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Authors: Pigeon, Karine E., and Jiskoot, Hester

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Meteorological Controls on Snowpack Formation and Dynamics in the Southern Canadian Rocky Mountains

Karine E. Pigeon and Hester Jiskoot*†*

*Department of Geography, University of Lethbridge, Lethbridge, Alberta, T1K 3M4, Canada
†Corresponding author: hester.jiskoot@uleth.ca

Abstract

Considerable spatial variability in snow properties exists within apparently uniform slopes, often resulting from microscale weather patterns determined by local terrain. Since it is costly to establish abundant weather stations in a region, local lapse rates may offer an alternative for predicting snowpack characteristics. For two Castle Mountain Resort weather stations, we present the 2003–2004 winter season weather and snow profile data and the 1999–2004 winter season lapse rates. A third site was sampled for small-scale spatial variability. Layer thickness, stratigraphy, temperature gradients, crusts, wind drift layers, stability, and settlement were compared between the sites and correlated with temperature, wind, and lapse rates. Average yearly snowfall was 470 cm at the Base and 740 cm at the Upper station. Average daily maximum and minimum temperature lapse rates are $-6.1°C\ km^{-1}$ and $-5.7°C\ km^{-1}$ when inversions are removed. Inversions occur mostly at night, adversely affecting lapse rate averages. Lapse rate modes are unaffected and most often $-6.3°C\ km^{-1}$. Snowpack spatial variability is ~25% of layer thickness and is controlled by wind and topography. Layer settlement is primarily related to initial snow thickness and wind drift. Snowpacks stabilize with age, unless rain crusts are present, which are important low-force failure horizons.

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Introduction

Snowpack spatial variability within a small mountain range or single slope has been studied extensively in North America and Europe (e.g. Birkeland et al., 1995; Harper and Bradford, 2003; Landry et al., 2004; Schweizer et al., 2006). Since considerable spatial variability exists within small, apparently uniform slopes (Landry et al., 2004) and even on flat terrain (Harper and Bradford, 2003), it is evident that microscale weather patterns determined by local topography can noticeably affect local snowpack formation. Knowing that the establishment of abundant weather stations over small areas is not a cost-efficient option to predict microscale snowpack developments, local lapse rates may offer a viable alternative. Simple predictions using local lapse rates could benefit small scale ski operations by serving as an indication of snowpack variability, leading to proper and efficient slope management and reducing the error in spatial extrapolations of snowpack stability test results.

An environmental temperature lapse rate of $6.5°C\ km^{-1}$ is often used to predict the elevation-related distribution of biological and geographical factors, including snowpack properties resulting from spatial distribution of snowfall, rainfall, wind drift, radiation balance, etc. However, this theoretical environmental lapse rate may not consistently suit mountainous regions because of the variability of topographical influences on meteorological elements such as temperature, precipitation, wind speed, and solar radiation (Rolland, 2003). Few studies (Pielke and Mehring, 1977; Bolstad et al., 1998; Rolland, 2003; Shea et al., 2004; Thayyen et al., 2005) have used measured lapse rates in mountainous regions, mainly because of a lack of weather stations and balloon data. Rolland (2003), using 640 stations in the southern European Alps over a period of 30 years, concluded that yearly temperature variations were regional and topographically controlled, while seasonal patterns were similar and had consistently higher summer lapse rates. However, many publications lack sufficient data in either years or weather stations to accurately depict lapse rates (Shea et al., 2004).

Steady snow layer settlement is generally an indication of densification and an increase in snowpack strength. While very low settlement rates indicate persistent potential instability, very high rates are associated with avalanche activity (McClung and Schaerer, 1993). The main factors influencing snow settlement are initial snow density, temperature, and snow and wind loading. At the mesoscale level, temperature, precipitation, and solar radiation vary with slope angle, aspect, and elevation (Shea et al., 2004). The same is true at the microscale level (<1 km), which is depicted in the single slope snowpack spatial variability found by Landry et al. (2004). Weather factors associated with avalanche occurrences and snow stability are assessed in a number of recent publications. Davis et al. (1999) ranked storm snowfall depth, snow water equivalent, wind-drift parameters, and yearly initial snow depth as important factors influencing dry slab avalanche activity in Utah and California, while Jones and Jamieson (2001) found 24-hour air temperatures and snowfall, as well as total and storm snowpack depths, the most relevant snow instability forecasting variables for the Columbia Mountains, British Columbia. Kozak et al. (2003) found the south temperature index (whereby degree-days above $-10°C$ are added over an index period to predict increases in settlement, density, and sintering caused by warm temperatures), maximum daily temperatures, and incoming shortwave radiation to be important predictors of new snow layer hardness on S-facing slopes in Wyoming. For new snow on N-facing slopes, maximum daily temperatures and previous day's

