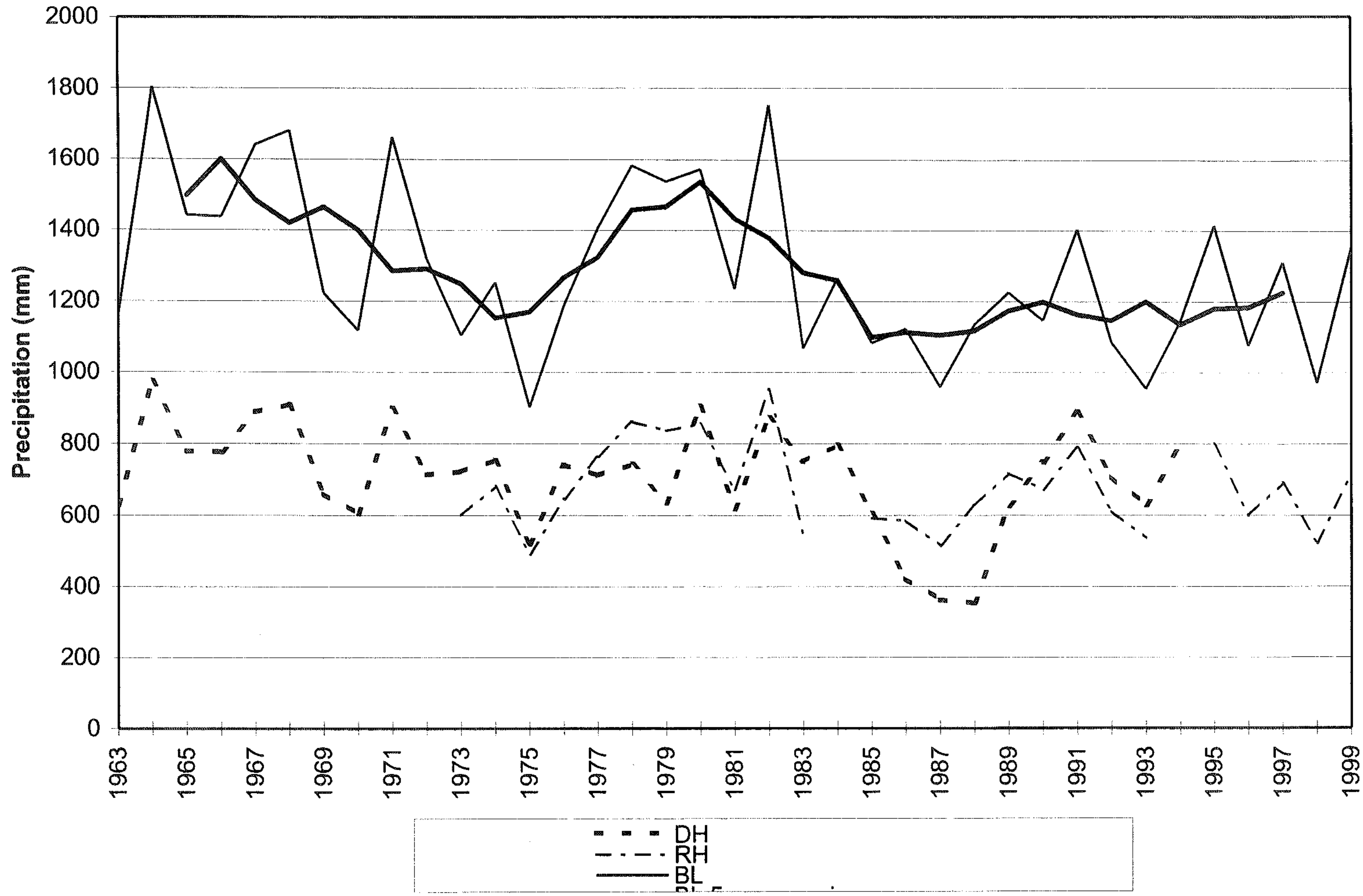


Figure 5.4 **Big Level and coastal plain (DH and RH) estimated snow accumulation season total precipitation.** Period from 1963 to 1999.

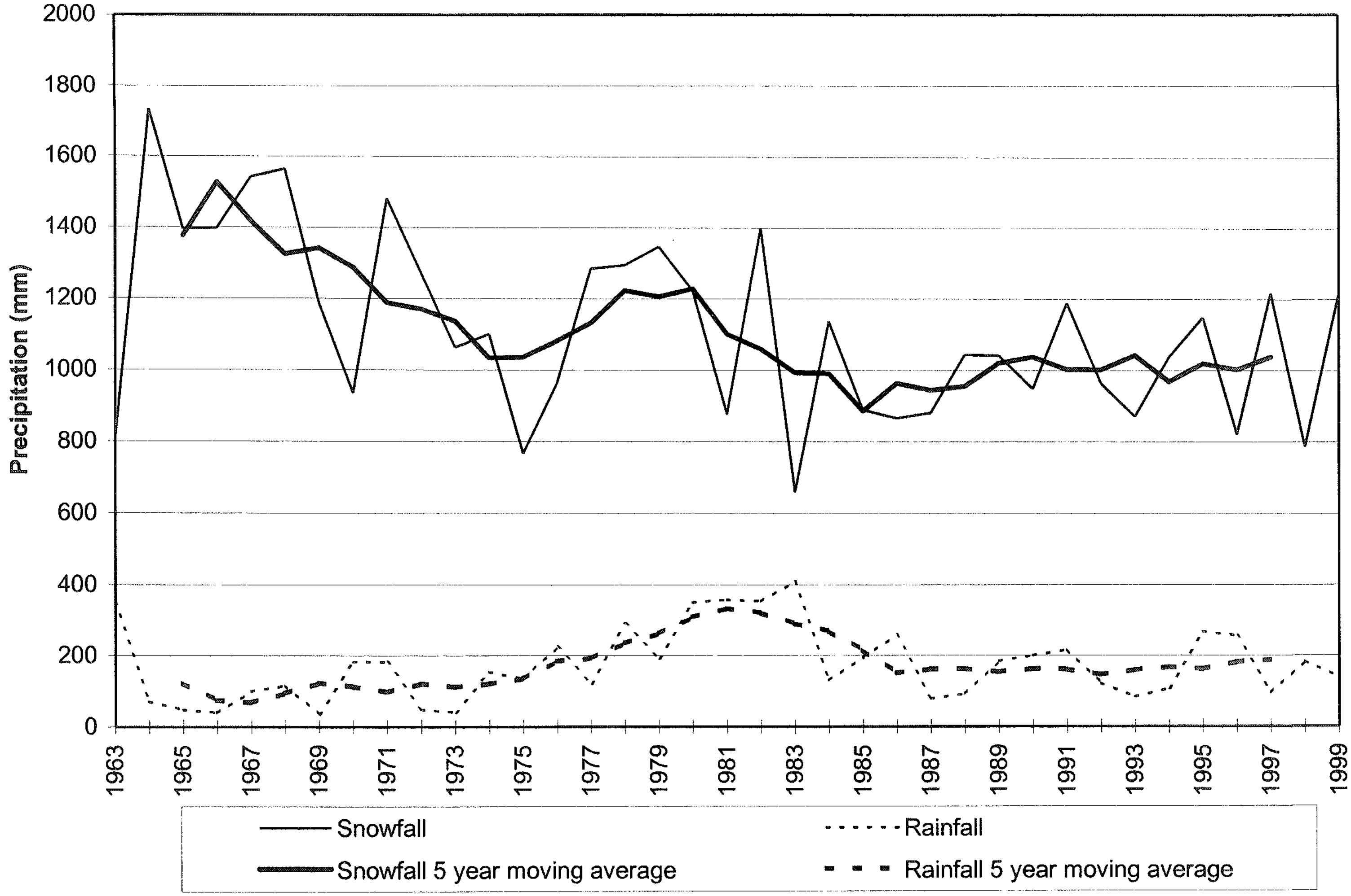


Figure 5.5 **Big Level estimated snow accumulation season snowfall and rainfall.**
Period from 1963 to 1999.

pronounced. Therefore, the complement is due to rainfall which is also shown on Figure 5.5. In fact, the 1980-1983 seasons were the rainiest of the estimated record, being well above the full period average of 170 mm. The 1983 SAS was perhaps the most outstanding, experiencing the highest rainfall and the lowest snowfall values in the same season, a phenomenon which occurred throughout the island during this season. Consequently, 38% of the total precipitation occurred as rain during the SAS at BL with, most likely, implications for the state of the snowpack. Habitually, less then 23% of the precipitation falls as rain on BL during the SAS (mean of 14% with a standard deviation of 8%), down to a low of 3-7 % during the mid-1960s.

A significant negative linear trend (95% level) exists for both total precipitation and snowfall during the SAS for the 1963-1999 period mainly due to the high snowfall values observed during the 1960s. A linear trend of -10.8 mm (WE) / year exists for snowfall with a resulting -9.2 mm / year trend for total precipitation. Indeed, the 1960s decade was known as an exceptionally snowy one in the context of the past few decades. The observed general negative trends are not apparent beyond the early 1980s, however, due to the stability of the record beyond this point. One must keep in mind that part of this stability of the BL record may be due to the estimation method, i.e. the RH and CH predictive equation was used for virtually every day after December 1982. However, the results for the RH method and the CH + RH method are virtually identical, and a method discontinuity problem could only be minor in this case.

The analysis of the snow season is completed by examining the snowmelt season

(SMS) (May -June) precipitation record. The average estimated total precipitation at BL for the 1963-1999 during the SMS is 350 mm with a standard deviation of 130 mm, compared with a coastal plain mean precipitation value of 181 at DH. The seasonal record is presented in Figure 5.6. From Figure 5.7 one can recognize that the total precipitation record in Figure 5.6 is largely determined by rainfall with an average of 89% of precipitation falling as rain during the SMS at BL. An estimated mean of 310 mm of rain is received during this two month period, with the balance of 50 mm falling as snow on average. The 1984 and 1981 seasons stand out with very large rainfall values on BL, being estimated at 600 and 530 mm respectively. A more recent large estimated rainfall amount fell in 1992 (480 mm) during this season. Concerning snowfall, few snowmelt seasons had relatively large amounts of snowfall on BL, with perhaps the exceptions of 1969 (140 mm WE), 1972 (170 mm), 1977 (120 mm) and 1996 (100 mm). During these types of season, one could expect the snow accumulation season, which normally ends near the end of April, to be prolonged by a few days to up to a few weeks, but ending before the latter part of May on the plateau.

A significant positive trend for rainfall during the SMS is noticeable, with a slope of 3.6 mm per year between 1963-1999. Concerning total precipitation and rainfall, a decadal cycle mostly driven by rainfall is noticeable from their 5 year moving average lines, with highs in 1968-71, 1979-85 and 1991-97 and lows in 1963-67, 1972-1978, 1986-90 and at the end of the record in 1999. No other clear decadal cycles exist for any other precipitation variables during the three stages of the snow season.

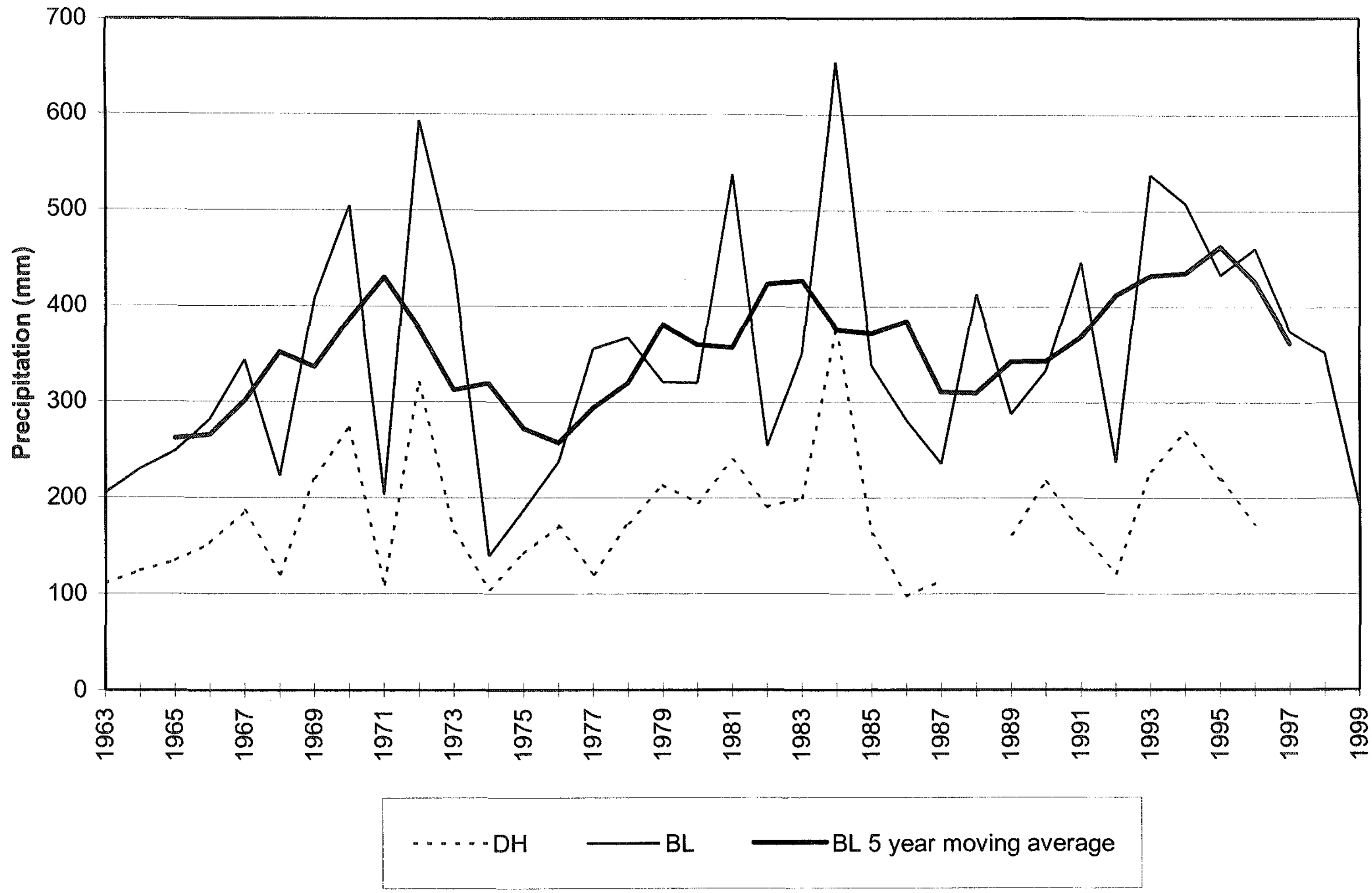


Figure 5.6 **Big Level estimated and coastal plain (DH) snowmelt season total precipitation.** Period from 1963 to 1999.

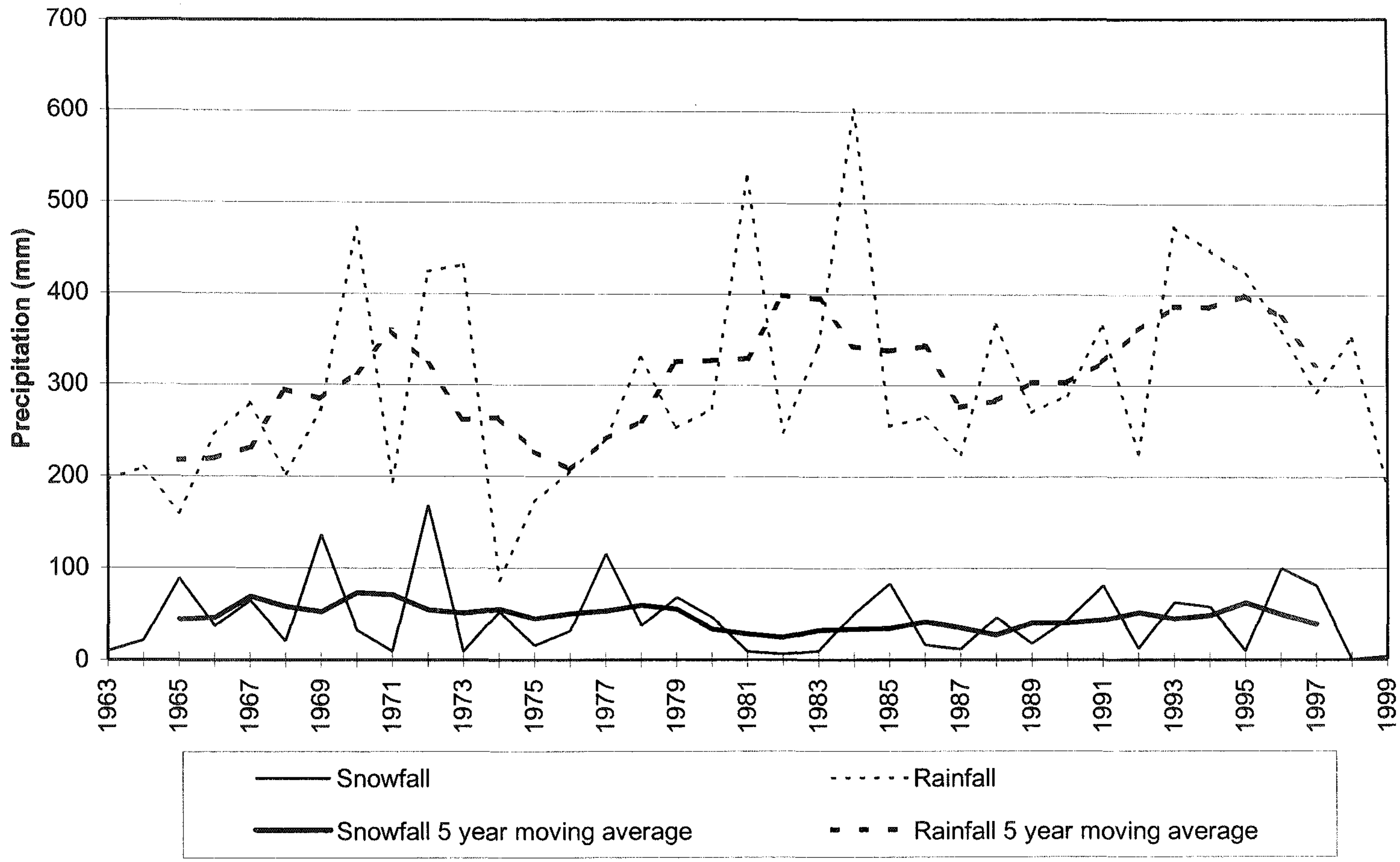


Figure 5.7 **Big Level estimated snowmelt season snowfall and rainfall.** Period from 1963 to 1999.

# 6 HISTORICAL BIG LEVEL SNOWCOVER

Previous chapters have treated meteorological variables (temperature, wind speed and precipitation) that are important in determining the state of a snowcover. Indeed, the characteristics of snowfalls and snow on the ground (SOG) are highly responsive to atmospheric conditions. The first section of this chapter explores critical concepts that are necessary to model a snowpack from the time it first accumulates on the ground up to the spring (or summer) melt. General characteristics of the snowcover in a mid-latitude site at various altitudes, with Gros Morne National Park considered in particular, are also discussed. This is followed by the methodology and the model results with a concurrent discussion.

## 6.1 Background

### 6.1.1 Snowcover statistics

The dates of appearance and disappearance of the snowcover, and the implicit snowcover duration they define are important to both human activities and ecosystems. However, one of the problems with determining this variable is that snowcover often appears and disappears several times during a season, especially near the beginning and end (McKay and Gray, 1981). It therefore becomes crucial to assign a methodology and apply it consistently from year to year. Some authors have defined the beginning (end) of the snowcover season as the first (last) day with 2 cm of snow on the ground or more, ignoring the snow free periods between (e.g. Brown and Braaten, 1998). Since SOG is of such large importance to the local climatology of a site, the notion of a continuous snow cover period has added merit. McKay and Thompson (1968) defined the date of formation of the snowcover as the first day of a period in which the snowcover is present for at least seven days. This logic can be extended to define the end of snowcover period as the last day with snowcover in which snow on the ground is absent for at least seven days following it. However, periods of several days without snow inside the continuous snowcover season need to be given special attention.

Other critical snow statistics include snow depth and density which can be combined to yield the snow water equivalent (SWE). SWE is defined as the depth of water a snowpack would produce if it was melted. The conversion is a simple one because 1 mm of water spread over an area o f 1 $m^2$ weighs 1 kg. Different snow densities can lead to very different SWE values for the same snow depth. For example, a snowpack with a depth of 10 cm can have a SWE value of 10 mm, 20 mm or even 40 mm or greater with densities of 100 kg $m^{-3}$, 200 kg $m^{-3}$ and 400 kg $m^{-3}$, or more, respectively.

Depth, density and SWE values can be expressed as means, medians, normals, departures, variances, frequencies, rates, extremes or duration for a site in a given time period (usually several years) (Goodison *et al.*, 1981). Maximum depth or SWE is also an important variable influencing water stored in the snowpack, insolation of the ground and vegetation, and implications for wildlife.

### 6.1.2 Snow Accumulation

Snow accumulation is normally measured in a straightforward manner, in which the precipitation in the form of snow is measured in a gauge or measured as a depth. However, melt, redistribution and sublimation can occur on the day of accumulation, altering the depth and density of snow present on the ground at any given site.

A mid-latitude seasonal snowcover is mainly the product of discrete winter storms (Male and Gray, 1981). Snowier winters tend to have more days with snowfall, but also larger snowfall events, as found for the western United States (Serreze *et al.*, 2001). The opposite is true for winters with less snow.

Snow accumulations at any given site are recognized as varying because of the amount of snowfall and also because of winds and topographic factors. In a windy and exposed environment, snow will preferentially accumulate around protruding objects such as boulders or shrubs and in topographic depressions. In areas that receive generally lower frequency of heavy snowfalls, such as the highlands of Scotland, wind-scouring can occur leaving behind large snow free areas on plateaus, and deep drifts on the slopes (Pomeroy, 1991). However, when large amounts of snow are regularly received, as on Big Level (BL), scouring and topographic factors would only be significant early on in the snow season, as the small roughness elements of the plateau would be quickly smoothed out. Furthermore, this study is concerned with mesoscale variations in snowcover, which involves the evaluation of snowcover conditions on an alpine plateau, and a comparison with coastal plain conditions. Thus, microscale variations due to slope, aspect and vegetation are essentially ignored. In any event, assuming that the plateau and the coastal plain are each homogeneous, flat surfaces would lead only to a small seasonal snow accumulation error because the surface types are close to this ideal, especially if the average conditions over longer periods of time are the studied elements.

Modelling snow accumulation is often approached in a simple manner, with new daily snowfall amounts simply being added to the day's total initial depth or SWE of the snowpack (Johnstone and Louie, 1983; Carr, 1989; Brown and Braaten, 1998). However, this method does introduce an error because the newly fallen snow can undergo alterations as mentioned in the first paragraph of this section (Pomeroy and Goodison, 1997).

### 6.1.3 Snow Density

Once deposited, a snowpack undergoes substantial changes before it finally melts. If an initial snow density of 100 $kg/m^3$ can be assumed at the time of deposition, such an approximation can no longer be used for snow which has aged for weeks or months. The density of a snowpack generally increases with time due to gravity, snow grain transformation, weight of additional overlying snowfalls causing settling, and wind compaction. Further increases in density occur due to melting and rain on snow events. McKay and Gray (1981) assess the density of snow in the taiga and the boreal forest, which includes coastal western Newfoundland, as being between 170 $kg/m^3$ to 210 $kg/m^3$ during mid-winter before melt. Near the end of the snow season, the boreal forest is known to have snow densities of 350 $kg/m^3$ or more (Pomeroy and Gray, 1995). For western Newfoundland, Banfield (1991) reviewed snow course data in the Grand Lake catchment for the 1934-1990 period and found an average density of 300 $kg/m^3$ for late March, with mean values for individual measurement sites varying from 284 $kg/m^3$ to 330 $kg/m^3$. Bilello (1967) developed an empirical equation relating snow season mean temperature and mean wind speed to late snow season mean density for North America:

$$\rho = 152 - 3.1T + 19u \qquad \text{Equation 6.1}$$

where:

$\rho$ = average late season snow cover density ($kg/m^3$)
T = average air temperature for season of snow cover (°C - must be below 0°C)
u = average wind speed for season of snow cover (m/s)

Using the above equation, a value of approximately 300 $kg/m^3$ was calculated for the Grand Lake catchment, which is inside the range of observed values for that area of western Newfoundland (Banfield, 1991). With a snowcover season assumed as being from December through March, Daniel's Harbour's (DH) mean late season density for the 1962-1996 period would be 300 $kg/m^3$ (T = -6.3°C, u = 6.9 m/s) and BL's mean late season density would be 360 $kg/m^3$ (T = -10.2°C, u = 9.4 m/s). The value for BL is approximately the same as a hard wind slab, while the value for DH approximates average wind toughened snow (McKay and Gray, 1981). Both estimated values appear to be realistic, on average.

The evolution of snow density during different months has been linked linearly to time, which results in a simple way to assign densities to different periods of the snow season. Sturm and Holmgren (1998) assigned average linear relationships to three different climate classes of snow. BL resembles maritime snow, the product of a relatively warm climate, which is deep (1 to 4 m) and comprises icy layers from melt and rain on snow events. Considering 1 January as the first day of applicability of the equation, they found a slope of 1.3 kg $m^{-3}$ $d^{-1}$ with an intercept of 140 kg $m^{-3}$ representing relatively rapid compaction. In comparison, tundra snow which is found in a cold, windswept environment has an intercept of 280 kg $m^{-3}$ and less than 1/5 of the slope. Finally, the density of a melting snowpack typically ranges from 350 to 500 kg $m^{-3}$ (Pomeroy and Gray, 1995).

### 6.1.4 Snow Sublimation

Aside from settling and compaction, another important process that a snowpack undergoes outside of melt is sublimation (i.e. from the solid to the gaseous phase). When a vapour pressure gradient exists decreasing away from the snow surface, and the snowpack receives energy through radiation or advection of sensible heat, snow particles can vaporize. However, with a latent heat of sublimation of 2.83 MJ $m^{-2}$ per mm of SWE, it occurs at a high cost of energy relative to the amount the winter season environment can provide (Pomeroy and Gray, 1995). Another limiting factor is that snow temperature cannot exceed 0°C, which means that the maximum vapour pressure at the surface is 6.11 mb (Oke, 1987). Consequently, under mild and humid conditions that are occasionally experienced in western

Newfoundland during the snow season, the vapour pressure gradient is small if not directed toward the surface leading to condensation. On the other hand, when temperatures are well below freezing, both energy availability and small vapour pressure gradients become limiting factors. As a result, the sublimation of a snowpack is a minor component of the physical changes that occur from deposition to ablation (Barry, 1992). However, the situation changes dramatically when blowing snow is considered (see section 6.1.6).

The assessment of snowpack sublimation through detailed energy balance studies requires the measurement of vapour pressure, wind speed and temperature at multiple levels to determine the latent heat term ($Q_e$), values that are not available for most study sites (Male and Gray, 1981). Several attempts to simplify sublimation evaluation have had mixed results. The most popular method has been the use of the following equation (Male and Gray, 1981):

$$Q_e = D_e u_z (e_s - e_a) \qquad \text{Equation 6.2}$$

where:

$Q_e$ = the latent heat of sublimation in MJ $m^{-2}$ $s^{-1}$ which, when divided by 2.83 MJ $m^{-2}$ $mm^{-1}$ and multiplied by 86400 s $d^{-1}$ yields the amount of sublimation in mm $d^{-1}$ SWE
$D_e$ = bulk transfer coefficient for latent heat transfer (MJ $m^{-3}$ $mb^{-1}$)
$u_z$ = wind speed at a reference height, usually 1 to 2 m above the surface
$e_s$ = vapour pressure of the snow surface (mb)
$e_a$ = vapour pressure of the air, at reference height (mb).

Male and Gray (1981) report several $D_e$ values, which vary from 2.17 $*10^{-6}$ to 8.0 $*10^{-6}$ MJ

$m^{-3}$ $mb^{-1}$ or more. Such wide range of values highlights the uncertainties involved in such calculations because of the complex surface-atmosphere interactions. However, only the lower value of 2.17 $*10^{-6}$ published by Granger (1977) for the Prairies, was calibrated for a period in which net sublimation rather than condensation was taking place. It is specified that equation 6.2 should not be applied to periods smaller than one day.

In order to roughly estimate the magnitude of snow sublimation in the study region, equation 6.2 was applied to average monthly conditions during the entire study period (1962-1999) for DH and BL during months with snow cover. Monthly average $e_a$ was obtained for DH from Environment Canada's climate normals (1961-1990) of relative humidity (rh - average winter value of 85%) while $e_s$ was obtained by combining vapour pressure and relative humidity data (Environment Canada, 1998). In BL's case, a 90% relative humidity was assumed for all months and $e_s$ was determined from the monthly mean air temperature. Winter measurements from the Big Level Autostation over one snow season (1997 - the only one available) suggest an average relative humidity of approximately 95%, however, this value is probably high due to the rh value being slightly over 100% on many days. For both locations, snow surface temperature is assumed to be at the average air temperature for $e_s$ estimation. Wind speed data were reduced from 10 m to 1 m using equation 4.1 and the procedure outlined in that chapter. For the remaining variable, $D_e$, the values used were between 2.0 $*10^{-6}$ MJ $m^{-3}$ $mb^{-1}$ and 8.0 $*10^{-6}$ MJ $m^{-3}$ $mb^{-1}$.

The results and input variable values are shown in Table 6.1. Only the months of December to April are reported for DH and November to April for BL. Not only does

Table 6.1 **Monthly sublimation over the DH and BL snowpacks during average conditions.** Values calulated using equation 6.2. Variables are average monthly wind speed (Ws), estimated saturation vapor pressure ($e_s$) and actual air vapor pressure ($e_a$). See section 6.1.4 for further details.

**Daniel's Harbour (DH)**

| Month | Ws at 1m (m/s) | $e_s$ (mb) | $e_a$ (mb) | Sublimation ($MJ\ m^{-2}\ d^{-1}$) | Sublimation (mm/d) | Sublimation (mm/month) |
|---|---|---|---|---|---|---|
| Dec | 4.8 | 4.9 | 4.1 | 0.67 - 2.66 | 0.24 - 0.94 | 7 - 28 |
| Jan | 5.0 | 3.9 | 3.3 | 0.52 - 2.07 | 0.18 - 0.73 | 6 - 23 |
| Feb | 4.4 | 3.8 | 3.1 | 0.54 - 2.15 | 0.19 - 0.76 | 6 - 24 |
| Mar | 4.3 | 4.7 | 3.9 | 0.59 - 2.36 | 0.21 - 0.83 | 6 - 23 |
| Apr | 3.9 | 6.1 | 5.5 | 0.40 - 1.61 | 0.14 - 0.57 | 4 - 18 |

**Big Level (BL)**

| Month | Ws at 1m (m/s) | $e_s$ (mb) | $e_a$ (mb) | Sublimation ($MJ\ m^{-2}\ d^{-1}$) | Sublimation (mm/d) | Sublimation (mm/month) |
|---|---|---|---|---|---|---|
| Nov | 6.3 | 4.8 | 4.3 | 0.52 - 2.09 | 0.18 - 0.74 | 6 - 22 |
| Dec | 6.5 | 3.3 | 3.0 | 0.37 - 1.48 | 0.13 - 0.52 | 4 - 16 |
| Jan | 6.7 | 2.4 | 2.2 | 0.28 - 1.11 | 0.10 - 0.39 | 3 - 12 |
| Feb | 6.1 | 2.1 | 1.9 | 0.22 - 0.89 | 0.08 - 0.31 | 2 - 9 |
| Mar | 5.9 | 3.3 | 3.0 | 0.34 - 1.35 | 0.12 - 0.48 | 4 - 15 |
| Apr | 5.6 | 4.8 | 4.3 | 0.46 - 1.84 | 0.16 - 0.65 | 5 - 20 |
| May | 5.4 | 6.1 | 5.5 | 0.56 - 2.23 | 0.20 - 0.79 | 6 - 24 |

snowcover not always exist outside these periods, but $e_a$ usually exceeds the maximum snow surface saturation value ($e_s$) of 6.1 mb outside of these months, preventing significant amounts of sublimation from taking place. With higher air temperature and accompanying higher $e_s$ values, DH averages 40 - 60 mm of snowpack sublimation per snow season, if one accepts that the lower end of the range is more likely (the higher end of the range applies to situations of condensation more so than sublimation). Despite the snow season being two months longer and higher wind speeds, BL's snowpack sublimation is also approximately 40 -60 mm per year, due to lower temperatures and a higher relative humidity. Although these values are only approximations, they agree well with snowpack sublimation values presented in Gray and Prowse (1993). Doty and Johnston (1969) found an average sublimation value during daytime hours from open sites in Utah of 0.30 mm $d^{-1}$, while Granger (1977) found and average sublimation value of 0.15 mm $d^{-1}$ during the snowmelt season on a Prairie site. Regardless of the exact sublimation value occurring at BL, it is obvious that values of snowpack sublimation loss on the order of 50 mm WE per season are only a small fraction of the total snowpack considering that average snowfall on the plateau is on the order of 1380 mm SWE per year on average (Table 5.2). However, sublimation of blowing snow is most likely large on a relatively flat and very windy site, due mainly to a huge increase in the surface to volume ratio in the case of suspended snow particles compared to a snowcover surface (see section 6.1.6).

### 6.1.5 Snowdrifting

The wind profile near the surface is heavily dependant on the surface features. Obstacles like hills, trees, or at a smaller scale shrubs and rocks tend to slow down the wind near the surface. Within the first few metres above the surface, wind speeds and direction are highly variable making exposed snow covered areas subject to erosion and other more protected areas subject to deposition. Thus, open areas with few obstacles will experience more snow drifting. Furthermore, because snowcovers envelop most of the smaller features and have small surface roughness values wind speeds near snow surfaces are relatively high. This inherent quality makes snow prone to drifting (Kind, 1981; Pomeroy and Goodison, 1997).

The vertical wind velocity gradient causes shear stresses between the various heights and is greatest at the surface. The force applied to the surface by this stress is what makes snow transport possible. In fact, drifting occurs when the shear stress on the snow surface exceeds the shear strength of the snow particles caused by the size, the shape and the weight of the snow grains as well as by the cohesion forces between these grains. The wind speed where this becomes possible is referred as the threshold velocity (Kind, 1981; Pomeroy and Goodison, 1997).

New and dry snow is most easily displaced because the cohesion forces are the smallest. Threshold velocities (at a height of 10 m) for it are between 3.5 and 4.5 m/s. Therefore, most snow is transported within a day of its deposition when the air temperature is below -2°C (Verge and Williams, 1981). For dry snow of moderate hardness a value of 5.5 m/s has been suggested (Pomeroy and Goodison, 1997). Li and Pomeroy (1997) found an average threshold value of 9.9 m/s for the Canadian prairies regarding wet snow, and an average value of 7.7 m/s for dry snow, although values have ranged from 4 to 14 m/s. Moreover, although older snow is supposed to be more difficult to move because of a harder and more cohesive surface (Kind, 1981), Li and Pomeroy (1997) found very little difference between new snow and old snow, with values of 7.5 and 8.0 m/s respectively. However, they found a positive relationship between threshold velocity and air temperature. Above -25°C, threshold velocity was found to follow the following relationship:

$$u^* = f + hT + kT^2 \quad \text{(m/s)} \qquad \text{Equation 6.3}$$

with:

$u^*$ = threshold wind velocity at 10 m
f, h and k= coefficients calibrated for the Canadian prairies (16 stations)
T = air temperature

The standard error for this equation was 1.97 m/s. Like most empirical equations, this one is probably of little use to other locations without proper and time consuming calibrations. No physically based method currently exists to estimate threshold velocities for snow grains. However, given that the characteristics of a snowcover are heavily related to meteorological conditions, an equation could eventually be built when the relationships between the meteorological variables and the snowcover are better understood (Li and Pomeroy, 1997).

Drifting snow can reach maximum heights ranging from several millimetres to several hundreds of metres. However, most of the snow is contained in the first few metres of the air column. Most of the snow moves by saltation at lower wind speeds, even though creep and suspension also occur (Pomeroy and Goodison, 1997). During episodes of high winds and with a snow cover having small particle sizes, the mass of blowing snow moved by suspension may be dominant, due to the total potential thickness of the blowing snow layer, i.e. 100 metres or more (Tabler *et al.*, 1990). Concerning the total amounts of snow moved by drifting, Verge and Williams (1981) suggested that it is proportional to the cube of the wind speed above the threshold value, although more recent studies have revised this estimate upward to the fourth power (Gray and Prowse, 1993). Furthermore, several equations have been devised linking the two variables (see Tabler *et al.*, 1990). What the relationships ultimately tend to show is that the amount of snow displaced by wind can dwarf the amount of precipitation received, particularly in dry, cold and windy locations (Trabant and Clagett, 1990). Transported amounts can exceed 3000 kg per metre perpendicular to the wind direction in one hour.

Davison (1987) used a method to compare the amount of snowdrifting in Scotland from one snow season to the next and the frequency of its direction. He classified every snowfall event according to amount of snowfall, the wind speed and the direction (8 cardinal points). Snowfall amounts and wind speeds were placed into classes which were multiplied together and then compared by year and by direction (e.g. a snowfall event of 12 cm (class 3) at a wind speed of 5 m/s (class 2) from the SW results in a snow drift amount of 6 from the SW). It was also found during the course of this study that higher amounts of snowdrift lead to more and larger late lying snowbeds during the summer.

Although the methods used by authors such as Davison (1987) are useful for determining broad climatological relationships, they offer little help to the investigator looking to quantify drifting snow amounts. Pomeroy (1988) devised the Prairie Blowing Snow Model (PBSM) which calculates both drifting snow amounts and sublimation values for an open, relatively flat surface, using meteorological data from principle observing stations. This type of model was unique in its genre and is frequently discussed in the snow literature. The majority of the model presented and used in this thesis for the examination of blowing snow is based on the PBSM, as discussed by subsequent papers (e.g. Tabler *et al.* (1990); Pomeroy *et al.*, 1991).

Goodison and Pomeroy (1997) suggested that wind transport of snow can be estimated from wind speed data alone, and mention the work by Tabler *et al.* (1990) in particular. The equation they propose sums up saltation and suspension components to give total transport:

$$q = u_{10}^{4.04} / 458800 \qquad \text{equation 6.4}$$

where:

$q =$ kilograms of snow transported to the end of the fetch per metre perpendicular to the wind direction, per second (kg $m^{-1}$ $s^{-1}$)

$u_{10} =$ hourly mean 10 metre wind speed (m $s^{-1}$).

According to Pomeroy and Gray (1995), equation 6.4 is valid for fetch distances greater than one kilometre, and has a threshold wind speed of 6.5 m $s^{-1}$ (Tabler *et al.*, 1990). The equation is not very sensitive to the threshold wind speed on a monthly basis in a windy environment, due to the immense amounts of blowing snow that occur at high wind speeds, well above the threshold conditions. This equation applies to an adequate upwind supply of snow and gives best results when data of blowing snow occurrences are available (Gray and Prowse, 1993).

It is thought by some authors that a notion of equilibrium condition may develop where snowdrifting may be considered to be negligible and erosion is a function of sublimation only. Gray and Prowse (1993) offer a value of between 150 and 500 m of fetch for such a condition to develop. However, if an entire snow covered area is considered such as the Big Level plateau, and snow cannot drift on the plateau from neighbouring areas due to steep slopes, it is clear that drifting off the plateau would result in net loss to the snowcover of the plateau. Therefore, if the average snow depth of a large but limited area is considered (e.g. a few square kilometres), snowdrifting cannot be ignored.

### 6.1.6 Sublimation of Blowing Snow

As mentioned earlier, sublimation of blowing snow occurs at a rate far exceeding that of snow on the ground in areas with frequent blowing snow events such as BL. This is due principally to the fact that blowing snow particles have 300 times more surface area than snow on the ground (Goodison and Pomeroy, 1997). Sublimation of blowing snow increases exponentially with wind speed (Pomeroy and Gray, 1995). Other factors significantly affecting sublimation rates are temperature and relative humidity. The sublimation rate approximately doubles for every 10°C increase in air temperature (Tabler *et al.*, 1990). Furthermore, sublimation of blowing snow increases with decreasing relative humidity due to the associated expansion of the vapour pressure deficit.

A particularly useful relationship to calculate blowing snow sublimation is the one offered by Tabler *et al.* (1990) adapted from Pomeroy (1988), due to its methodological simplicity. It calculated the sublimation rate as a function of snow transport (q), mean air temperature ($T_{mean}$) and relative humidity (rh):

$$q_{evap} = mq - b \qquad \text{Equation 6.5}$$

where:

| | |
|---|---|
| $q_{evap}$ = | snow evaporation rate (mg $m^{-2}$ $s^{-1}$) |
| q = | total snow transport (g $m^{-1}$ $s^{-1}$) |
| m and b = | slope and intercept value of the equation, which are a function of $T_{mean}$ and rh. |

Tabler *et al.* (1990) give a few values for the m and b coefficients for different $T_{mean}$ and rh as calibrated for the Canadian Prairies. However, the m and b values presented are each for a specific temperature and precipitation value which complicates the interpolation to different values. Nevertheless, interpolation is possible due to several coefficients being offered by Tabler *et al.* (1990) for different rh values at -15°C, and for different temperatures at a 70% rh value. The rh values were converted to vapour pressure deficits, which more accurately reflect the drying power of the air. It is logical to expect that other regions should respond similarly when under the same atmospheric conditions than in the calibration region of the Canadian Prairies, because the most important factors are all taken into account.