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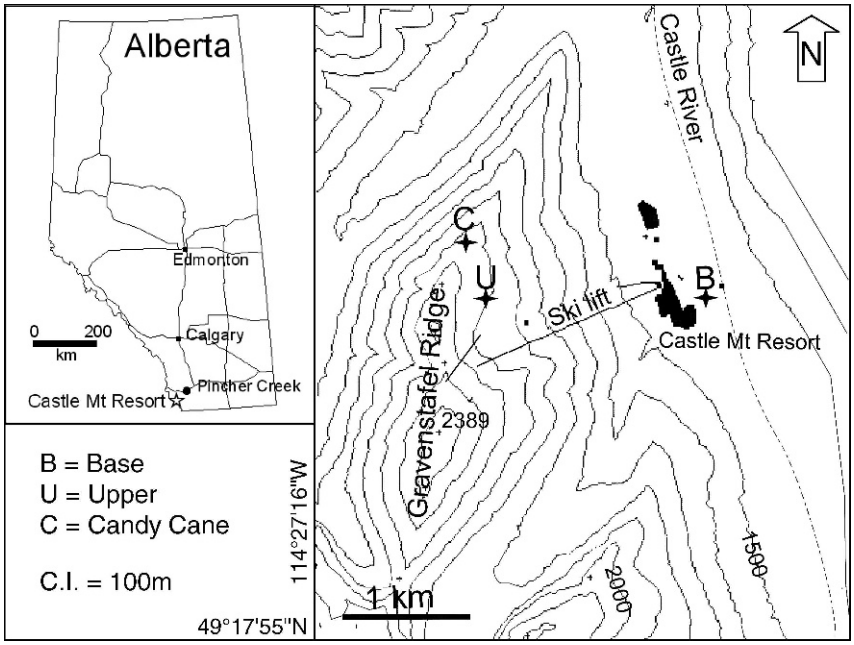

FIGURE 1. Castle Mountain Resort map with Base (B), Upper (U), and Candy Cane (C) weather station and snow profile locations. C.I. is contour interval.

wind speed ranked highest. For older snow layers, only the temperature index was ranked as a significant hardness predictor (Kozak et al., 2003). Reviewing these findings suggests that a good knowledge of local area lapse rates and meteorological processes can be combined with snow profiles to potentially provide simple, cost-efficient predictions of snowpack dynamics and properties.

The central aim of this paper is to assess the impact of local weather conditions on snowpack dynamics in a small study region on the eastern slopes of the southern Canadian Rocky Mountains. During a 9-week period in winter 2004, snow profiles and weather observations were recorded at Castle Mountain Resort at two locations with 630 m vertical and ~1.5 km horizontal distance. Additionally, a 6-year period of winter temperature data are presented for the same sites, and a one-day snowpack spatial variability study was conducted on the ski hill.

Study Area

In order to test the meteorological and local lapse rate controls on snowpack properties, a study was established at Castle Mountain Resort, Alberta (49°19′N, 114°25′W; Fig. 1), during winter 2003/2004. This ski resort is located in the Westcastle Valley, formerly the Rocky Mountain Forest Reserve, southern Canadian Rocky Mountains, east of the continental divide and about 25 km northwest of Waterton Glacier International Peace Park. It is affected by a typical continental climate yielding relatively low winter precipitation (<8 m) and cold winters (McClung and Schaerer, 1993). However, so-called "chinook events" bring strong westerly winds and warm air to the region, and cause rapid snowmelt on the prairies. These events occur frequently throughout the winter season and disturb Castle Mountain Resort by bringing strong winds and snow drift, especially at higher elevations, and above-zero temperatures at the base of the mountain.