Concerning the effect of fetch distance on sublimation rates, Tabler *et al.* (1990) and Pomeroy and Gray (1995) explain that the sublimation rate increases linearly with fetch distance, above the reference case of one km for Equation 6.4. However, it was found by Dery and Yau (2001) and Xiao *et al.*(2000) that this method overestimated sublimation, as applied to the Canadian Arctic, because blowing snow events occurring over large fetch distances, or in large amounts, would be self-limiting in terms of sublimation due to the reduction in the air's drying power, i.e. vapour pressure deficit. A further phenomenon which tends to reduce ablation rates is the reduction of ambient air temperature as energy is used to convert snow from the solid to the gaseous phase during sublimation (Xiao *et al.*, 2000). Xiao *et al.* (2000) add that our understanding of the physics of blowing snow is still rudimentary, but this should obviously not prevent us from trying to incorporate its effects in studying the evolution of snowcovers.

Using the method described above in equation 6.5, Pomeroy and Gray (1995) found average annual sublimation rates of 64 mm SWE at Regina and 28 mm at Yorkton over large fetches during the 1970-76 period, which may represent on the order of 50% of annual snowfall in these dry and windy environments (Goodison and Pomeroy, 1997).

6.1.7 Snowmelt

Snowmelt can be calculated precisely using a detailed energy balance approach. Assuming that the site is horizontally large enough to be considered homogeneous, lateral exchanges of energy and water may be considered negligible. The energy balance equation can thus be expressed by the following:

$$Q_m = Q^* + Q_h + Q_e + Q_g + Q_p - \Delta U/\Delta t \qquad \text{Equation 6.6}$$

where:

$Q_m$ = energy available for melt (MJ $m^{-2}$ $d^{-1}$)
$Q^*$ = net radiation (MJ $m^{-2}$ $d^{-1}$)
$Q_h$ = sensible heat (MJ $m^{-2}$ $d^{-1}$)
$Q_e$ = latent energy (MJ $m^{-2}$ $d^{-1}$)
$Q_g$ = ground heat (MJ $m^{-2}$ $d^{-1}$)
$Q_p$ = heat from rain (MJ $m^{-2}$ $d^{-1}$)
$\Delta U/\Delta t$ = internal energy change per unit time, U is the negative heat storage (MJ $m^{-2}$ $d^{-1}$).

Energy fluxes toward the snowpack are positive while those directed away from the snowpack are negative. The time interval assumed above is one day, but the equation can apply to any other period. Under melting conditions, $Q_h$, $Q_p$, and $Q_g$ are normally positive. Concerning $Q^*$, the shortwave portion is always positive, while the long wave portion is usually significantly positive under cloudy conditions during melting. Consequently, radiation can be a large portion of the energy available for melt. On the other hand $Q_e$ and $\Delta U/\Delta t$ can be either positive or negative, often playing secondary roles (Viessman and Lewis, 1995).

The largest factor affecting the Q* at a snow surface is the surface albedo, the fraction of incoming shortwave radiation that is reflected from the surface. The albedo of new snow is approximately 0.8-0.9, while old melting snow at the end of the snow season usually has a value of around 0.4 (U.S. Army Corps. of Engineers, 1956; Pomeroy and Goodison, 1997). This factor will tend to make snow melt occur faster near the end of the snow season because of the larger shortwave input. For more information on albedo, see Male and Gray (1981).

Ground heat, $Q_g$, is often ignored in the equation because of its negligible daily contribution to melt (Viessman and Lewis, 1995). In an area with a thick snowcover, its overall seasonal contribution is also likely to be small.

The heat provided by rainfall, $Q_p$, comes in two forms. The first is the sensible heat derived from its temperature above freezing. However, rainfall during the snowcover season usually occurs close to the freezing point, and releases a limited amount of heat. The second source of energy is the latent heat of fusion from the rainwater, which is released when the water freezes in the snowpack. This warms the snowpack to its freezing point relatively quickly compared with other sources of energy. However, under melting conditions, the contribution from the latent heat of fusion stops once the pack is isothermal at 0°C (Gray and Prowse, 1993). Furthermore, rainfall has the special property of adding mass to the SWE until the snow becomes saturated with liquid water.

The snowpack releases meltwater usually only once its liquid water holding capacity is reached. The snow is considered primed for melt when three to five percent of its mass is liquid water, which concentrates around snow grains and in pockets throughout the pack (Male and Gray, 1981). Under high melt rates, the water holding capacity can reach proportions as high as 20%. If temperatures fall again below freezing, the water must freeze before the snow temperature can drop below 0°C. Furthermore, liquid water in the snowpack has the effect of reducing the thermal quality of snow. The thermal quality is the ratio of energy needed to melt snow in comparison to ice (U.S. Army Corps of Engineers, 1956). When water is present, the thermal quality drops in proportion to the amount of liquid water present, with a value of 0.97 for three percent water content and 0.95 for five percent water content thus decreasing the amount of energy needed to melt the snowpack (Male and Gray, 1981).

Although usually considered separately, liquid water content can be part of what is called the internal energy of the snowpack, $\Delta U/\Delta t$, if ablation is being modelled rather than the melt at the top layers of the snowpack. Internal energy is proportional to the mass of the snowpack. Thick snowcovers can require a considerable amount of energy simply to increase in temperature to the freezing point throughout their depth and to satisfy the liquid water storage capacity before ablation can begin. The amount of energy required for the snowpack to reach an isothermal state at the freezing point is called its "cold content" (in kJ $m^{-2}$). The notion of cold content is very useful in simulating the delay of snow ablation, because the initial melt is often restricted to the upper layers of the pack, with the water freezing again at deeper layers, resulting in a negligible loss of total mass (Gray and Prowse, 1993). The major result is a rise in the temperature of the deeper layers of the snowcover, priming the pack for ablation at a later time. Furthermore, although it takes time for heat to diffuse through the pack, using a daily time step essentially removes this complexity for snowpacks on the order of a few metres thick or less, because this factor operates on a scale of hours at limited thicknesses. Bloschl and Kirnbauer (1991) have used a bulk approach to measuring snowpack ablation in which the snowpack is considered as a block rather than as separate layers to measure the cold content. Although the method does not simulate the cold content well, it does predict the timing of the isothermal state in an acceptable way, which is just the property that is sought when the snowpack SWE is modelled. They also used the concept of a maximum cold content value to prevent the cold content from reaching unrealistically high values when used in conjunction with a melt algorithm. For a more complete consideration of energy balances of snowpack, the reader is referred to a hydrology text such as Maidment (1993) or the Handbook of Snow (Male and Gray, 1981).

### 6.1.8 Temperature Index Method

A full consideration of the energy balance of a snowpack is the best way to calculate melt. However, the data requirements for this approach are large, and usually not available at most sites. On a given site, there is typically little more to work with than daily temperature and precipitation data, with humidity and wind speed data at one level occasionally available (Viessman and Lewis, 1995). Temperature is therefore regularly used as an index, i.e. a less accurate substitute, for the energy available for melt. The following equation outlines the degree-day approach:

$$M = a(T_a - T_b) \qquad \text{Equation 6.7}$$

where:

$M$ = melt ( in SWE, mm $d^{-1}$)
$a$ = degree-day factor (mm $°C^{-1}$ $d^{-1}$)
$T_a$ = daily mean air temperature (°C)
$T_b$ = base temperature (°C), i.e. temperature above which melting occurs.

Using degree-day factors that increase during the season to incorporate the effects of increasing solar radiation and of decreasing snow albedo can partially correct the shortcomings of the method (Rango and Martinec, 1995).

The base temperature is often assumed to be 0°C. However, days with mean temperatures slightly below the freezing point commonly have several hours of above freezing air temperatures leading to some melt. Open environments in particular often have melt during days with mean temperatures below 0°C due to the input of solar radiation. For this reason, the selected base temperature can be below freezing.

It has been widely thought that temperature index models do not predict snowmelt accurately in open environments compared to forested areas, mainly because of the wide range of solar radiation amounts that can be experienced in the open (Goodison and Pomeroy, 1997). However, in cloudy environments, such as maritime climates or in mountainous areas with large amounts of cloud cover, the accuracy of temperature index models can improve considerably (Gray and Prowse, 1993). Furthermore, although the daily values of the melt factors have large variations in open environments, the average values over monthly intervals can be considered to be reasonably consistent (Rango and Martinec, 1995).

Gray and Prowse (1993) suggest that degree-day factors of 3.5-6 mm $^{o}C^{-1}$ $d^{-1}$ are common during spring melt, with lower values for fresh snow. During mid-winter with less solar radiation and fresher snow, values closer to 2 mm $^{o}C^{-1}$ $d^{-1}$ may be considered reasonable (see Rango and Martinec, 1995). An analysis of melt occurring at DH and Rocky Harbour on the coastal plain during days without rain has yielded average melt factors of 2 mm $^{o}C^{-1}$ $d^{-1}$ for the period up to February, and approximately 3 mm $^{o}C^{-1}$ $d^{-1}$ for April. Larger values would be expected on BL during May, June and July. The WMO (1964) suggested average values of 6 mm $^{o}C^{-1}$ $d^{-1}$ for mid-latitude open sites in May and 7 mm $^{o}C^{-1}$ $d^{-1}$ for June, although smaller values on BL may be more realistic given the high cloudiness of the site.

On days with rain, the method can be adjusted to include an algorithm dealing with the amount of precipitation received, to estimate the energy supplied by the rain water. The most widely used expression was presented by the U.S. Army Corps of Engineers (1956) and is still widely referred in the literature in one form or another (Gray and Prowse, 1993; Viessman and Lewis, 1995; Brown and Braaten, 1998). The equation was designed for forested environments, but it has since been adapted to suite many environments due to its simplicity. The version presented below in the methodology (Equation 6.8) was inspired from Brown and Braaten (1998), which has two variables, the mean air temperature and the total rainfall. The principle innovative feature of their approach is the inclusion of a local calibration factor, which can be adjusted to suit many different environments.

### 6.1.9 Snowcover in the Gros Morne Area

Snowcover in western Newfoundland can be described as "stable", lasting four to six months depending on latitude and altitude (McKay and Gray, 1981). At the lowest elevations a complete snowcover is typically established in early December and disappears by late April; maximum depths are often in the vicinity of 100 cm and occur in late February in low altitude locations (McKay and Gray, 1981). Table 6.2 shows statistics for Daniel's Harbour (DH) on the coastal plain calculated from the MSC daily data set of snow on the ground (SOG) for the 1962-1996 period. The DH record becomes too patchy after 1996 to extract meaningful information from it. Other information on coastal plain snowcover is presented in section 6.3.

### 6.1.10 Alpine Snowcovers

There are relationships between snow cover depth and altitude. Due to an increase in the number and size of snowfall events with altitude (orographic effect) and a decrease in

Table 6.2 **Snow Statistics for Daniel's Harbour (1962-1996) as compiled from the daily snow on the ground (SOG) record**. The term "continuous snowcover" applies to both the spatial and temporal dimensions, see methodology for specifics.

| **Variable** | **Value** |
|---|---|
| Mean timing of maximum snow depth | Late February - early March |
| Mean maximum snow depth | 87 cm |
| Largest maximum seasonal snow depth | 166 cm (1987) |
| Smallest maximum seasonal snow depth | 20 cm (1996) |
| | |
| Mean date of continuous snowcover appearance | 10 December (JD 344) |
| Earliest date of continuous snowcover appearance | 19 November 1976 (JD 314) |
| Latest date of continuous snowcover appearance | 15 February 1970 (JD 46) |
| | |
| Mean date of continuous snowcover disappearance | 19 April (JD 109) |
| Earliest date of continuous snowcover disappearance | 20 February 1981 (JD 51) |
| Latest date of continuous snowcover disappearance | 15 May 1985 (JD 135) |
| | |
| Mean continuous snowcover duration | 131 days (over 4 months) |
| Minimum continuous snowcover duration | 73 days (1981) |
| Maximum continuous snowcover duration | 177 days (1977) |

melt and sublimation (in many cases), the general conclusion is that the depth increases with height (Pomeroy and Goodison, 1995). In fact, some investigators have found a linear relationship between seasonal snowcover depth and altitude (e.g. U.S. Army Corps of Engineers, 1956). However, other factors such as snowdrifting or topographic factors can make snow depth highly variable at the microscale, leading to the appearance of less snow at higher elevations.

Atmospheric conditions also have an impact on the differences that exist between sites separated by altitude. Caine (1975) found in Colorado that low pressure systems which generate snow storms tend to deposit relatively more snow at low altitude locations, while light accumulations which are more often driven by local factors tend to predominate at high altitude sites. This finding, combined with the fact that the length of the snow season increases with altitude, leads to the initial hypothesis that total snow accumulations would tend to be less variable as elevation increases for areas that experience considerable winter snowfall amounts. Such a relationship may also hold true for BL. However, wind redistribution may be the deciding factor, as in the Norwegian Mountains where Killingtveit and Sand (1991) found that snow depth decreased in open areas with height due to greater snow relocation at higher elevations. Furthermore, snowcovers above the tree-line in mountainous terrain (i.e. alpine snowcovers) usually undergo substantial erosion and higher densities due to wind packing (McKay and Gray, 1981).

Gray and Prowse (1993) mention that sublimation of blowing snow is a function of elevation due to increasing solar radiation and decreasing vapour pressure, although this principle does not appear to be a critical factor in moderate altitude and cloudy environments such as BL compared to high elevations in large mountains (e.g. Rockies).

High elevations in Newfoundland have amongst the largest snowfall amounts in eastern Canada because of the regional moisture source (Gulf of St. Lawrence), orographic cooling and the frequency of favourable synoptic scale weather patterns. Furthermore, in southern Canada overall, snowcover depth and duration generally increase with elevation (with micro-scale topography and vegetation held constant). This is due to a combination of more snowfall and less melt with increasing height. However, larger snow transport values due to higher wind speeds at higher elevations can partially erase the difference (Pomeroy and Goodison, 1997).

### 6.1.11 Climate Change and Snowcover

It is generally expected that higher temperatures due to climate change would cause an upward altitudinal shift of climate zones by a of approximately 170 m/ °C, due to the mean environmental lapse rate (IPCC, 2001b). Mountain ecosystems within a few hundred metres of the highest elevations are particularly at risk. With an increase of a few degrees Celsius in the annual mean temperature at BL, there is a possibility that the alpine tundra could disappear in Gros Morne National Park with important impacts on wildlife such as caribou.

Concerning alpine snow and ice cover, the global 20$^{th}$ century warming of approximately 0.6°C has caused a widespread retreat of glaciers in most regions. However, in a few maritime regions, the increase in temperature has been overcompensated by larger snowfall, thus causing glaciers to advance (IPCC, 2001a).

The winter season climate in Newfoundland is affected by the North Atlantic Oscillation (NAO) as studied by Banfield and Jacobs (1998a), which influences the strength of the westerlies and the synoptic weather types experienced on the island. From the early 1970s through the early 1990s, the winter NAO was mainly in a positive phase (IPCC, 2001a), which caused stronger mid tropospheric westerly winds and lower temperatures over the region including Gros Morne National Park in winter, due to the arctic front being more often south of Newfoundland. However, the winter cooling experienced over the northwest North Atlantic and Newfoundland has reversed in the late 1990s (Banfield and Jacobs, 1998a; IPCC, 2001a). Nevertheless, climate change could involve local and regional changes in climate that counter the hemispheric trends due to atmospheric circulation shifts, such as caused by the NAO.

In studying climate change, analog scenarios can be of assistance (Barry and Chorley, 1998). Unusual seasons in the past can perhaps offer a glimpse as to what may be expected more frequently in the future. Many GCMs currently project warming and more precipitation in Newfoundland due to anthropogenic climate change, although the strength of the changes is expected to be smaller than most areas, particularly compared with continental interiors. Generally speaking, the IPCC reports an expected decreased snowcover amount and duration for the region including Newfoundland, although such projections would not necessarily take into account the situation at higher altitudes, especially for the Great Northern Peninsula which is near the edge of the Eastern Canadian (and Northeast USA) region (IPCC, 2001b).

## 6.2 Methodology

In order to estimate snow cover conditions at the BL site, several mathematical relationships were needed to convert temperature, precipitation and wind speed data into a reasonable estimate of SWE. The snow model used will be discussed in terms of accumulation, wind transport (drifting) and sublimation, and melt.

Snow accumulation was calculated in a straightforward manner with the snowfall falling on any given day, in mm of SWE, appearing as a snow on the ground at the beginning of the next day. The snow on the ground was presented as a SWE depth in mm. However, not all the snow falling on any given day was partially or totally reported as SWE on the next day, as it is subjected to wind transport, sublimation and melt on the day of precipitation. Rainfall could also add to SWE if its internal energy was insufficient to raise the snowpack temperature to an isothermal 0°C state and saturate it with liquid water. These factors were dependent on the snowpack temperature (cold content) as well as its SWE. Appendix 1 discusses further the equations used in the snow model.

Snowdrifting was calculated using the reconstructed ten metre hourly wind speed record on BL, following the expression suggested by Tabler *et al.* (1990) presented as

Equation 6.4 in section 6.1. According to Pomeroy and Gray (1995), Equation 6.4 is valid for fetch distances greater than one kilometre. With an area of approximately 12 $km^2$, the BL summit plateau was assumed to be a perfect circle with a diameter of 4 km for the purposes of snow transport (drifting) and sublimation. Given that only the average SWE on BL was sought, and the coarseness of the blowing snow equation, the increase in error due to this assumption was considered negligible. In addition, snow blown off the plateau was assumed to be a loss to the SWE and unable to blow back on to the plateau in the event of a reversal of wind direction. This assumption was consistent with some observations for windy plateaus, for example by Pomeroy (1991). The two major assumptions allowed wind direction to be ignored as a model parameter. The permanence of blown snow loss was consistent with the terms of use of equation 6.4, and approximately correct given the steep slopes of the edge of the plateau. Hourly transported snow values were summed for the entire day to give daily values. Daily transport losses from the plateau were converted to a SWE using the fetch distance of 4 km and the density of water.

The blowing snow loss from the plateau was limited by two factors. The first is the minimum threshold wind speed required for transport. A value of 6.5 m/s was used, irrespective of snow conditions. Although the threshold wind speed could introduce a considerable error in low wind speed environments, it becomes increasingly less important in windier locations, due to the fact that blowing snow totals near the threshold value were very small compared to those that occur at high wind speeds. The second factor was the snow available for transport (A). Blowing snow totals could not surpass what was judged to be available for transport, which was the sum of 1) the snow available from the previous day, 2) snowfall, 3) snowdrifting and 4) blowing snow sublimation (see Appendix 1). However, if the average daily temperature ($T_{mean}$) rose above -2°C or rainfall occurred, than the resulting surface melt and subsequent refreeze was assumed to lock away the snowpack from future relocation. Although transportation of the snowpack with a surface ice layer was theoretically possible at very high wind speeds, a more detailed algorithm would have been too complex and furthermore the error caused by this assumption would have been small. Consequently, the frequent snowfalls during the snowcover season at BL reduce the susceptibility to underestimation of snow transport due to this assumption.

Sublimation of snow was only considered significant for blowing particles. The relationship used to calculate blowing snow sublimation was the one offered by Tabler *et al.* (1990), as adapted from Pomeroy (1988), presented in section 6.1 as Equation 6.5. The equation is a function of snow transport (q), air temperature ($T_{mean}$) and relative humidity (rh). Tabler *et al.* (1990) gave various values of their coefficients m and b for different air temperatures and relative humidity as calibrated for the Canadian Prairies. However, other regions should respond similarly when under those selected atmospheric conditions. In order to obtain a methodology which was not unreasonably complex, only one set of m and b values were used for all meteorological conditions experienced on the BL plateau. The m and b values were estimated using linear interpolation and extrapolation from the values published by Tabler *et al.* (1990). Average conditions assumed were -8°C, which is the BL mean snow accumulation season temperature from 1962 to 1999, and 95% rh. Although rh values at BL are typically closer to 90% during non-blowing snow conditions in winter, the evaporation from blowing snow would make the rh even closer to saturation.

The most useful published m and b values were those for -15°C / 70% rh, -1°C / 70% rh and -15°C/ 90% rh conditions. Accordingly, m and b values for BL were estimated for -8°C/ 70% rh and then decreased for the higher relative humidity expected on BL. This was accomplished using the vapour pressure deficit change from the conditions at -15°C / 70% rh to 90% rh at the same temperature and then extrapolated linearly to 95% rh (at -8°C). Using vapour deficits instead of just rh values accounts for the fact that the same rh values have different drying powers at different temperatures. As a result, the estimated m and b values were 2.0 and 0.9 respectively, which were close to the published values for the -15°C and 70% rh conditions.

However, it was shown by Dery and Yau (2001) and Xiao *et al.* (2000) that this method overestimates sublimation, because large blowing snow events would be self-limiting in terms of sublimation. The 1998 and 1999 BL test years verified this as the sublimation values were too large on a seasonal basis, especially due to storm days with high wind speeds. Consequently, an upper maximum to daily sublimation values was set. The maximum value applied was 15 mm WE per day, which is completely empirical in nature, being set at the value that minimized the error between modelled and measured mid-season BL SWE values for the 1998 and 1999 snow seasons. Blowing snow sublimation was calibrated in this fashion because it was judged to be the term with the largest error in the snow model because of the size of the sublimation values and the significance of the assumptions made.

The final major component of the snow model was the melt algorithm. It was based on the degree-day approach (Equation 6.7). The degree-day factors used for BL increased as the snow season progressed. A value of 1.5 mm $^{\circ}C^{-1}$ $d^{-1}$ was applied from the beginning of the snow season until the end of January, with the melt factors increasing by 0.5 mm $^{\circ}C^{-1}$ $d^{-1}$ per month until July, when the value assumed is 4.5 mm $^{\circ}C^{-1}$ $d^{-1}$. These values were lower than the values suggested by the WMO (1964) for open environments because of the lower base temperature, and also the high cloudiness on the BL plateau and the measured coastal plain values for winter up to the month of April. The base temperature used was -2°C because of the exposed nature of the BL site. For days with rain, a different expression was used to account for the energy supplied by the rain water. Similarly to Brown and Braaten (1998), the expression adapted from the U.S. Army Corps of Engineers (1956) was used here:

Equation 6.8

$$M = c*[0.074 + 0.007(R*0.03937)((9/5*T_a + 32) - 32) + 0.05]/0.03937$$

where:

M and $T_a$ = same as above
R = rainfall (mm $d^{-1}$)
c = local calibration factor (dimensionless)

The local calibration factor was set to 0.5, based on the period between JD57 and JD135 during the 1999 snow season, during which almost all the melt occurred on days with rain.

However, there were indications that melt on days with rain varies widely. In 1995, a rather large melt event due to rain indicates a calibration factor of an order of magnitude larger than that used, but such variability in the amount of melt cannot be accounted for accurately with estimates of air temperature and rainfall only.

The degree-day method works best when a snowcover is ripe for melting, i.e. isothermal at 0°C and saturated with water. For thicker snowcovers, however, the degree-day method tends to overestimate melt due lack of consideration of the cold content and meltwater requirements necessary before ablation may begin. The simplest model to determine the change in internal energy ($\Delta U$) of the snowpack is expressed as follows:

$$\Delta U = \mu C_p(T_f - T_i) \qquad \text{Equation 6.9}$$

where:

$\Delta U$ = change in internal energy of the snowpack per square metre (kJ $m^{-2}$)
$\mu$ = mass of snowpack (kg $m^{-2}$)
$C_p$ = specific heat of snow, 2.09 kJ $kg^{-1}$ $°C^{-1}$
$T_f$, $T_i$ = final and initial temperature of the snowpack, respectively, with a daily time step (°C)

When the temperature of the snowpack fell under 0°C, a cold content developed, which needed to subsequently be removed before melt could begin. The snowpack temperature as determined essentially by air temperature determined the cold content. A negative degree day factor of -2 mm °C $d^{-1}$ within the temperature index equation was used for temperatures below the degree day base temperature multiplied by a refreeze factor of 0.5 to determine the heat loss from the snowpack as discussed in Bloschl and Kirnbauer (1991). The refreeze factor was used in combination with a maximum cold content value set by the air temperature to prevent the calculation of unrealistic cold content values. The latter adjustments were undertaken to prevent the snowpack temperature from falling below or rising above the air temperature during cooling and warming episodes respectively, which often occurred when the snowcover was shallow. For shallow snowcovers, the energy required to alter the temperature of the entire snowpack is rather small, and the degree-day method overestimated the change in internal energy. For example, a snowpack of 10 kg $m^{-2}$ at -4°C subjected to a day of mean air temperature -10°C would lose 2001 kJ according to the negative degree-day method. However, because of the thin snow cover, such a loss would reduce the temperature of the snow by 95°C, which is clearly impossible. The snow temperature is thus set to -10°C, which would mean a decrease in internal energy, or an increase in the cold content, by 125 kJ, rather than by 2001 kJ suggested by the negative degree-day equation. However, for thick snowcovers, the snow temperature would vary less than the air temperature, introducing a lag between relatively higher temperatures and snow ablation in the early part of the snowmelt season.

Another algorithm inserted into the model was the liquid water content, which required the snowpack to reach a saturation point of 5% of SWE before the snowpack could start ablating. Meltwater as well as rainwater can contribute to the liquid water content. For rainwater to contribute to liquid water content however, the cold content of the snowpack first needed be reduced to zero, otherwise the rain simply reduces the cold content of the snowpack by releasing its latent energy of fusion by freezing and contributing to the SWE.

When liquid water present in the snowpack was subjected to sub-freezing temperatures, the water froze inside the snowpack before the cold content value could start increasing.

The snow model was run for every snow year from 1962 to 1999. A daily record was built, and monthly and seasonal statistics were compiled. A key part of the results deals with the notion of continuous snowcover. Following the discussion in section 6.1.1, the continuous snowcover season is a spatial and temporal concept where a snow covers blankets the entire surface and is present for long time periods. The start of the continuous season is the date when SOG is present for at least seven consecutive days and does not disappear for seven days or longer until the end of the season. The point in time where the snowpack SWE drops to zero mm for seven days or longer is considered to be the disappearance date or end of the continuous season. Periods of less than seven days with no SOG within the continuous season are ignored in this definition. Special seasons with important periods of SOG outside the continuous season are given attention in the results and discussion section (6.3).

## 6.3 Results and Discussion

### 6.3.1 Comparison of the Modelled SWE Results with the Measured Values for Big Level

Very few measurements of snow depth or SWE have been taken on the Big Level plateau (BL). This is due to the difficulty of accessing the site, and to the harsh weather conditions experienced on the plateau. In 1998 two snow surveys were taken at the climate station on the plateau, one on JD 54 and one on JD 121 in conjunction with ongoing research related to this study. In 1999 an additional two measures were taken, through snow pits, on JD 57 and JD 135. In 1997 a series of eight surveys were taken at Puncheon Rock east of the plateau, and in 1995 four surveys were also undertaken at Puncheon Rock, all taken by National Park wardens. This was a small sample to work with. In particular, the fact that the Puncheon Rock site is lower in elevation and is a more sheltered site subjected to less net drifting loss and blowing snow sublimation was an important consideration.

Figures 6.1-6.4 compare the modelled values with the real measurements for BL. The 1998 and 1999 estimations are fairly close to the measured values. However, these two seasons were used for the calibration of the model and therefore do not constitute independent test cases. Despite the calibration efforts, there is still a considerable discrepancy of approximately 170 mm between the JD 54 1998 measured and modelled values. The other three measured values agree fairly closely to the modelled ones however.

The 1997 modelled SWE values performed poorly against the measured ones.

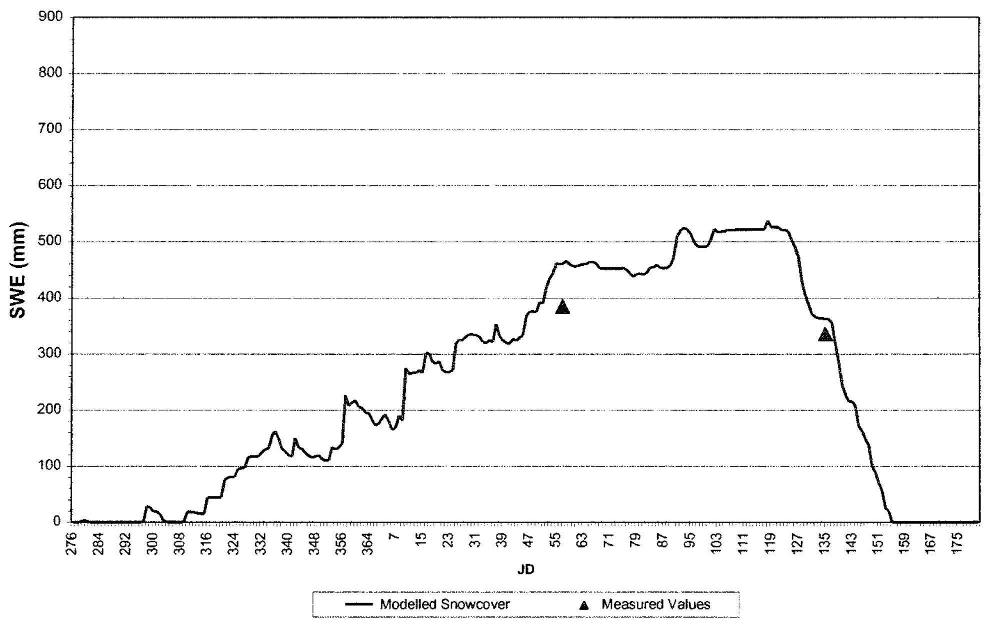


Figure 6.1 **Big Level modelled snowcover for the 1999 snow year.** The two measured values were already in SWE units.

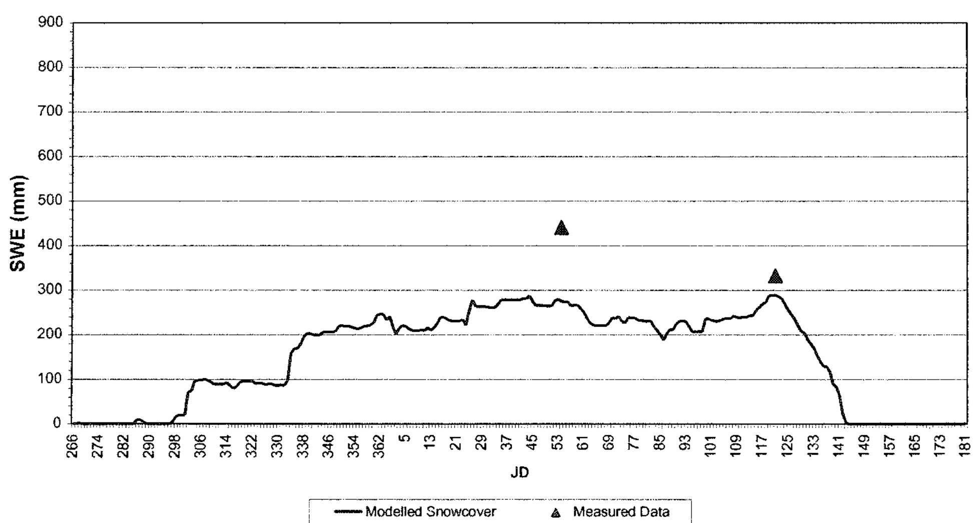


Figure 6.2 **Big Level modelled snowcover for the 1998 snow year.** The measured values were taken in depth units and were converted using a mean estimated density of 0.3 in Feb and 0.35 in early May.

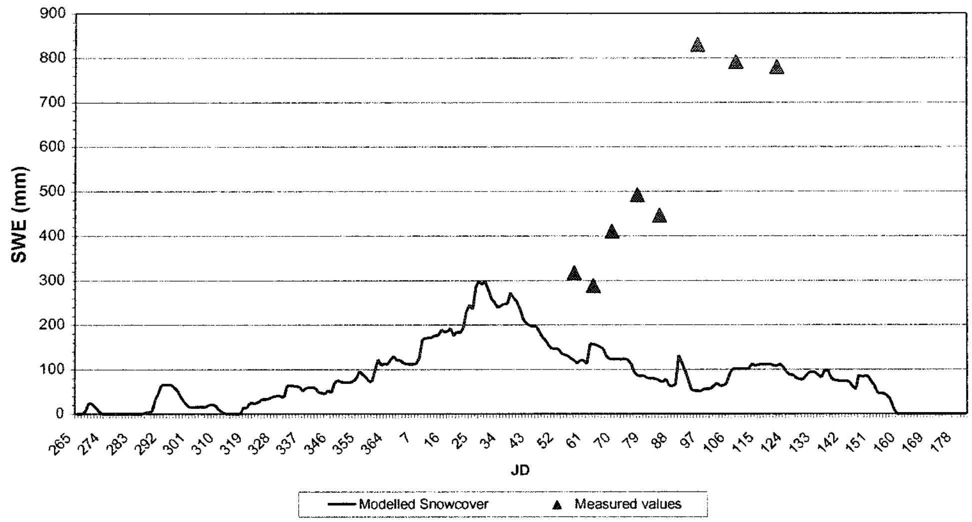


Figure 6.3 **Big Level modelled snowcover for the 1997 snow year.** The measured values were taken at Puncheon Rock, a lower elevation and less exposed site east of the summit plateau. The 1997 year was windy in comparison to most other years.

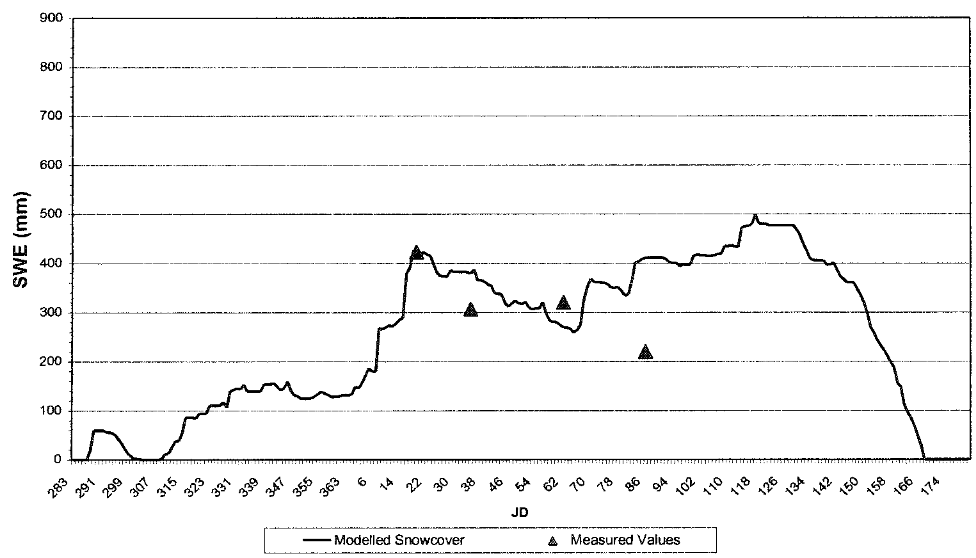


Figure 6.4 **Big Level modelled snowcover for the 1995 snow year.** The measurements were taken on Puncheon Rock, east of the summit plateau. The 1995 snow year was less windy than average.

However, given that there is a site difference between both locations, it is possible that Puncheon Rock could have experienced more SWE than the BL summit plateau. This could be explained by the former site being downslope and downwind from the latter. Furthermore, the 1997 season was particularly windy, and the model estimated large snowdrift and blowing snow sublimation values. It is likely however, that the drifting and sublimation values estimated for BL are too large due to the size of the discrepancy between modelled and measured values.

Finally, the 1995 measured values agreed reasonably well with the modelled ones. That season was below average in terms of windiness which likely explains why the Puncheon Rock measured values are closer to the BL modelled values. Nevertheless, the modelled SWE values are crude. A conservative estimate of the probable error of the estimated late season SWE would be ±100 mm. Thus, modelled SWE values should only be relied upon in a relative way as year to year indicators. It is essential to note that the discrepancy between the measured and modelled values could also be due to errors in the temperature, precipitation and wind speed estimates. Without further information, it is impossible to know and caution must be used in interpreting all the results of this thesis.

### <u>6.3.2 The Modelling Results</u>

The modelling results were compiled as Microsoft Excel spreadsheets with one file for each snow year on a daily time scale. These were named *BL 19XX.xls*; one sample is provided on the CD (Appendix 2). The results were then rounded up into one file, namely *BL daily stats.xls* where only the principal findings of snowdrifting, blowing snow sublimation, rain accumulation in the snowpack and SWE were kept. In addition to these snowcover variables, the daily values of temperature and precipitation are also present. These results were then compiled into monthly statistics on one spreadsheet having the file name *BL monthly statistics.xls*. These three files are included on the CD as Appendix 2. The following is a discussion of these results.

### 6.3.3 The Big Level Snowcover

Table 6.3 shows some important estimated snow related variables by year for the 1963-1999 snow years. At the bottom of the table are the means, standard deviations, maximum and minimum values for each variable. Each row (year) tells the story of the snowcover for a particular year.

The first frost of the year has important climatological and biological significance. During the study period (October 1962- July 1999), the estimated average date of the first frost on BL has been JD 266 (23 September). However, the first frost has varied in time from late August to early October. The last frost is estimated to be typically in mid-June with an estimated average of JD 164 (13 June). Similarly to the first frost, the last frost can vary in time across a range of over a month. During the study period the last frost has probably occurred from the end of May to the end of June on BL (see Chapter 3 for a detailed

Table 6.3 **Summary of key estimated Big Level climate parameters for the study period (October 1962 - July 1999)**.
Snow on Ground (SOG) is recorded only if the snow is estimated to remain until the following day. The (temporally) continuous SOG (or snowcover) seasor can have some periods with no SOG. The snow accumulation season extends from November to April. See text for a definition of the continuous snowcove season.

| Snow Year | First frost (JD) | First snowfall (JD) | First SOG (JD) | Start of continuous SOG (JD) | End of continuous SOG (JD) | Last SOG (JD) | Last snowfall (JD) | Last frost (JD) | Duration of continuous SOG season (days) | Periods with no SOG within the continuous SOG season (JD) |
|---|---|---|---|---|---|---|---|---|---|---|
| 1963 | n/a | n/a | 287 | 347 | 140 | 140 | 129 | 165 | 159 | |
| 1964 | 257 | 274 | 288 | 320 | 200 | 200 | 142 | 182 | 246 | 335-338 |
| 1965 | 275 | 275 | 276 | 287 | 180 | 180 | 145 | 148 | 260 | 294-298 |
| 1966 | 262 | 272 | 273 | 300 | 186 | 186 | 134 | 162 | 252 | 324 |
| 1967 | 273 | 276 | 292 | 320 | 185 | 185 | 151 | 165 | 231 | 329, 331-336, 346 |
| 1968 | 277 | 281 | 313 | 313 | 190 | 190 | 161 | 162 | 243 | |
| 1969 | 283 | 293 | 294 | 302 | 174 | 174 | 148 | 153 | 239 | |
| 1970 | 274 | 275 | 277 | 327 | 163 | 163 | 146 | 157 | 202 | 340 |
| 1971 | 264 | 292 | 293 | 299 | 167 | 167 | 146 | 180 | 234 | |
| 1972 | 271 | 271 | 272 | 292 | 176 | 176 | 145 | 164 | 250 | |
| 1973 | 271 | 287 | 288 | 288 | 148 | 148 | 127 | 170 | 227 | |
| 1974 | 267 | 286 | 287 | 295 | 167 | 167 | 157 | 162 | 238 | 306-307 |
| 1975 | 267 | 274 | 285 | 285 | 150 | 150 | 133 | 162 | 231 | 81-82 |
| 1976 | 273 | 277 | 278 | 288 | 92 | 154 | 154 | 165 | 170 | 299-303, 313-315 |
| 1977 | 275 | 285 | 286 | 290 | 156 | 156 | 147 | 174 | 233 | |
| 1978 | 257 | 256 | 264 | 324 | 169 | 169 | 153 | 168 | 211 | |
| 1979 | 256 | 256 | 274 | 293 | 162 | 162 | 132 | 176 | 235 | |
| 1980 | 273 | 270 | 274 | 289 | 171 | 171 | 155 | 168 | 248 | 298-300, 308-309 |
| 1981 | 231 | 264 | 266 | 289 | 97 | 141 | 140 | 164 | 175 | 44 |
| 1982 | 237 | 283 | 283 | 314 | 163 | 163 | 143 | 164 | 215 | |
| 1983 | 251 | 280 | 281 | 312 | 119 | 140 | 138 | 151 | 173 | |
| 1984 | 278 | 293 | 294 | 304 | 109 | 161 | 158 | 175 | 171 | 312 |
| 1985 | 272 | 271 | 280 | 325 | 153 | 153 | 163 | 162 | 195 | 349 |
| 1986 | 266 | 287 | 288 | 292 | 127 | 142 | 149 | 157 | 201 | |
| 1987 | 261 | 266 | 272 | 290 | 146 | 146 | 137 | 151 | 222 | |
| 1988 | 284 | 270 | 285 | 304 | 148 | 161 | 159 | 160 | 210 | |
| 1989 | 273 | 273 | 275 | 288 | 161 | 161 | 124 | 164 | 240 | 300-303 |
| 1990 | 271 | 271 | 272 | 290 | 162 | 162 | 151 | 163 | 238 | |
| 1991 | 274 | 284 | 294 | 305 | 186 | 186 | 167 | 170 | 247 | |
| 1992 | 236 | 244 | 275 | 312 | 158 | 158 | 149 | 162 | 212 | |
| 1993 | 248 | 276 | 277 | 277 | 156 | 156 | 154 | 165 | 246 | 286-289 |
| 1994 | 274 | 287 | 293 | 293 | 164 | 164 | 159 | 161 | 237 | |
| 1995 | 262 | 284 | 288 | 288 | 167 | 167 | 181 | 181 | 245 | 304-308 |
| 1996 | 272 | 281 | 282 | 324 | 155 | 174 | 173 | 173 | 197 | |
| 1997 | 266 | 266 | 268 | 287 | 158 | 158 | 154 | 157 | 238 | 314-317 |
| 1998 | 268 | 267 | 268 | 297 | 143 | 143 | 119 | 157 | 212 | |
| 1999 | 276 | 276 | 279 | 297 | 155 | 155 | 135 | 160 | 224 | 304-309 |
| | | | | | | | | | | |
| **average** | **266** | **276** | **282** | **301** | **157** | **163** | **148** | **164** | **222** | |
| **standard dev** | **13** | **11** | **10** | **16** | **24** | **15** | **13** | **8** | **27** | |
| **maximum** | **284** | **293** | **313** | **347** | **200** | **200** | **181** | **182** | **260** | |
| **minimum** | **231** | **244** | **264** | **277** | **92** | **140** | **119** | **148** | **159** | |