No previous snowpack or local weather data were ever published for this region, and meteorological records only go back to winter 1999. Two stations (Base and Upper) on Gravenstafel Ridge were used to collect all weather and snowpack data. The Base weather station (1410 m a.s.l.) is located in the "village" area of the resort and the Upper weather station (2040 m a.s.l.) is on the ski hill. The horizontal distance between the stations is about 1.5 km. Both stations are somewhat sheltered by trees. The Base station is in relatively flat, hummocky terrain (average slope 0°), while the Upper station is on a uniform E-facing slope of 14°. Additionally, in order to evaluate snowpack spatial variability on a uniform slope, a single day multi-profile campaign was performed on "Candy Cane," the upper section of an out-of-bounds N-facing run with a 31° slope. This site encompasses a 20 × 20 m area and ranges in elevation from 2050 to 2070 m a.s.l.

Methodology

Daily weather and weekly snowpack data were collected from the Upper and Base weather stations between 9 January and 7 March 2004 (9 weeks). Weather observations were taken twice daily (8:00 and 16:00) from the Base and once daily (12:00) from the Upper station. Maximum and minimum temperatures were automatically recorded for the periods between these observations. Long-term temperature lapse rates were calculated from the recorded daily maximum and minimum temperatures and weather observations obtained for the 6-year period during which both weather stations were operational (winters 1998/1999 to 2003/2004).

SNOW PROFILE AND WEATHER DATA

This paper uses the Canadian Avalanche Association (CAA) snow profile and weather data collection methods defined in the "Observation Guidelines and Recording Standards for Weather, Snowpack and Avalanches" (CAA, 2002). Table 1 lists all weather variables collected during each daily or twice-daily visit to the weather stations. Daily snowfall data were collected using snow boards and storm boards. Temperature and humidity were measured using the weather station's thermistors, correlating

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TABLE 1

Weather and snow observations collected daily for the Base and Upper weather stations.

Measurement	Method	Abr./unit
Cloud cover fraction	(0/8) clear, (8/8) overcast	Sky/0/8–8/8
Snowfall rate	(S-1) <1 cm/hr, (S4) >4 cm/hr	S-1 to S4
Maximum air temperature	24 hr interval	T_{max}/°C
Minimum air temperature	24 hr interval	T_{min}/°C
Air temperature at time of observation	Nearest 0.5°C	T_{pres}/°C
Relative humidity	Nearest percent	RH/%
Snowpack temperature at 10 cm depth	Nearest 0.5°C	$Ts_{10\ cm}$/°C
New snowfall amount	Nearest cm	$NS_{24\ Hr}$/cm
New snowfall amount since storm start	Nearest cm	S_{Storm}/cm
Snowpack height	Nearest cm	HS/cm
Surface form type	Symbol (CAA, 2002)	Sfc_{Form}/symbol
Surface form size	Nearest 0.5 mm	Sfc_{Size}/mm
Wind speed at time of observation	Calm, light, moderate, strong, extreme	$Wind_{Spd}$/nominal
Wind direction at time of observation	S, SE, E, NE, N, NW, W, SW	$Wind_{Dir}$/nominal
Blowing snow occurrence and direction	Present, absent, intermittent, previous	BS/nominal
Barometric pressure at time of obs.	Nearest mb	P/mb
Barometric pressure trend	Rising, falling, stationary	P_{trend}/arrow

thermographs, and hygrographs. These instruments were enclosed in a Stevenson screen at 1.5 m above the snow surface. The Stevenson screen was raised according to new snowfall throughout the season. Nominal wind speeds and directions were recorded daily at the two stations using the CAA (2002) five-category ranking system (Calm: 0 km h^{-1}; Light: 1–25 km h^{-1}; Moderate: 26–40 km h^{-1}; Strong: 41–60 km h^{-1}; Extreme: >60 km h^{-1}). Snow drift occurrence and direction were logged concurrently. Additionally, for comparison and verification, numerical wind speeds and directions were obtained from the Pincher Creek weather station (Environment Canada, 2005) located approximately 40 km northeast of Castle Mountain Resort.

Weather station snow profiles were assessed weekly, on Mondays, except for 31 January and 7 February. Each week a new snow profile was dug ~30 cm behind the previous week's, with a total of nine profiles over the study period. At the Candy Cane site, seven snow profiles were recorded on 25 January 2004. All profiles were recorded by the same individual (K. Pigeon) in order to limit human-induced interpretation variations. Eight snow profile variables were collected: snowpack height, snow temperature, stratigraphy, crystal types and sizes, layer resistance, density, and shovel compression test results (CAA, 2002). However, due to irregularities in the snow density measurements, these could not be used for further analysis. Snowpack height was measured from the ground up to the nearest cm. Temperatures were measured in 10 cm increments using digital snow thermometers calibrated in an ice/water mixture. Snow crystals were considered using a 10× magnifying lens and measured to the nearest 0.5 mm. Layer resistances were assessed using the CAA (2002) hand hardness categories from softest to hardest: fist (F), four fingers (4F), one finger (1F), pencil (P), knife (K), and ice (I), making 14 numerical categories by adding + to categories F and K, and − or + to 4F to P. For the compression tests, force (number of taps), depth of fracture, and type of fracture were recorded (CAA, 2002).