Table 6.3 (cont.) See chapter 5 for snowfall, chapter 3 for temperature and chapter 4 for wind speed.

| Snow Year | Total snowfall (mm WE) | Maximum snowpack SWE (mm) | Month of maximum SWE (1-12) | Snow accumulation season mean temperature (°C) | Snow accumulation season rainfall (mm) | Snow accumulation season mean wind speed (km/hr) | Snowdrift loss (mm) | Blowing snow sublimation (mm) |
|---|---|---|---|---|---|---|---|---|
| 1963 | 1180 | 260 | 3 | -7.4 | 340 | 38 | 190 | 770 |
| 1964 | 2820 | 1650 | 5 | -8.1 | 70 | 37 | 220 | 930 |
| 1965 | 2120 | 1060 | 5 | -7.8 | 50 | 35 | 170 | 840 |
| 1966 | 2180 | 1140 | 5 | -7.0 | 40 | 34 | 180 | 860 |
| 1967 | 2240 | 1320 | 4 | -8.4 | 100 | 34 | 140 | 860 |
| 1968 | 2250 | 1200 | 4 | -6.8 | 110 | 36 | 170 | 920 |
| 1969 | 1850 | 720 | 5 | -6.2 | 30 | 34 | 190 | 920 |
| 1970 | 1180 | 290 | 5 | -6.0 | 180 | 33 | 90 | 640 |
| 1971 | 1810 | 820 | 4 | -7.2 | 180 | 34 | 130 | 850 |
| 1972 | 1840 | 710 | 5 | -9.9 | 50 | 34 | 170 | 930 |
| 1973 | 1170 | 260 | 3 | -10.0 | 40 | 34 | 160 | 790 |
| 1974 | 1200 | 380 | 5 | -9.7 | 150 | 35 | 130 | 670 |
| 1975 | 900 | 130 | 4 | -9.2 | 140 | 32 | 90 | 620 |
| 1976 | 1060 | 200 | 1 | -8.7 | 230 | 33 | 130 | 680 |
| 1977 | 1540 | 330 | 5 | -8.8 | 120 | 34 | 190 | 960 |
| 1978 | 1500 | 580 | 4 | -7.7 | 290 | 32 | 110 | 670 |
| 1979 | 1530 | 720 | 4 | -7.4 | 190 | 32 | 100 | 640 |
| 1980 | 1360 | 530 | 2 | -7.4 | 350 | 31 | 120 | 680 |
| 1981 | 1010 | 130 | 12 | -6.2 | 360 | 33 | 130 | 690 |
| 1982 | 1430 | 460 | 3 | -8.4 | 360 | 33 | 140 | 810 |
| 1983 | 710 | 130 | 3 | -6.5 | 410 | 31 | 60 | 430 |
| 1984 | 1230 | 110 | 2 | -8.0 | 130 | 38 | 190 | 1000 |
| 1985 | 1010 | 130 | 5 | -9.7 | 200 | 35 | 120 | 720 |
| 1986 | 1000 | 160 | 2 | -8.7 | 260 | 34 | 110 | 710 |
| 1987 | 960 | 220 | 2 | -8.5 | 80 | 32 | 90 | 600 |
| 1988 | 1140 | 270 | 3 | -8.0 | 90 | 34 | 110 | 750 |
| 1989 | 1170 | 470 | 4 | -9.7 | 190 | 30 | 70 | 550 |
| 1990 | 1100 | 420 | 5 | -10.8 | 200 | 30 | 70 | 550 |
| 1991 | 1360 | 670 | 5 | -9.9 | 220 | 30 | 80 | 600 |
| 1992 | 1010 | 280 | 4 | -10.0 | 120 | 31 | 90 | 670 |
| 1993 | 1050 | 250 | 3 | -10.1 | 80 | 30 | 80 | 610 |
| 1994 | 1230 | 270 | 5 | -9.3 | 110 | 35 | 200 | 750 |
| 1995 | 1230 | 500 | 4 | -8.5 | 270 | 31 | 100 | 610 |
| 1996 | 950 | 190 | 4 | -7.2 | 260 | 30 | 90 | 580 |
| 1997 | 1410 | 300 | 1 | -8.2 | 100 | 34 | 230 | 920 |
| 1998 | 930 | 290 | 2 | -6.7 | 190 | 30 | 90 | 550 |
| 1999 | 1270 | 540 | 4 | -6.3 | 140 | 32 | 100 | 650 |
| | | | | | | | | |
| **average** | **1380** | **490** | **4** | **-8.2** | **170** | **33** | **130** | **730** |
| **standard dev** | **470** | **380** | **2** | **1.3** | **100** | **2** | **50** | **140** |
| **maximum** | **2820** | **1650** | **5** | **-6.0** | **410** | **38** | **230** | **1000** |
| **minimum** | **710** | **110** | **12** | **-10.8** | **30** | **30** | **60** | **430** |

discussion of BL temperatures).

The first snowfall is often not far behind the first frost. On BL, the estimated average date of first snowfall during the study period was JD 276 (3 October), but has likely varied over the period from early September (JD 244) to mid-October (JD 293). In addition, the last snowfall is estimated to have occurred on JD 148 (28 May) on average for BL. The latest estimated last snowfall was on JD 181 (30 June), while the earliest was on JD 119 (29 April). The last snowfall typically precedes the last frost by a couple of weeks on average (see Chapter 5 for a detailed discussion of BL precipitation).

The first occurrence of snow on the ground (SOG), or snowcover, is usually an ephemeral one. However, an estimated eight years during the study period saw the first SOG occurrence on BL last for a period of seven days or longer, including five where the continuous snowcover season started simultaneously with the first SOG occurrence (1968, 1973, 1975, 1994 and 1995). On BL during the study period, the first SOG was estimated to occur on JD 282 (9 October) on average. The first SOG typically trails the first snowfall by approximately a week due to the temperature near the ground being too high for the snow to remain until the following day. Figure 6.5 shows the first seasonal snow on ground event and the start date of the continuous snowcover season for BL during the study period. The earliest estimated appearance of snow cover was in 1977 on JD 264 (21 September in 1978 snow year), while the latest was in 1967 on JD 313 (9 November in 1968 snow year). With an estimated standard deviation of 10 days, most first occurrences of SOG would have happened in early to mid October on BL during the study period. In fact only six years are

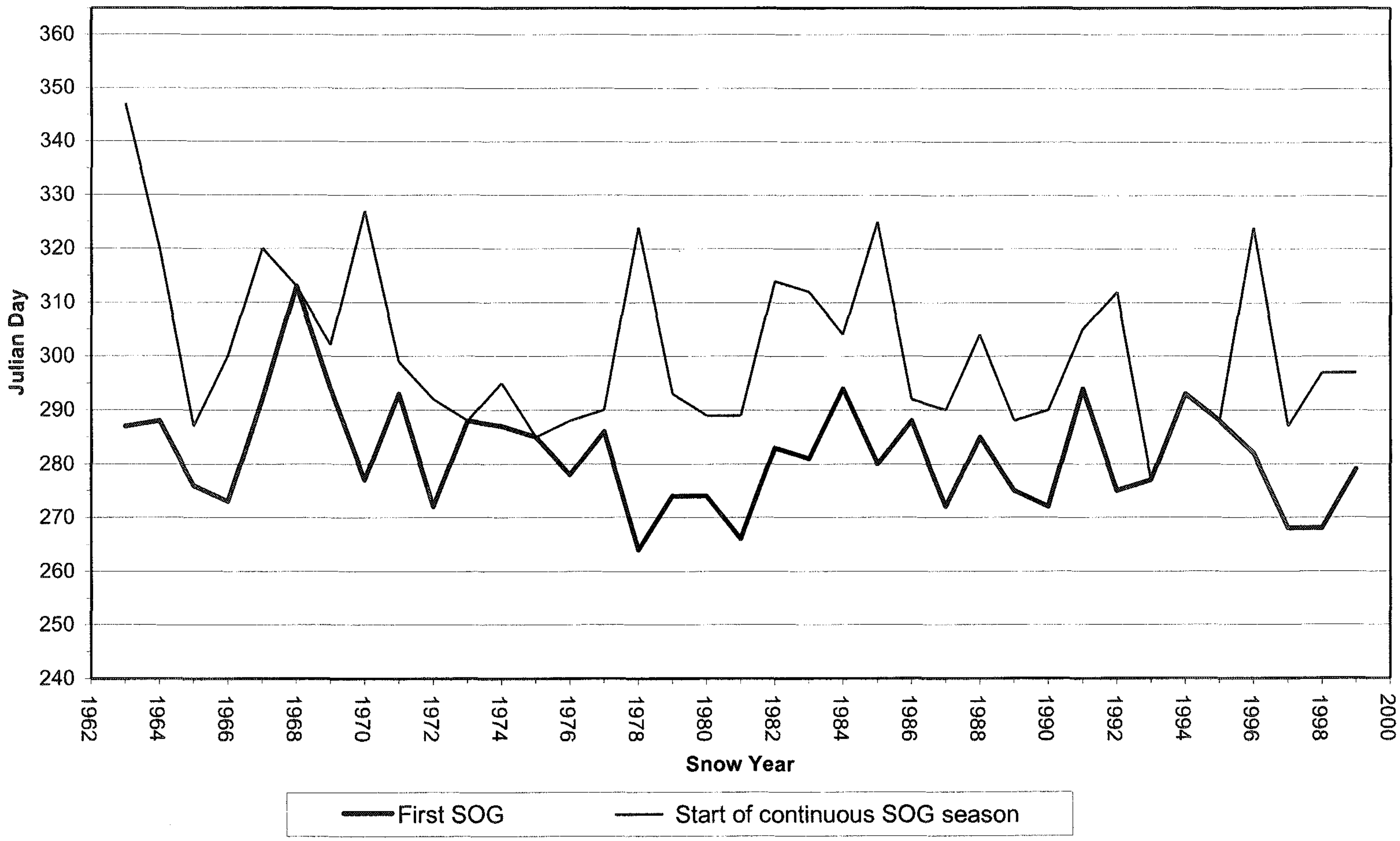


Figure 6.5 **First BL seasonal snow on ground (SOG) event and start of the continuous snowcover (or SOG) season.** A snow year starts in August of the previous year.

estimated to have had SOG occurrences during September. There appear to be no temporal trends in the study period concerning this variable on BL.

The last occurrence of snow on the ground (SOG) normally coincides with the end of the continuous snow cover season, meaning that the snow had been in place for at least a couple of months (see Figure 6.6). In this analysis we ignore late lying snow beds off the edges of the plateau that are the result of snowdrifting. Instead the last occurrence of SOG is the point in time when most of the plateau becomes snow free. It is estimated that seven snow years did not have their last occurrence of SOG simultaneous with the end of the continuous snow cover season, concentrated mainly in the 1980s: 1976, 1981, 1983, 1984, 1986, 1988 and 1996. Almost all these years had estimated last snowfalls within one standard deviation of the mean date and continuous snowcover seasons ending before the mean date, which means that the detached SOG occurrences were not the result of particularly late heavy snowfalls except for 1996, which saw its last snowfall on JD 174 (23 June). In fact, the last occurrence of SOG was estimated to occur, on average, 15 days after the last snowfall. The mean last occurrence of SOG was estimated to be JD 163 (12 June) with most of the SOG occurrences ending in June (standard deviation of 15 days). Six snow years had estimated last SOG occurrences in late May, with the earliest estimated to be in 1963 on JD 140 (20 May), the others being 1973, 1975, 1981, 1983 and 1998. With respect to years with late ending of the snowcover year, five snow years are estimated to have had their last SOG occurrence in July, with all but one of them (1991) occurring in the 1960s (1964, 1966, 1967, 1968). The latest SOG disappearance is likely to have occurred a year

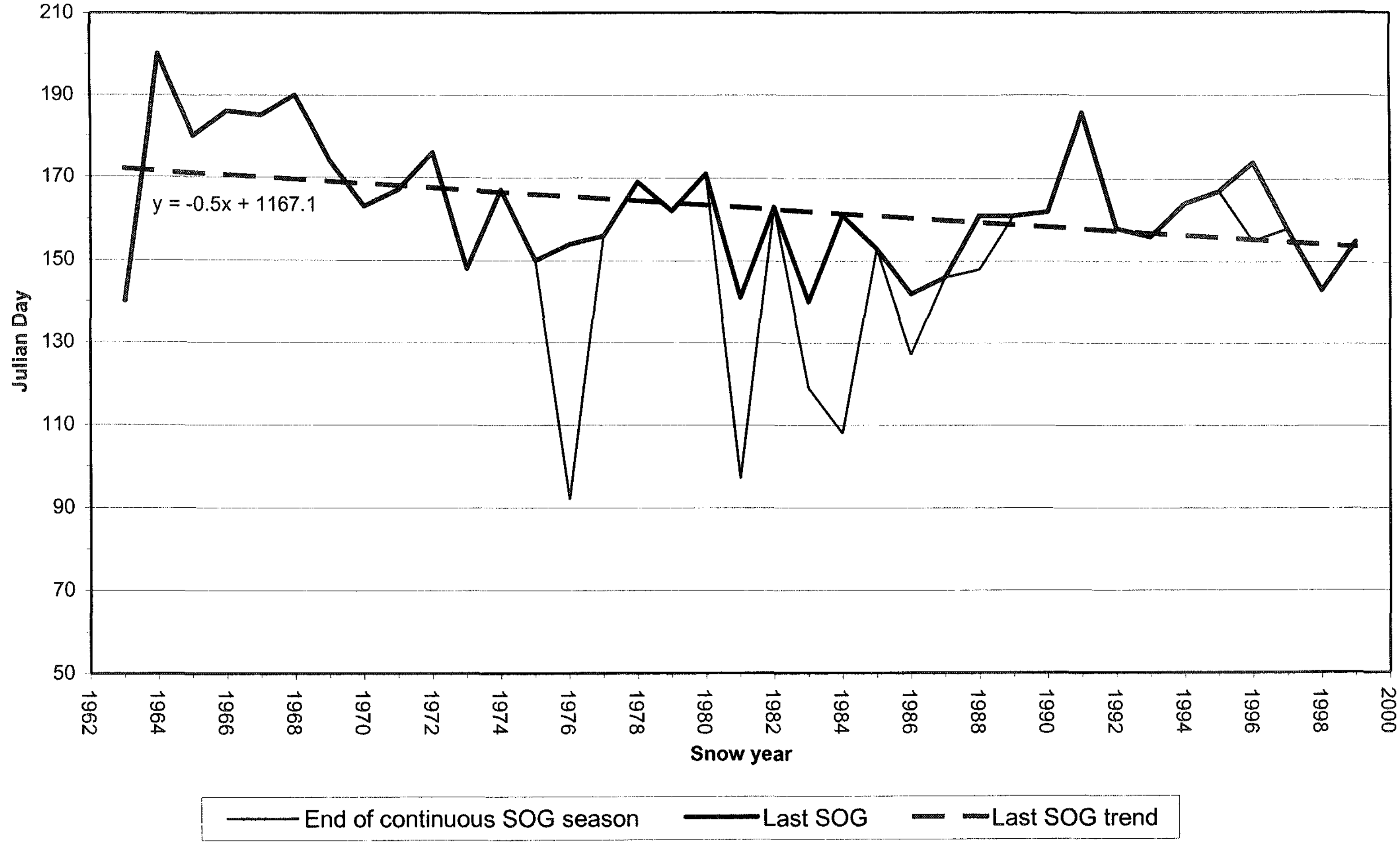


Figure 6.6 **BL last seasonal SOG occurrence and end of the continuous snowcover (or SOG) season**. On most years, the two dates are the same. The significant linear trendline equation for the last SOG occurrence is indicated on the graph.

after the earliest disappearance, i.e. 1964. On BL during this extended snow season, it is estimated that the last SOG occurrence would have been on JD 200 (18 July). During the study period (1963-1999), there is a significant negative linear trend of last SOG occurrence, i.e. becoming earlier in the year by the rate of -0.5 days per year. This translates into an 18.5 day difference between the beginning and the end of the study period.

The continuous snowcover season distinguishes itself from the presence of SOG by the length of time the snow remains. The estimated date of appearance of the continuous snowcover trails the first occurrence of SOG by approximately 20 days on average on BL for the study period, with the mean being JD 301 (28 October). In comparison, the continuous snowcover season starts approximately 50 days later on average on the coastal plain, as measured with the Daniel's Harbour (DH) record.

The standard deviation for the appearance of the continuous snowcover season on BL is large at an estimated 16 days, meaning that the date at which the snowcover establishes itself for the winter is highly variable during the study period. The earliest appearance of the continuous snowcover is estimated to be in 1992 (1993 snow year) on JD 277 (3 October). Another seven snow years also had their continuous snow cover appear in the first half of October: 1965, 1973, 1976, 1980, 1989, 1995 and 1997. As for late continuous snow cover appearances only one started in December, which is the 1963 snow season that started on JD 347 (13 December). Only five other seasons began after mid-November (1964, 1967, 1978, 1985, and 1996). There are no significant trends to be mentioned concerning the appearance of the continuous snow cover season.

The initial data set revealed some suspiciously late (estimated) starts to the continuous snowcover seasons. In fact, during some years the continuous snowcover season start was estimated to be later on BL than at DH, which is highly improbable. It was discovered that the snow drifting and blowing snow algorithms in the model were scouring the snowpack to a SWE of 0 mm. This was a problem early in the snow season, before melt or rain on snow events could lock in the SWE and make it unavailable for transport. The data were visually inspected and all the 0 values caused by wind scouring were changed to 2 mm SWE, to allow for traces of SOG to remain present as it would not be possible for the snowcover to be reduced to 0 mm given the presence of low lying vegetation. This small adjustment changed the dates of appearance of the continuous snowcover season for nine snow years making it considerably earlier in some cases (119 days earlier in 1974) and removed most of the initially reported periods of no SOG within the continuous season up to mid-winter. It is acknowledged that due to this problem, the snowcover SWE on BL is underestimated during many seasons.

The end of the continuous snowcover season closely mirrors the last day of SOG presence. The estimated average date of the end of continuous snow cover season for BL during the study period is JD 157 (6 June), six days before the estimated mean last SOG occurrence. When the continuous snowcover season ends after this mean date, it always corresponds exactly to the estimated last day of SOG presence. A large factor explaining the end date of the continuous snowcover season is the maximum snowpack SWE which explains 56% of the variance of the continuous snowcover season end date. The end of the continuous snowcover season at DH is approximately 55 days earlier on average. The standard deviation of this variable is very large on BL with a value of 24 days. Only four years saw their estimated continuous snowcover season end before May (1976, 1981, 1982, 1984), with the earliest being on JD 92 1976 (1 April). However, all continuous snow seasons estimated to end before JD 140 (20 May) during the study period on BL had some SOG reappearing until this threshold date (JD 140). Surprisingly, there is no significant negative trend for the timing of the end of the continuous snowcover season, despite the existence of such a trend for the date of last SOG occurrence. This is likely mainly the result of the 1976 snow season when the continuous snow season ended 62 days before the last SOG occurrence.

The continuous snowcover season for BL is estimated to last for 222 days on average for the study period (almost 7½ months), with a standard deviation of 27 days. This is approximately 3.5 months longer than on the coastal plain on average. Four snow seasons had less than a two month difference between the two locations, while six snow seasons had a difference of over four months. The time series for this variable is shown in Figure 6.7.

In accordance with the methodology, there can be periods of up to six consecutive days within the "continuous" snowcover season with no SOG. Of the 17 snow years that have some estimated days with no SOG within the continuous snowcover season, all of them have their snow free spells before 16 December (JD 350) except for one estimated snow free day in 1981 on JD 44 (13 February). This is the only day in the BL record estimated not to have SOG from JD 350 (16 December) to JD 91 (1 April). This is not surprising given that

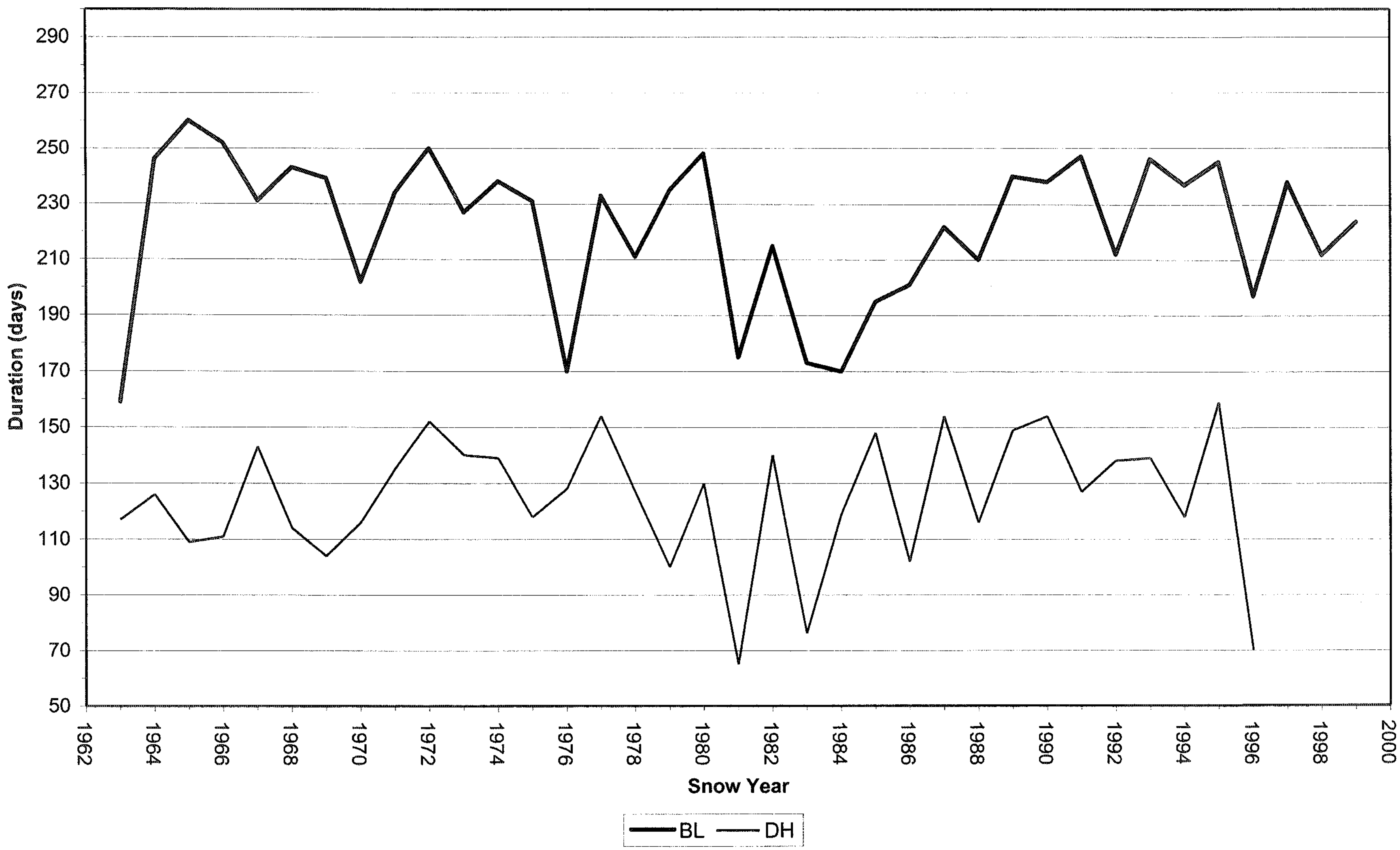


Figure 6.7 **Duration of the continuous snowcover (or SOG) season for BL and DH.** There are no significant trends at either station.

1981 had the warmest estimated February on record from 1946 to 1999 with a mean temperature of -6°C on BL. That month also had 80 mm of rain leading up to JD 44 and only 100 mm of snowfall during the entire month, with most of the new snowfall being quickly ablated through drifting and sublimation on cold windy days between the warm and rainy spells.

An estimated 10 snow years had a continuous snowcover duration of 240 days (eight months) or more (1964, 1965, 1966, 1968, 1972, 1980, 1989, 1991, 1993 and 1995). The longest continuous snowcover season was in 1965 with an estimated duration of 260 days. In contrast, five years had continuous snowcover durations of less than 180 days (six months) (1963, 1976, 1981, 1983 and 1984) with the estimated shortest season in 1963 at 159 days. The snowcover duration record for BL during the study period shows no significant trends.

The snowpack on BL at its largest seasonal depth has an estimated snow water equivalent (SWE) of 490 mm during the 1963 to 1999 period, which occurs most frequently in April. In contrast, the maximum estimated seasonal value at DH on the coastal plain is less than half this value on average at 240 mm SWE. The standard deviation at BL is large at 380 mm (see Figure 6.8 for the estimated maximum SWE values for DH and BL during the study period). Flowing from the discussion of snow density in section 6.1.3, the DH snow depth record was converted to SWE by using mean densities of 200 kg/m$^3$ in the fall up to December, and 350 kg/m$^3$ for the end of the season (May). Snow densities for months in between were estimated using linear interpolation: 233 kg/m$^3$ for January, 266 kg/m$^3$ for February, 300 kg/m$^3$ for March and 333 kg/m$^3$ for April.

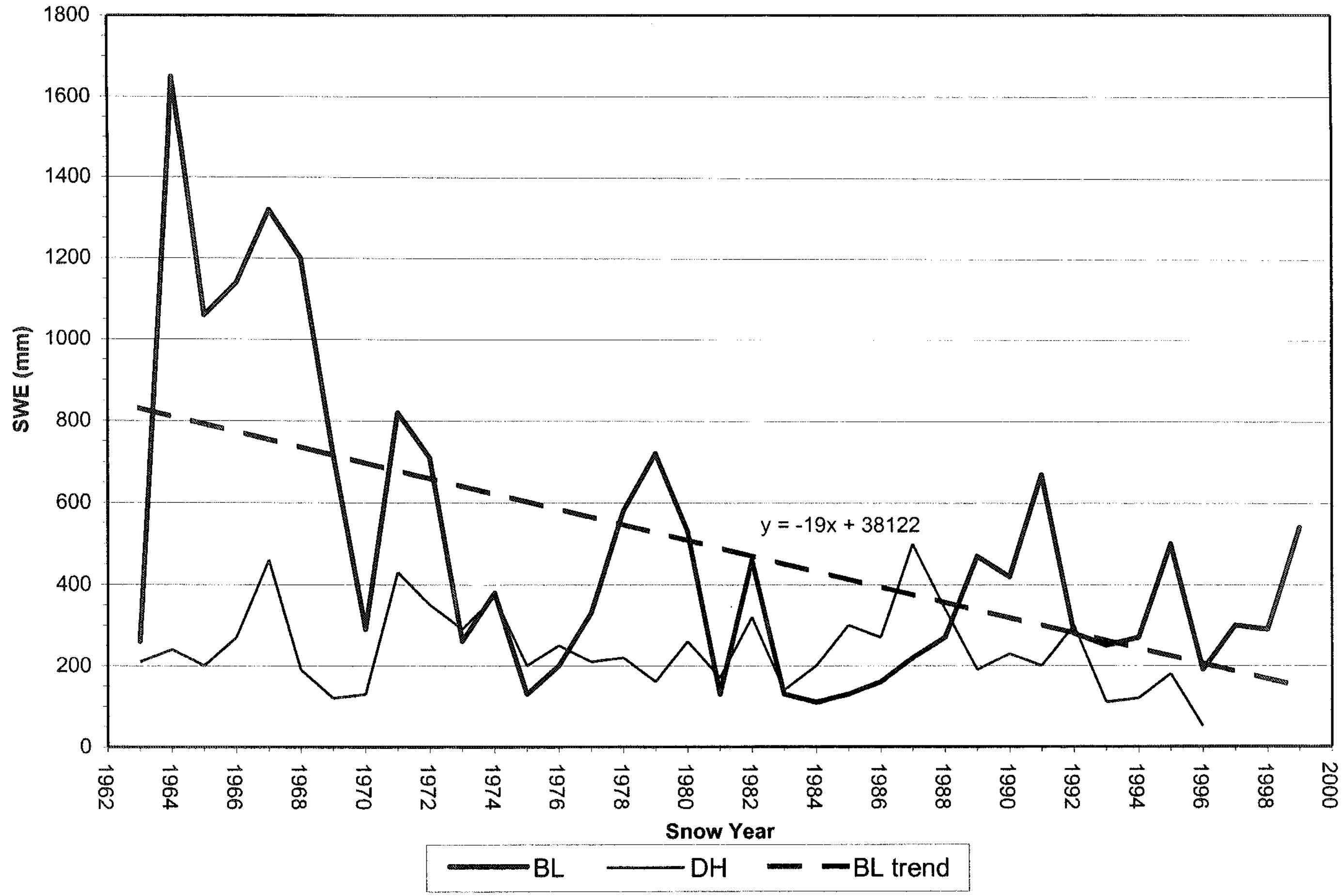


Figure 6.8 **BL and DH maximum seasonal snowpack SWE.** The significant linear trendline equation for BL is indicated on the graph. Due to the probable errors in the BL estimates, caution must be used in comparing with DH.

Six snow seasons on BL had exceptionally large estimated maximum SWEs with 800 or more mm concentrated mainly in the 1960s (1964-68 and 1971). The largest estimated value was estimated to have been achieved in May of 1964 with a value of 1650 mm SWE. Concerning the small maximum SWE amounts, seven snow seasons had maximum values of less than 200 mm, mainly in the 1980s (1975, 1981, 1983-86 and 1996). The lowest maximum value was estimated to have been realized in February 1984, with only 110 mm SWE. There is a significant negative trend during 1963-1999 period of -19 mm per year, leading to a difference of approximately 700 mm between the beginning and the end of the period of interest. This trend is largely governed by the high values of the 1960s and the low values of the 1980s.

For some years, the BL maximum SWE was estimated to be less than the DH maximum value mainly because of the high levels of snowdrifting and sublimation occurring on the plateau that would erase much of the difference between seasonal snowfall and melt quantities at both locations. This was the case in the snow years of 1973, 1975-76, 1981 and 1983-88. Despite the large amount of snowdrifting and blowing snow sublimation that occurs on BL, this outcome is doubtful and again points to an overestimation of wind related ablation. Furthermore, given the error involved in estimating the BL SWE, it cannot be asserted with confidence that BL would have had a smaller maximum SWE than DH during many of those years. Nevertheless, the possibility of more SWE at DH than at BL under certain conditions remains.

Throughout the snow season, the estimated mean snowdrifting loss on BL during the study period was 130 mm WE, with a year to year standard deviation of 50 mm WE. The drifting loss is assumed to be uniform across the plateau. Furthermore, the loss is assumed to be blown off the plateau and unable to return due to steep slopes. Snowdrifting off the plateau accounted for 10% of the estimated snowfall (i.e. a proxy for snowcover ablation) on average. A total of seven snow seasons had estimated transport values higher than 180 mm (1963, 1964, 1969, 1977, 1984, 1994 and 1997), with the highest value estimated to be in the 1997 snow year with 230 mm SWE blown off the plateau. Some snow seasons had considerably less transport. Eleven snow seasons had estimated transport values of less than 100 mm, all but two of them occurring after 1982. The 1983 snow season, a mild winter, had the least amount of transport with an estimated 60 mm removed from the plateau. Over the study period, there is a significant negative trend of snow transport of -2 mm per year, leading to a 70 mm difference between the beginning and the end of the study period.

Sublimation of blowing snow is the largest means of snowcover ablation on BL, estimated to be larger than melt during in all but eight snow seasons (1964-68, 1978, 1979 and 1991). An estimated mean of 730 mm of SWE was sublimated per year on BL during the study period with a standard deviation of 140 mm. This is over five times larger than what is transported off the plateau, and represents 53% of the estimated snowfall on BL. Figure 6.9 shows the estimated blowing snow sublimation and snowdrifting time series on BL for the study period. The lowest and highest estimated blowing snow sublimation totals were only one year apart, with the largest value occurring in 1984 at 1000 mm SWE and the lowest in 1983 with 430 mm SWE. Four snow seasons had totals below 600 mm (1983,

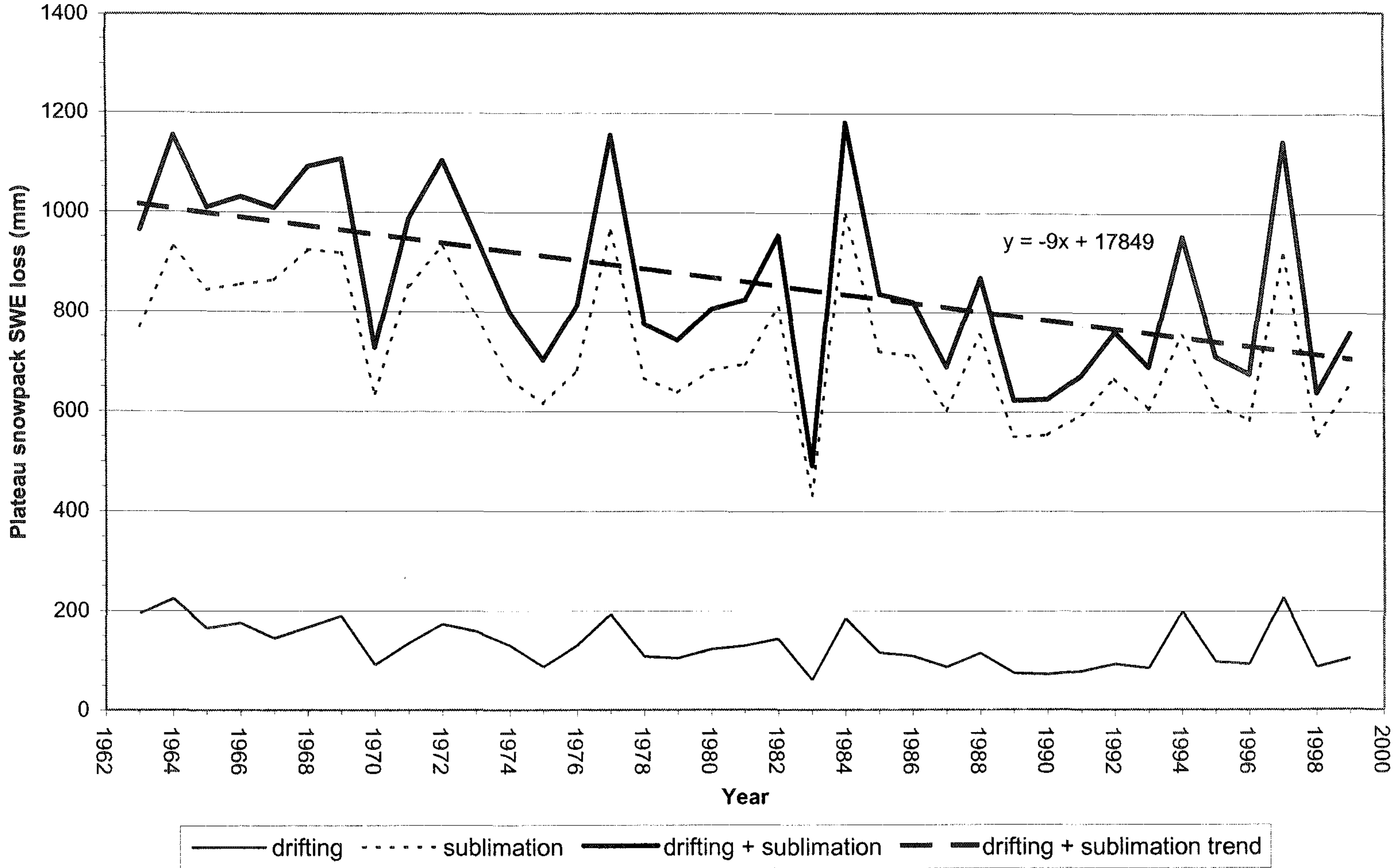


Figure 6.9 **Snowdrifing and blowing snow sublimation loss on BL.** All three quantities have significant negative trends with drifting + sublimation being the largest at -9 mm per year. The line and equation for this trend are shown on the graph.

1989, 1990, 1996 and 1998) while eight had totals above 900 mm (1964, 1968, 1969, 1972, 1977, 1984 and 1997). Over the study period, there is a significant negative trend of -7 mm SWE per year for blowing snow sublimation on BL. This represents a 260 mm SWE difference between the beginning and the end of the study period.

Ablation of the snowpack in the form of snow drifting and sublimation of blowing snow accounted for an estimated average of 900 mm of SWE, or approximately 62% of the estimated yearly snowfall received on BL. As expected due to the individual trends, these two forms of ablation have an estimated significant negative trend over the study period of -9 mm SWE per year, which represents a difference of 330 mm from the beginning to the end of the study period. However, there are reasons to believe that these wind related ablation figures are too high, namely that during several seasons the estimated SWE values are larger at DH than at BL, and also that wind scouring on the plateau was a common modelled occurrence in the first half of the snow season. This is unlikely due to the presence of low lying vegetation, and also because of the wind induced densification of the snowpack that would result from multiple days of blowing snow. There is little doubt however that wind related ablation is a principle mean of shrinking the snowpack on BL, perhaps even larger than melt during most seasons as the modelled results have shown.

Table 6.3 has some additional columns relating to temperature (Chapter 3), wind speed (Chapter 4) and precipitation (Chapter 5). These are not discussed here, but they complement the other important results in the table.

Table 6.4 (a and b) shows the mean monthly SWE statistics for both BL and DH while Figure 6.10 shows the mean values and their standard deviations in graphic form. The standard deviations for DH are omitted for clarity, but it is obvious from Table 6.4 that the standard deviations for the monthly mean SWE values at DH are much smaller. This definitely calls into question the theory that higher altitude sites have more stable snowcover conditions from year to year for Western Newfoundland. Although the snowcover is more consistently present on BL from December until March than at DH, the depth of the snowpack and the timing of the presence of snowcover at both ends of the snow season is far more variable at BL. This seems to be the result of a higher interannual variability in rainfall at higher elevations as well as a higher variability in snowdrifting and blowing snow sublimation.