The Candy Cane site encompasses about 400 m^2. Profiles were organized by staggering pairs of snow pits approximately 5 m apart, with numbers 1 and 2 being the lowest pair at 2050 m a.s.l., 3 and 4 the next pair up, 5 and 6 the next, followed by a single plot 7 at 2070 m a.s.l. There is considerable disagreement about the optimal spacing of snow pits used for testing of spatial variability of snow properties and stability, varying from <5 m to >10 m (Conway and Abrahamson, 1988; Harper and Bradford, 2003; Schweizer et al., 2006). Apart from process-related optimal spacing, logistical considerations suggest that a maximum of 5 pits in a 30 × 30 m area are representative of an avalanche forecaster's routine snowpit seasonal survey (Landry et al., 2004). Our spacing and vertical distribution was chosen so that it represents the uniform character of the terrain, and allows analysis of both the suggested 5 m and 10 m distance spatial variability.

TEMPERATURE LAPSE RATES

Lapse rates were obtained by linear interpolation of daily maximum (T_{max}) and minimum temperatures (T_{min}) between the Base and Upper stations, for the period 9 January to 7 March for the years 1999 to 2004 (347 days). Because daily weather observation times differ between Base and Upper stations, the observation time was adjusted for both the T_{max} and T_{min} lapse rates (LR_{Tmax} and LR_{Tmin}). Daytime T_{max} occurrences vary with aspect and season (Barry, 1981), but since our Base station is on flat terrain and our Upper station has a minimal E-facing slope, 90% of our 2004 T_{max} occurred prior to 16:00. Therefore, T_{max} was taken from the record prior to 16:00 for the Base station (with 8:00 and 16:00 observations), and prior to 12:00 the next day for the Upper station (with once-daily 12:00 observations). Similarly, T_{min} is expected to occur in the early morning hours and hence prior to 8:00 for the Base station and prior to 12:00, on the same day, for the Upper station.

Bolstad et al. (1998) and Rolland (2003) found daily LR_{Tmin} more variable than daily LR_{Tmax}, as T_{max} is mainly affected by daytime solar radiation, while T_{min} also fluctuates with valley bottom cold air drainage. Cold air drainage, which is frequent during the winter months, is responsible for "inversion days," where temperature increases with elevation gain. This phenomenon suggests that separating inversion days from normal days would give a more accurate lapse rate approximation of a particular area. Hence, we separated our daily lapse rates into normal and inversion lapse rate days. We further divided the normal lapse rate category into dry and wet (precipitation) days to obtain approximate dry and wet adiabatic lapse rates for the area. A day was categorized as having a wet lapse rate in the event of any precipitation being recorded at either one or both the Base and Upper stations.

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Although some argue that linear regressions may not accurately represent regional lapse rates because factors such as topographic effects, local precipitation, and latent heat release causing local anomalies to be extrapolated over large areas (Bolstad et al., 1998; Shea et al., 2004), simple linear regressions remain an efficient method to assess temperature lapse rates with station temperatures adjusted for topographical differences (Rolland, 2003). For this study, the extrapolation error caused by cold air drainage in valley bottom was eliminated by categorizing lapse rates into normal and inversions days and with a further separation of dry and wet categories lapse rates we diminished additional linear extrapolation errors.

Rolland (2003) and Shea et al. (2004) argued that a minimum of 30 years of data is necessary to account for abnormal years and multiyear climatic events such as El Niño–Southern Oscillation (ENSO). However, since human development in the Westcastle valley is relatively recent (~1965) and still sparse, only a 6-year record was available for our study. In an attempt to alleviate errors caused by this short data set, winter lapse rates were compared with other relevant studies (Pielke and Mehring, 1977; Bolstad et al., 1998; Rolland, 2003; Shea et al., 2004; Thayyen et al., 2005). To assess whether the mean lapse rates of the 2004 study period differed from the mean lapse rates of 1999–2003, we performed non-parametric Wilcoxon's tests for LR_{Tmax} and LR_{Tmin} categories including and excluding inversion days for the 9 January to 7 March period in these years.

SNOWPACK TEMPERATURE GRADIENT

Snowpack temperature gradient (TG) was calculated from the 10 cm increment snowpack temperature data. A large (≥1 °C 10 cm^{-1}) TG indicates a faceting, weakening snow layer while a small (<1 °C 10 cm^{-1}) TG indicates a rounding, strengthening snow layer (McClung and Schaerer, 1993). The average TG was calculated for each profile, using all temperatures deeper than 20 cm in order to account for heat loss to the air at the top of the snowpack. In order to test "cold wave penetration" into the active layer, T_{min} of each profile was plotted against air T_{min} and T_{max} of the 4 nights and 4 days prior to each profile analysis. The best overall fit for the Base and Upper stations was accepted as the most prominent air temperature factor influencing snowpack TGs. Temperature at the time of observation (T_{pres}), T_{min}, and T_{max} were plotted against the daily 10 cm depth snow temperature ($Ts_{10\ cm}$) to assess heat loss at the top of the snowpack. In order to test air temperature persistence in the upper region of the snowpack, we also introduced an experimental 4-day continuous weather data time lag in both the T_{min} and T_{max} and associated this with $Ts_{10\ cm}$ through linear regression.

LAYERS AND SNOWPACK SETTLEMENT

Snow profile layers were interpreted weekly from each weather station and were recorded in Snowpro software and later plotted as stratigraphic columns. Correlation of layers from week to week was done using multiple criteria (stratigraphic principles, hardness, crystal shape, and boundary characteristics) and allowed calculation of weekly settlement rates for up to two weeks after the recorded snowfall event. Common layers, present at both weather plots, were compared and associated with their respective LR_{Tmax} and LR_{Tmin}. Overall snowpack settlement rate could not be determined since weekly profiles were dug ~30 cm behind the previous profiles and no correction for bottom topography was made, so no level datum could be established. Also, discontinuous solar and rain crusts as well as windblown layers further affected our ability to assess overall settlement rates.

Results

Castle Mountain Resort received an average of 469 ± 104 cm (Base) and 743 ± 242 cm (Upper) of snow per year in the last six winter seasons. In the 2003–2004 winter, the total snow accumulation, including that from wind drift, amounted to 347 cm at the Base and 944 cm at the Upper weather station. HS during the open seasons of 1999 to 2004 averaged 76 ± 24 cm (Base) and 234 ± 57 cm (Upper), and was 75 ± 19 cm (Base) and 211 ± 51 cm (Upper) in 2003–2004. Temperatures ranged from −38 °C to +15°C, with a mean of −5 ± 13°C during the open ski seasons of 1999 to 2004. The open ski season varied between 80 (2003) and 112 (2002) days, opening as early as 6 December and as late as 13 January, and usually closing in the first week of April.

TEMPERATURE LAPSE RATES

The 1999–2004 lapse rate data for 9 January to 7 March reveal an average LR_{Tmax} of −4.4 ± 5.7°C km^{-1} and LR_{Tmin} of −2.6 ± 6.8°C km^{-1}, including inversions, and an average normal LR_{Tmax} of −6.1 ± 3.2°C km^{-1} and LR_{Tmin} of −5.9 ± 3.4°C km^{-1}. Although the average lapse rates vary between the years as well as over the categories (Table 2), they are statistically not significantly different. Kolmogorov-Smirnov tests (p = 0.05) indicate that lapse rate observations for individual years as well as for the entire 6-year period are not normally distributed (α = 0.05), with or without inversion days, apart from year 2000 without inversions. Wilcoxon's tests reveal no significant difference (α = 0.05) between LR_{Tmax} and LR_{Tmin} of 2004 and the 5-year control including or excluding inversion days (LR_{Tmax}: χ^2 = 0.68, n = 347; and LR_{Tmin}: χ^2 = 0.08, n = 344. LR_{Tmax}: χ^2 = 0.64, n = 304; and LR_{Tmin}: χ^2 = 0.06, n = 252, respectively). Our 2004 year of observation is therefore representative of the general lapse rate distribution in the period since 1999.