There are several estimated significant linear trends with respect to the mean monthly or seasonal SWE on BL, though not at DH. These results are shown in Table 6.5. On BL, the months of February to June have significant negative trends for their monthly mean SWE during the study period with a smallest trend in February at -8 mm per year and the largest value in May at -21 mm per year, although much of this trend is caused by the large values observed in the 1960s. In addition, the mean SWE of the snow accumulation season (SAS - November to April) and the snow melt season (SMS - May to June) also have significant negative trends. The month of November also has a significant trend, but in the opposite direction. Its mean SWE has a significant increase of 1 mm per year throughout the study period. Surprisingly, the estimated DH monthly mean SWE record reveals no significant trends. Figure 6.11 shows the mean monthly SWE record during the SAS for BL and DH

Table 6.4a **Big Level mean monthly SWE statistics (in mm) (October 1962- July 1999).**

| | Aug | Sep | Oct | Nov | Dec | Jan | Feb | Mar | Apr | May | Jun | Jul |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Mean** | 0 | 0 | 10 | 40 | 100 | 200 | 280 | 370 | 420 | 390 | 160 | 10 |
| **Standard deviation** | 0 | 0 | 10 | 30 | 80 | 170 | 230 | 330 | 280 | 400 | 270 | 40 |
| **Maximum mean** | 0 | 0 | 30 | 110 | 350 | 770 | 860 | 1180 | 1560 | 1630 | 1120 | 240 |
| **Year(s) of occurrence** | All | All | 1969 | 1989 ,93 | 1968 | 1966 | 1964 | 1964 | 1964 | 1964 | 1964 | 1964 |
| **Minimum mean** | 0 | 0 | 0 | 0 | 10 | 10 | 10 | 20 | 0 | 0 | 0 | 0 |
| **Year(s) of occurrence** | All | All | 1962-63, 65-68, 70, 75, 81-84, 86-87, 90-91, 95, 98 | 1966, 77 | 1984 | 1974, 81, 85 | 1975, 81 | 1975, 81 | 1984 | 1983-84, 86 | 1963, 73, 75, 76, 81, 83-88, 96, 98, 99 | Only 1964, 66-68, 91 were not 0. |

Table 6.4b **Daniel's Harbour mean monthly SWE statistics (in mm) (October 1962- July 1999).**
Includes conversion from depth using the following density factors: Jul-Dec= 0.2, Jan = 0.233, Feb = 0.266, Mar = 0.3, Apr = 0.33, May= 0.35 and Jun = 0.35.

| | Aug | Sep | Oct | Nov | Dec | Jan | Feb | Mar | Apr | May | Jun | Jul |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Mean** | 0 | 0 | 0 | 5 | 30 | 80 | 140 | 150 | 55 | 0 | 0 | 0 |
| **Standard deviation** | 0 | 0 | 0 | 0 | 10 | 20 | 30 | 35 | 20 | 0 | 0 | 0 |
| **Maximum mean** | 0 | 0 | 5 | 15 | 115 | 200 | 370 | 420 | 250 | 35 | 0 | 0 |
| **Year(s) of occurrence** | All | All | 1972 | 1990 | 1976 | 1977 | 1987 | 1987 | 1974 | 1974 | All | All |
| **Minimum mean** | 0 | 0 | 0 | 0 | 5 | 5 | 20 | 5 | 0 | 0 | 0 | 0 |
| **Year(s) of occurrence** | All | All | All except 1972 | All except 1964-65, 67-68, 71-72, 74-76, 78, 81-82, 84, 86-87, 89-90 | 1964, 73, 97 | 1963 | 1996 | 1981 | 1979, 81, 96 | All except 1967, 69, 72, 74, 85, 90 | All | All |

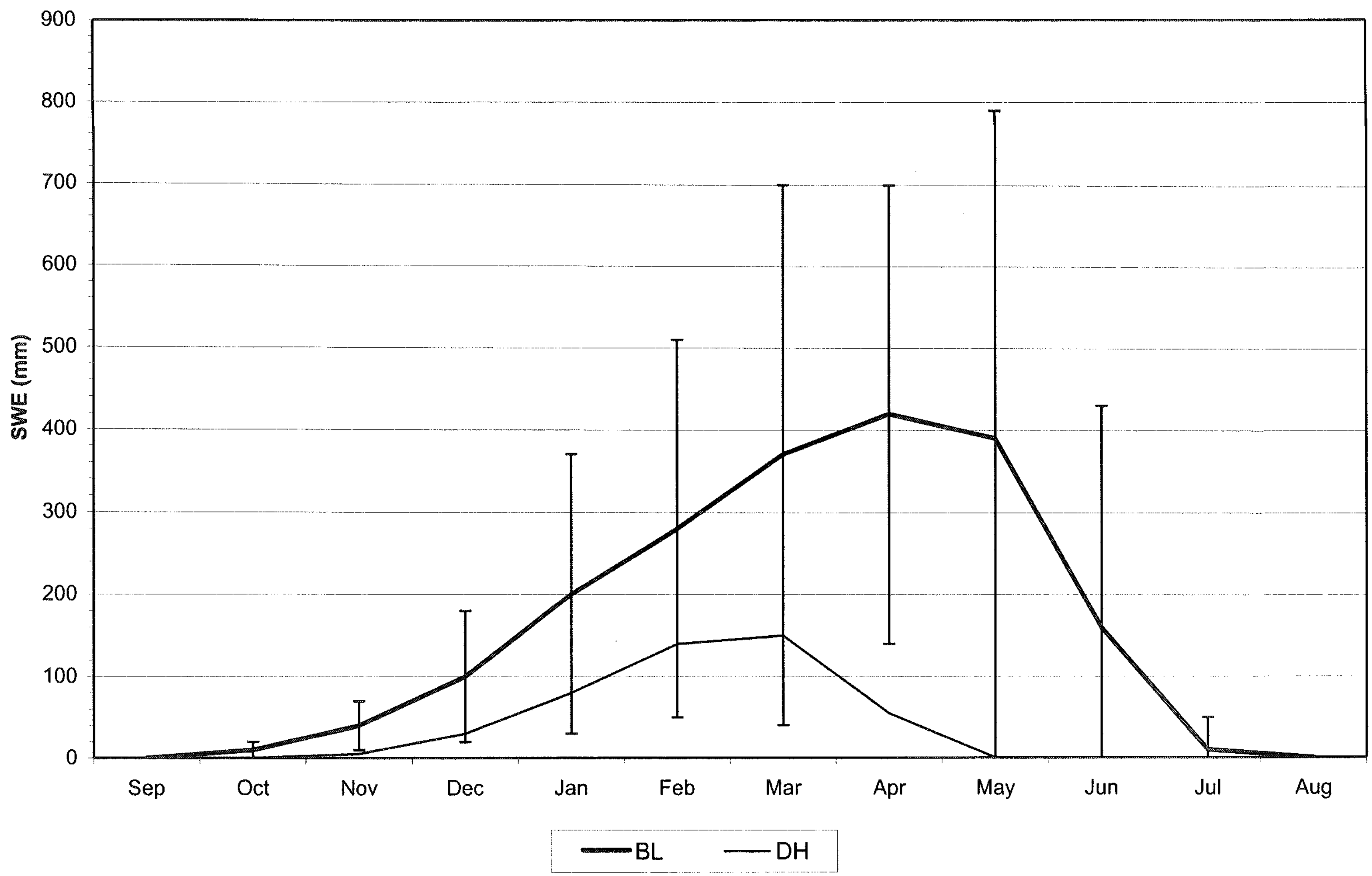


Figure 6.10 **Big Level and coastal plain (DH) monthly mean SWE for the study period (1963-1999).** Error bars indicate the standard deviation of monthly mean values for BL.

Table 6.5 **Significant linear trends for mean SWE at Big Level (BL) and Daniel's Harbour (DH) for various periods.** (snow years 1963-1999)

| | Significant trend at 95% level (mm per year) | |
|---|---|---|
| | **BL** | **DH** |
| October | none | none |
| November | 1 | none |
| December | none | none |
| January | none | none |
| February | -8 | none |
| March | -16 | none |
| April | -19 | none |
| May | -21 | none |
| June | -14 | none |
| SAS* | -8 | none |
| SMS* | -17 | none |

*SAS = Snow Accumulation Season (November - April)

*SMS = Snow Melt Season (May - June)

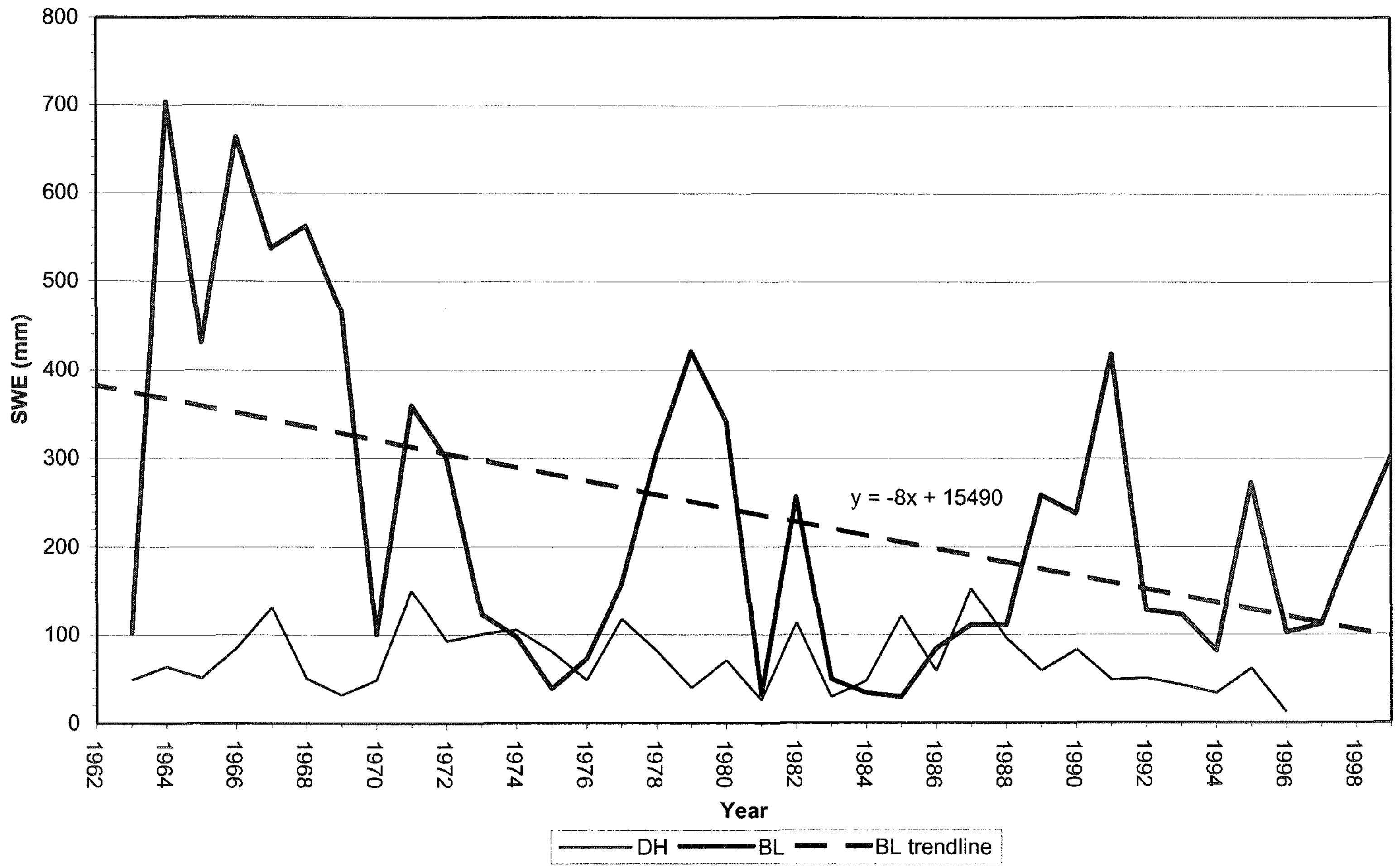


Figure 6.11 **Big Level (BL) and coastal plain (DH) mean snow accumulation season SWE (November-April).** The significant linear trendline and equation for BL is shown on the graph. There is no significant trend for DH.

with BL's estimated significant negative trend of -8 mm per year. Figure 6.12 does the same with the month of May. The estimated significant negative trend in this case is at it's peak with a value of -21 mm per year. The DH record is not shown due to the absence of SWE during this month for most snow seasons.

#### 6.3.3.1 Discussion of Short and Long Continuous SOG Seasons on BL

The estimated absence of snowcover for one day in the middle of winter during 1981 has already been highlighted above. Beyond this characteristic, this season also had one of the shortest estimated continuous snowcover seasons at 175 days. The start of the continuous SOG season was near the period average, but the end was a full two months (60 days) before the average date. This situation was caused by above average mean temperatures (Jan -9°C, Feb -6°C, March -6°C), below average snowfall and above average rainfall from January to March. Furthermore, large estimated drift and sublimation values in late December did much to reduce the SWE that had been put into place early in the season. The stage was set for an early end to the continuous snowcover season at the end of March because the SWE was reduced to only a few mm by the end of that month. With an estimated seasonal maximum SWE of 130 mm, it does not take that much rain or above freezing temperatures to remove the snowpack. However, large snowfalls in late April meant that there would have been significant amounts of SOG until JD 136 (16 May) a month before the average date of continuous snowcover end.

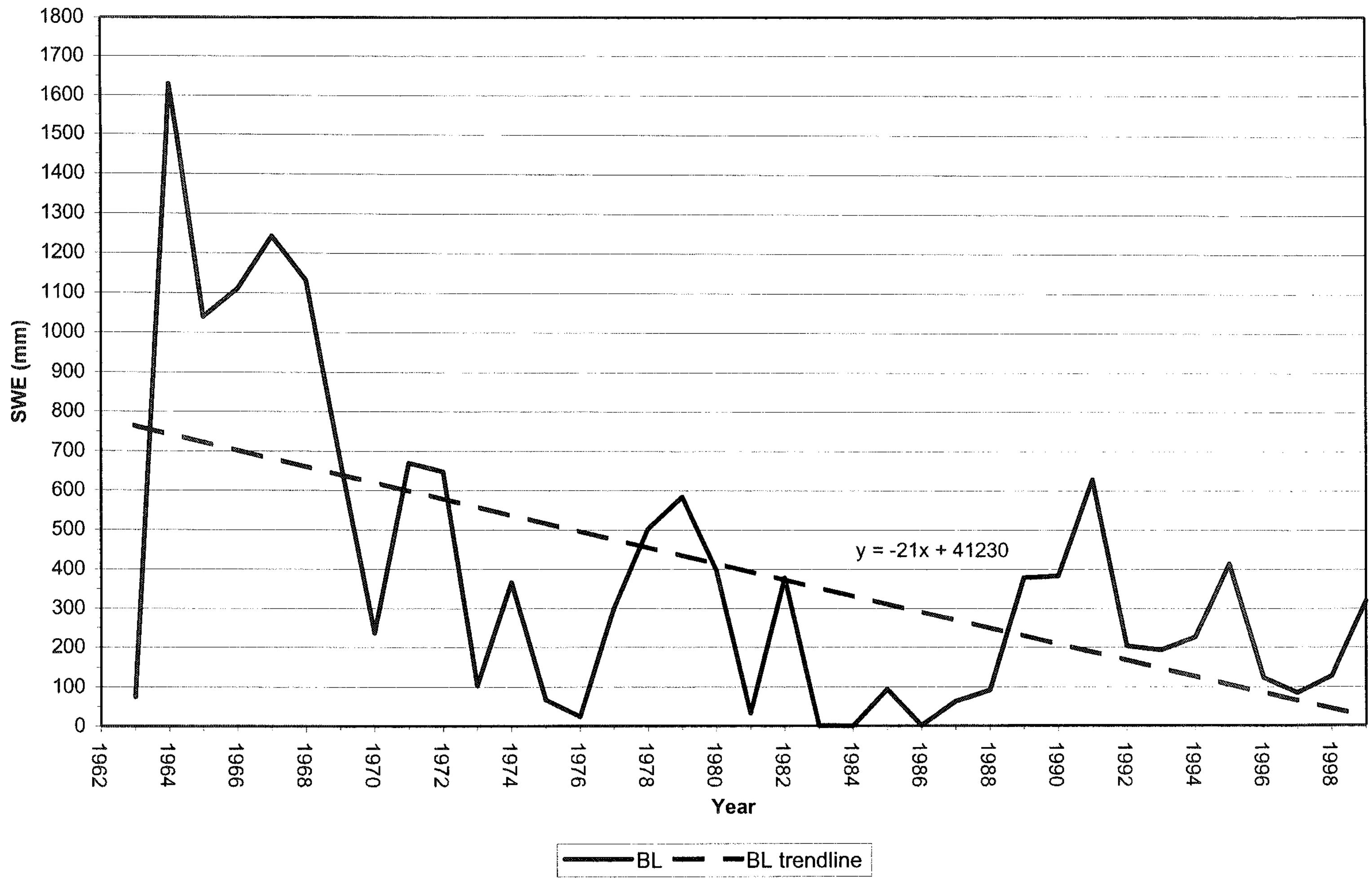


Figure 6.12 **Big Level mean May SWE for the study period.** The significant linear trendline and equation for BL is shown on the graph. Daniel's Harbour does not have significant amounts of snow in May (not shown).

Other short continuous snowcover seasons deserve particular attention. The year 1963 had the shortest season at 159 days, approximately 63 days below average. This continuous season started 46 days after the mean date and ended 17 days before the mean date. October was over three degrees warmer than the mean ($T_{mean}$ 2°C), preventing the onset of the continuous snowcover season. November was also a degree warmer than the mean for that month, but had well below average snowfall at 30 mm SWE and well above average rainfall at 230 mm. Each month for this season other than December and February had below average snowfall. Although this shortfall was mainly overcome by the largest February snowfall in the estimated record (600 mm SWE), since there was no rain or thawing temperatures over half of it was estimated to be lost to drifting and blowing snow sublimation. The maximum snowpack SWE (260 mm) was well below average and the pack was lost under above average May temperatures ($T_{mean}$ 2.5°C), near average rainfall (130 mm) and below average snowfall (10 mm).

The 1976 snow season had a continuous snowcover for 170 days, 52 less than the period mean. This season actually started 13 days before the period mean on JD 288 (15 October 1975). However, its end date was the earliest on record at a value of JD 92 (1 April). This early disappearance was the result of below average snowfall and below average temperatures in February and March ($T_{mean}$ -13.5°C and -10.5°C and snowfall of 110 mm and 90 mm SWE respectively). Drifting and sublimation values were estimated to be larger than snowfall in these two months, resulting in a thin snowcover (max SWE of 200 mm achieved in January) that could not withstand a few days of warm weather at the end of March.

However, given the fact that April was not warm ($T_{mean}$ -4°C) , the snowcover reappeared on JD 102 and lasted until JD 137 (16 May) after which SOG appeared only ephemerally.

The 1983 continuous snow season lasted for only 175 days. It began on JD 312, 11 days after the mean date and ended on JD 119, 38 days before the mean date. As for the other years mentioned above, the maximum seasonal SWE value was well below average at 130 mm (achieved in March). However, the 1983 season differs in that it had the least amount of drifting and blowing snow sublimation during the study period (60 mm and 430 mm respectively). Nevertheless, large values of drifting and sublimation relative to the amount of snowfall occurred in December and January. This snow year had the least amount of snowfall during the study period at 710 mm SWE with all months having below average snowfall. In addition, November, January and April received above average amounts of rain (220, 80 and 70 mm respectively). April 1983 was the warmest on record with a mean temperature of 0.5°C and following an above average March (-6.5°C), this melted away the SOG. However, unlike 1963, 1976 and 1981, no significant amounts of SOG occurred after the end of the continuous snowcover season.

The 1984 season continuous snowcover lasted an estimated 171 days, the third shortest on record. It also had the thinnest maximum snowcover SWE at 110 mm. The continuous snowcover season began on JD 304 which is near the mean, but ended on JD 108, 49 days before the mean date. The reasons behind this particular short season are the largest drifting and blowing snow sublimation loss quantities of the record (1190 mm combined) on top of a near average snowfall season. The thin snowcover was unable to last through the month of April due to a heavy rainfall event on JD 108, and low amounts of snowfall during that month. The rest of the spring was snow free except for an early June snowfall which melted away a few days later.

Thus, from the above examples it is argued that drifting and sublimation of blowing snow is the dominant factor in the regime of a short continuous snowcover season through the maintenance of a thin snowcover throughout the winter. Furthermore, winter month mean temperatures of above -7°C, the threshold above which thawing degree-days become more common on BL, accompanied by low snowfall and high rainfall can also lead to a short season as was the case in 1981. But even in 1981, the high sublimation and drifting values of December played a large role in the early end to the season. One must keep in mind however, that it is believed that the drifting and blowing snow values estimated with the model used here may be too high, potentially having an impact on the results. Nevertheless, the results given here do offer some key insights into the snowcover regime on BL.

The longest estimated continuous snowcover season was determined to be in 1965 lasting 260 days. The dominant factor in this season as with the other large snowcover seasons of the 1960s, is the well above average snowfall values (totalling 2120 mm SWE in 1965). The snowcover season began in mid-October despite a mean temperature during that month of 3°C above the average value for that month. It was then sustained through above average snowfalls throughout the season and attained a maximum SWE of 1060 mm in May. In 1963, March received the largest amounts of snowfall on record (590 mm), while April and May were colder than average (1.0 and 1.5°C below average respectively) and received well below average amounts of rainfall. Therefore, the long lasting and thick snowcover in that year was maintained despite above average amounts of drifting and blowing snow sublimation (1010 mm in total).

Of the other 10 seasons that lasted more than 240 days (8 months) certain patterns are apparent. The long snow seasons in the 1960s (1964, 1965, 1966 and 1968) and in 1972 are due mainly to large snowfall amounts and accompanying large maximum snowcover SWEs, rather than due to particularly cold conditions. In fact, most of these continuous snowcover seasons actually started near or after the period average, but more than made up for this with snowcover disappearances delayed until late June or July. These snow seasons also had larger than average drift and blowing snow sublimation amounts which might normally indicate a thinner snowcover. Later in the study period, the longer continuous snowcover seasons have a different character. The years 1980, 1989, 1991, 1993 and 1999 had a tendency to start early and to finish nearer to the average date. They generally had snowfall amounts near the study period mean, but had less than average drifting and blowing snow amounts. Their maximum SWE values were also close to the average, and they tended to have colder winters than average.

Finally, with regard to the late lying summer snowbeds on BL that form during the snow accumulation season, it is argued that their magnitude has a tendency to be related to the snowdrifting amounts rather than the total snowfall amounts or spring temperatures. For example, although the 1984 season was shorter than average on the plateau, the snowbeds would have probably persisted fairly late into the summer due to the large snowdrift loss (190 mm), accumulating in depressions and on the margins of the plateau. Based on snowdrift amounts, other seasons with particularly late lying snowbeds are estimated to have occurred in 1964, 1969, 1977, 1984, 1994 and 1997. The 1964 snowbeds would have been particularly impressive with a very thick general snowcover lasting until mid-July, depressing the local temperatures below the already sub-average values in July and August. Unpublished digital maps of BL from a GIS database for the mid to late summer period show the position of the late lying snowbeds. The large snowbeds typically reach up to 500 m in width (longest axis, across the slope). The possibility of some snowbeds or patches persisting into the fall season or beyond cannot be ruled out for some years.

6.3.3.2 Possible Implications of Climate Change

The IPCC (2001a - Figure 20), using some of the middle of the road scenarios of climate change, projected an increase of approximately 5°C in average mean temperature for Newfoundland by the end of the 21$^{st}$ century over 1961-1990 normals. Assuming this temperature change would be distributed evenly year round as a conservative estimate of the effects on the winter season, the temperature record at BL would likely resemble the current record at Stephenville A (annual mean temperature of 4.6°C for 1971-2000, (Environment Canada, 2003)). Stephenville receives important amounts or rainfall year round, with the rainfall total representing 70% of the total precipitation rather than the 50% experienced at

BL during the study period (1962-1999). The snow accumulation season at BL would be reduced from December to March, and the snowcover would likely appear continuously in December only, and disappear by mid-May at the latest given that considerable amounts of snowfall would still be expected (740 mm a year at BL, assuming everything else is held constant).

The effects on the extremes would be more pronounced. The cold winters of the future (similar to those of 1989 and 1973) would likely resemble the warmest ones experienced during the 1962-1999 study period, while the average snow accumulation season temperature (November-April) would be 3°C warmer than the warmest ones experienced under the recent climate. In essence, the thin snowcovers experienced in the early 1980's would likely become above average in terms of snowcover amounts and duration.

Over two months of snowcover would be lost, with all the effects on the BL ecosystem that this could imply in terms of favouring warmer climate species. However, the effect on the late lying snowbeds would be major, both because of the lower amount of snowfall received, and also because of the larger amounts of rainfall and thawing temperatures that would reduce the amount of snowdrifting. The effects on the fragile micro-scale ecosystems dependent on the snowbeds would be a serious cause for concern.

# 7 CONCLUSION

The daily climatological record including snowcover snow water equivalent (SWE) was estimated for the Big Level Summit Plateau (BL) in Gros Morne National Park, western Newfoundland, for the 1962-1999 study period from coastal plain station records and the Big Level Autostation record (1993-1999). Widely used statistical techniques were employed in the case of temperature, wind speed and precipitation. Physically based relationships were added for the snow model, which was based on the reconstructed climatological elements named above.

The estimated daily climatological record for BL is included on the accompanying CD (Appendix 2) with the file name *BL daily stats.xls*. The monthly totals and averages are under the name *BL monthly stats.xls*. A sample of the snow model is also provided, named as *BL 1999(sample).xls*.

The model constructed in this thesis can be used to estimate climate and snowcover conditions on BL in the future, from coastal plain records, on a daily basis although the results are principally useful on a monthly and seasonal basis, due to the large uncertainties in the model results. The results of the model must be used with caution, serving mainly as an interannual climate comparison tool, rather than as an accurate predictor of the BL climate on a daily time scale.

The daily temperature record for BL was estimated for 1946-1999 from coastal plain station records and the BL Autostation record, using the multiple linear regression technique. Numerous equations were prepared for all months using different coastal stations, which can be used in the future to provide estimates of BL daily maximum, minimum and mean temperatures. The reconstructed record yielded an average daily mean temperature of -0.7 ±0.1°C, an average daily maximum temperature of 2.6 ±0.1°C and a daily minimum temperature of -3.8 ±0.1°C for the entire period (Table 3.3). BL temperatures were, on average, 3.8 ±0.1°C lower than at Daniel's Harbour (DH) on the coastal plain. The snow accumulation season (SAS - i.e. winter) on BL was identified as being from November to April due to mean daily maximum temperatures being below 0°C during these months. October was identified as a shoulder month (season), with significant snow accumulations possible during many years. The snowmelt season (SMS) was identified as the months of May and June. In a few exceptionally snowy years, the snowcover extended into July during the period of study for the snowcover, confined to 1962-1999 due to limitations by precipitation and wind speed measurements. There were no significant linear trends (95% confidence level) in the temperature record for the period of estimation (1946-1999). However, there was a significant positive linear trend (95% level) for thawing degree-days using maximum temperature of 1.3 - 1.8 degree-days per year for the SAS. There was also a small significant positive trend (95% level) for the number of thawing days for the SAS (0.2°C per year). In spite of these trends, the total amount of thawing degree-days during the SAS remained low compared to the months of October and May.

The hourly wind speed record at BL for the 1962-1999 study period was estimated using linear regression from the DH and the Stephenville upper air (SUA) data. An hourly breakdown of the estimation equations did not lead to significantly better results than single equations for the entire day, i.e. the diurnal signal over the course of the year was weak. The average monthly wind speeds on BL varied from 29 to 36 km/h (±0.5 km/h) with the highest values for December and January and the lowest values in May and June (Table 4.3). Wind speeds were on average 8 ±0.5 km/h higher at BL than at DH. The periods of October, the SAS and the SMS all had negative significant linear trends (95% level) for average wind speed, with values of -0.23 km $h^{-1}$ $yr^{-1}$, -0.12 km $h^{-1}$ $yr^{-1}$ and -0.15 km $h^{-1}$ $yr^{-1}$ respectively. The 1960s was a windy decade, whilst the years with coldest winter seasons tended to have the lowest wind speeds, probably due to more frequent cold anticyclonic air circulations over Gros Morne National Park (GMNP).

The daily precipitation record at BL for the study period (1962-1999) was estimated using mean precipitation ratios between coastal plain stations' record and the Big Level Autostation record. The estimated BL winter precipitation record was separated into snowfall and rainfall based on the estimated temperature record and the SUA freezing levels, and the measured BL precipitation values estimated to be snowfall were corrected for wind induced undercatch. The mean estimated precipitation ratio (BL: coastal plain) was approximately 1.9 ±0.3. Precipitation measurements are made from primitive and imprecise instrumentation. The uncertainty given above is a monthly estimate and would be smaller for longer time periods (see pages 97-98 and section 5.3.1 for a discussion of this). The estimated mean annual precipitation total at BL is 2610 ±60 mm, divided roughly equally between rainfall and snowfall (Table 5.2). December and January typically have the largest snowfall values. There are significant negative precipitation and snowfall trends (95% level) for the SAS, with values of -10.8 mm (SWE) per year and -9.2 mm per year respectively. The large amounts of the 1960s heavily influence these. There was also a significant linear positive trend (95% level) for SMS rainfall of 3.6 mm per year. The SMS rainfall pattern seems to follow a decadal cycle in addition to its increasing trend.

The BL daily snowcover SWE was estimated using the estimated daily precipitation record and the temperature index method, using increasing degree-day factors as the snow season progressed. Snowdrifting off the plateau and sublimation of blowing snow were also estimated using simple expressions that were a function of wind speed only. Sublimation directly from the snowpack was judged to be a small portion of total snowfall (40 - 60 mm per year) and was thus ignored. The method used to calculate SWE was compared with a few measurements taken on site in the mid to late 1990s, and agreed reasonably well given all of the estimations involved in reaching the final product (temperature, wind speed, precipitation, wind related ablation and melt).

The single largest factor influencing the snowcover SWE, other than snowfall, was found to be blowing snow sublimation, which ablates approximately 53% of the snowfall accumulation on average. An estimated average of 730 mm per year was ablated by blowing snow sublimation during the study period. Values of SWE and ablation are crude and should not be considered more accurate than ±100 mm. The numerical values are more properly used as relative indicators between the various components and for year to year comparisons. Mean blowing snow sublimation figures were accompanied by a significant linear trend (95% level) of -7 mm SWE per year, as a function of wind speed according to the methodology used. Another major factor influencing snowcover SWE is snowdrifting off the plateau, accounting for approximately 10% of the average accumulated snowfall, leaving approximately 400 mm to be ablated by melt (32% of the snowfall accumulation). Thus, wind related ablation is more important than melt during most seasons on BL.

Table 6.3 shows the results pertaining to each snow season of the period of study. The first appearance of snow on the ground (SOG) is typically around JD 282 (9 October) and snow becomes seasonally permanent (i.e. continuous) around JD 301 (28 October). The continuous snowcover season lasted an average of 222 days per year (standard deviation of 27 days), which is 3.5 months longer than on the coastal plain. There were no trends for the snowcover season length, or for the appearance of the snowcover. However, there was a negative significant linear trend for the last occurrence of snow on the ground (SOG), with a value of 0.5 days earlier per year (18.5 day difference between the beginning and the end of the study period), varying around an average value of JD 163 (12 June). The mean maximum snowcover SWE was estimated to be 490 mm for BL. The maximum estimated value was 1650 mm SWE in 1964, while the minimum was 110 mm SWE in 1984.

Trends in the monthly average SWE values were presented in Table 6.5. The months from February to June all showed negative significant trends (95% level) with respect to mean SWE, with the trends increasing as the season progresses (minimum: Feb -8 mm SWE per year, maximum May -21 mm SWE per year). Interestingly, the DH snow depth record did not show any trend over the 1962-1999 study period.

It is clear from this thesis that although the snowcover on BL is consistently present for long periods from year to year, its thickness varies considerably compared with the coastal plain. It was found that an average monthly temperature of -7°C is an important value, as months exceeding this temperature tend to have more thawing events, with these events being of larger magnitude as the temperature increases above this point. However nearly all months from December to March during the study period experienced temperatures well below this threshold value. Exceptions do exist, however. The 1981 snow season is a case in point as warm temperatures for February and March (averaging -6°C) were associated with below average snowfall and above average rainfall.

A 5°C annual warming for Newfoundland could be seen by 2100 due to a global increase in greenhouse gas concentrations in the atmosphere. In such a world, a snow season like that of 1981 season would likely become above average in terms of snowcover SWE and duration. It is also evident that one of the most important factors with respect to SWE on BL is sublimation of blowing snow, which tended to be greater with snowier and windier seasons. Future seasons with below average snowfall could also be occasionally associated with windy conditions and above average sublimation as in 1984, leading to a very thin snowcover or even snow free periods on the plateau in mid-winter. Any remaining snowcover would disappear quickly with a warm spring. A mean temperature of 5°C above the study period average would likely reduce the snow accumulation season to December-March and result in the disappearance of the snowcover by mid-May at the latest. Warming on this order of magnitude would result in more winter rainfall, less winter snowfall and less snowdrifting on BL due to more snowcover particle cohesion, with major impacts on late lying snowbeds and associated ecosystems.

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## Appendix 1: Calculations in the Snowpack SWE Model

The Big Level snowpack SWE was calculated with an Excel spreadsheet for each season on a daily timescale. The various steps undertaken are discussed here as an accompaniment to Chapter 6. Essential variables brought over from previous chapters were the daily mean temperature ($T_{mean}$) , daily total rainfall (R) and snowfall (S) and, indirectly through blowing snow quantities, hourly mean (10 m) wind speed ($u_{10}$). The theoretical underpinnings of this model are discussed in Chapter 6.

### Potential Snowdrifting

The potential amount of snowdrifting (i.e. blowing snow, or snow transport) was calculated for BL using the hourly mean wind speed and the snow transport equation presented by Tabler *et al.* (1990, see reference section). The initial quantities were in kg per meter across the wind per second for fetch distances of 3 km or more. The fetch on BL was assumed to be 4 km, the approximate diameter of a circle with an area of 12 $km^2$. The blowing snow quantities were assumed to be blown off the plateau, unable to return and the loss was assumed to be uniformly distributed on the plateau. The blowing snow quantities were multiplied by 3600 seconds to give hourly quantities and then summed over 24 hours to give daily potential blowing snow quantities. The descriptor "potential" assumes that enough snow is available to yield the calculated amount

The potential blowing snow loss was converted into a mm SWE loss to the snowcover by assuming that 1000 kg of water are contained in 1 $m^3$ and by assuming an upwind area of 4000 $m^2$ (4000 m long x 1 m wide). The resulting depth in m was then converted into mm by multiplying by 1000. Thus, in summary, the potential blowing snow quantity in kg $m^{-1}$ $d^{-1}$ was divided by 4000 to yield potential SWE loss in mm.

## Snow Available for Transport

The potential amount of snowdrifting needed to be measured up against what was available for transport (A). If the $T_{mean}$ for any given day was below the -2°C threshold and no rainfall was estimated, the snow available for transport on that day was assumed to be:

$$A_n = A_{n-1} - Dr_{n-1} + S_n \qquad \text{Equation A}$$

where:
$A_n$ = snow available for transport on any given day n (mm SWE)
$A_{n-1}$ = snow available for transport on the previous day
$Dr_{n-1}$ = drifting loss on the previous day (mm SWE).
$S_n$ = snowfall (mm SWE) on the given day.

However, if the $T_{mean}$ rose above the threshold or rainfall was estimated to have fallen, the snow on the ground (SOG), if any, was assumed to be locked away by a wet layer at the surface that would subsequently freeze into an ice layer also making the SOG unavailable. In this case, "A" was therefore assumed to be 0 mm.

Actual Drifting

In this step the snow available for transport (A) was compared to the potential calculated snow drifting. The actual drifting amount was not allowed to surpass the amount available for transport.

Blowing Snow Sublimation

Blowing snow sublimation is dependent on drifting snow. If actual drifting was 0 mm, than so was blowing snow sublimation. The blowing snow sublimation equation used was presented by Tabler *et al.* (1990) modified after Pomeroy (1988). It yielded quantities in mg $m^{-2}$ $s^{-1}$. This expression was multiplied 86400 s/d and divided by 1000 mg/g and 1000 g/$m^2$ to give blowing snow quantities in mm SWE averaged over the plateau. Within the expression, the actual drifting amounts needed to be converted from mm SWE/d to g $m^{-1}$ $s^{-1}$ by multiplying q by 1000 g/kg, 4000 kg $m^{-1}$/mm SWE and 1 d / 86400 s. Furthermore, the amounts of sublimation of drifting snow were limited between 0 and 15 mm SWE per day. The upper bound was determined through calibration with the Big Level snowcover measurements as discussed in Chapter 6.

## Melt Factors

The melt factors used for the degree-day melt equation for days with no rainfall varied from 1.5 mm $^{\circ}C^{-1}$ $d^{-1}$ for the start of the snow season until January to a value of 4.5 mm $^{\circ}C^{-1}$ $d^{-1}$ in July. The values increased by 0.5 mm $^{\circ}C^{-1}$ $d^{-1}$ per month from February to July. The base temperature was $T_{mean}$ = -2°C.

## Melt Potential Equations

Melt potential was distinguished from actual melt due to two main factors. First, enough snow must be present for all the melt potential to be realized. Secondly the cold content and free water content conditions of the snowpack must be satisfied.

For days where there was rainfall with a $T_{mean}$ of above 0°C, the expression used to calculate snowmelt potential (MP) in mm / d was:

Equation B

$$MP = 0.5*((0.074+0.007*(R*0.03937))*((9/5*T_{mean}+32)-32)+0.05)/0.03937$$

where:

R = daily total rainfall (mm).
$T_{mean}$ = daily mean temperature (°C).

For days with rainfall but with a $T_{mean}$ of 0°C or less, the $T_{mean}$ was assumed to be 1°C in order for the equation to work. For days without rain, the potential melt was calculated with the following equation on days with $T_{mean}$ above -2°C:

$$MP = MF * (T_{mean}+2) \qquad \text{Equation C}$$

where:

MF = melt factor which varies according to the month of the year as a proxy for net shortwave radiation (mm $°C^{-1}$ $d^{-1}$).

Snow Cold Content

The cold content (i.e. heat deficit) of new snow from the current days snowfall was assumed to be a function of the amount of snowfall received and the mean temperature of the air for values below 0°C:

$$NCC = 2.09 \text{ kJ mm}^{-1}\ °C^{-1} * S*T_{mean} \qquad \text{Equation D}$$

where:

NCC= new snow cold content (kJ)
S = current day's snowfall (mm SWE)
$T_{mean}$ = daily mean air temperature (°C)

For older snow, the cold content was a cumulative function of the amount of snow on the ground and the weather from previous days. The old snow cold content was the cold content left over from the previous day's conditions in kJ and corresponded to a snow mean temperature. Once the current day's cold content was added up (old and new snow), the melt potential was subtracted from the cold content by converting the mm SWE potential melt values into kJ using the latent heat of fusion of snow (333.5 kJ/ kg or kJ/mm SWE) and a thermal quality of 0.95. When the $T_{mean}$ was below -2°C, a different melt factor equation was used to change the cold content of the snowpack. That step was accomplished using the following equation:

$$\Delta CC = 0.5*MF*(T_{mean} - T_{osnow})*333.5 \qquad \text{Equation E}$$

where:

$\Delta CC$ = change in cold content (kJ)
MF = (negative) melt factor 2 mm $°C^{-1}$ $d^{-1}$.
$T_{osnow}$ = temperature of the old snow (°C).

Snowpack cold content was linked to snow temperature through the 2.09 kJ $mm^{-1}$ $°C^{-1}$ constant an the estimated SWE. The cold content (CC) change could not lead to a snow temperature that was below the air temperature when CC is increasing, while a CC decrease could not lead to a snow temperature that is above the air temperature. When the equation would have led to such an occurrence, the snow temperature was capped at the prevailing air temperature, and that temperature was converted to a cold content in kJ. The snow temperature could not exceed 0°C at which point the CC was assumed to be 0 kJ.

The cold content of the snowpack was assumed to be 0 kJ if liquid water was present. If there was liquid water in the snowpack due to rainfall or superficial melt, that water needed to be frozen inside the pack before the cold content of the snowpack could start increasing as shown in the following equation:

$CC_f = CC_i + R * 333.5\ kJ/mm + WC * 333.5\ kJ/mm$ Equation F

where =

$CC_f$ = snowpack cold content after accounting for internal water refreeze (kJ).
$CC_i$ = estimated cold content before accounting for internal water.
R = rainfall (mm)
WC = water content of the snow (mm)

## Liquid Water Requirements

Before the snowpack could start ablating through melt, five percent of the mass of the snowpack needed to be converted to liquid water. The liquid water requirements were expressed through the following equation:

$WR = 0.05 * (S + SWE - Dr - Sub)$ Equation G

where:

WR = liquid water requirements for melt to begin (mm).
S = current day's snowfall (mm SWE)
SWE = snowpack snow water equivalent at beginning of day (mm)
Dr = SWE lost to drifting during the current day (mm)
Sub = SWE lost to blowing snow sublimation during the current day (mm).

## Rainfall adding to SWE

When the cold content of the snowpack was large enough to absorb the heat of the rain as expressed through equation B, converted to ice and added to the snowpack SWE. The rain that was retained in liquid form within the snowpack also added to the SWE. If some ablation occurred, that portion was subtracted from the addition to the SWE.

Real Melt

Once the CC and liquid water requirements were subtracted from the potential melt, whatever heat remaining was assumed to be real melt and ablated the snowpack. The heat in kJ was reconverted to mm SWE melt using the following equation:

$$M = Q_m / (333.5 \text{ kJ/mm} * 0.95) \qquad \text{Equation H}$$

where:

$M$ = real melt occurring (mm SWE)
$Q_m$ = energy remaining to melt the snowpack (kJ)

Finally, if M was larger than the snowpack SWE, the M value was capped at the SWE value, leaving no snowcover behind.

Snowpack SWE

The snowpack SWE value was then recorded at the start of the subsequent day. The equation for calculating the SWE of the following day was thus:

$$SWE_{n+1} = SWE_n + S + R_a - Dr - Sub - M \qquad \text{Equation I}$$

where:

$SWE_{n+1}$ = snowpack SWE present at the beginning of the following day (mm)
$SWE_n$ = snowpack SWE present at the beginning of the current day (mm)
S = snowfall (mm SWE)
$R_a$ = rainfall added to the SWE (mm) (only if RM = 0)
Dr = snowdrift loss (mm SWE)
Sub = blowing snow sublimation loss ( mm SWE)
M = real melt (ablation) after CC and liquid water requirements fulfilled (mm SWE).