The 2004 lapse rates (Table 2) including inversions show an average LR_{Tmax} of −4.1 ± 7.4°C km^{-1} and an average LR_{Tmin} of −1.1 ± 7.6°C km^{-1} (n = 59). However, normal lapse rates (excluding inversion days) show an overall larger LR_{Tmax} of −6.2 ± 2.9°C km^{-1} (n = 52); −6.1 ± 2.9°C km^{-1} for dry days and −6.4 ± 2.7°C km^{-1} for wet days. Normal LR_{Tmin} is −4.9 ± 2.5°C km^{-1} (n = 41), with similar values for dry and wet days. Inversion lapse rate averages are positive and extremely variable (LR_{Tmax} = 11.5 ± 11.8°C km^{-1}, n = 7; LR_{Tmin} = 7.7 ± 8.0°C km^{-1}, n = 18). From these data, it is evident that LR_{Tmax} and LR_{Tmin} for inversion days increase lapse rate variability and lower the overall average accuracy. Of the 6 years of lapse rate data (n = 347), inversion days occur in 26% of the LR_{Tmin} observations, while only in 11% of the LR_{Tmax} observations. This distribution is similar for individual years.

Contrary to the lapse rate averages, the modes for LR_{Tmax} and LR_{Tmin} are quite consistent for days including and excluding inversions in all years (Table 3). For the 6-year period, LR_{Tmax} and LR_{Tmin} modes are most often −6.3°C km^{-1}, and frequency histograms reveal that this value occurs about 20% of the time when taking inversions into account, but up to 25% of normal lapse rate days. In 2004, LR_{Tmax} modes are mostly −7.9°C km^{-1} and LR_{Tmin} modes generally −4.8°C km^{-1}. When comparing our 2004 lapse rate modes with the 5-year control, we can detect some consistencies in LR_{Tmax} and LR_{Tmin} between lapse rate categories, but some inconsistencies between years. The anomalously large

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TABLE 2

Study period (9 Jan to 7 Mar) average of daily maximum and minimum temperature lapse rates (LR_{Tmax} and LR_{Tmin}) for 1999–2004. Categories include overall, normal, dry, and wet normal lapse rates, and inversion lapse rates. Bold numbers represent the normal lapse rates (excluding inversion days) that we suggest are used in lapse rate extrapolation of local climate data. See explanation in text.

	LR_{Tmax} average						LR_{Tmin} average					
Year	*1999*	*2000*	*2001*	*2002*	*2003*	*2004*	*1999*	*2000*	*2001*	*2002*	*2003*	*2004*
All days	−6.8	−3.8	−4.5	−3.7	−3.2	−4.1	−5.7	−2.6	−0.5	−2.6	−1.9	−1.1
Excl. inversions	**−6.9**	**−6.2**	**−5.9**	**−5.8**	**−5.4**	**−6.2**	**−7.6**	**−6.3**	**−5.1**	**−5.3**	**−4.7**	**−4.9**
Wet	−6.7	−5.9	−6.1	−6.6	−6.0	−6.4	−7.9	−6.4	−5.3	−5.9	−5.9	−4.8
Dry	−7.5	−5.1	−6.0	−5.4	−5.3	−6.1	−7.0	−5.0	−5.2	−5.7	−3.6	−5.1
Inversions only	3.2	6.5	5.3	8.3	11.6	11.5	4.2	5.9	6.5	5.7	10.5	7.7

LR_{Tmax} and LR_{Tmin} modes in 1999 (−9.5°C km^{-1} and −7.9°C km^{-1}) might be due to 1999 having 36 days of wet LR_{Tmax} and 29 days of wet LR_{Tmin}, which is 9 to 18 more wet LR_{Tmax} and 2 to 11 more wet LR_{Tmin} days than in other years.

Although it is expected that dry lapse rates should be steeper than the wet, our data indicate that 5 out of 6 times the wet average LR_{Tmax} and LR_{Tmin} is higher than the dry measure, while this occurs 1 out of 6 times for the LR_{Tmax} mode and 3 out of 6 times for the LR_{Tmin} mode.

WIND CONDITIONS AND SNOW EVENTS

Weather observations from Castle Mountain Resort and Pincher Creek for the 9-week study period in 2004 show wind speed and direction similar to the dominant light western and southwestern winds in this region (Environment Canada, 2005). The wind data indicate a prevalence of southwestern winds for both the Base (88%) and Upper (69%) weather stations, followed by northern winds (8% versus 21%). Clearly, wind direction at the Base station is partly influenced by valley topography, funneling the winds into a southwestern direction (Fig. 1). Even with our nominal wind recordings we observe a wind lapse rate where wind speed increases with elevation, which is consistent with evidence showing increased snow transport at high elevations in our study area and from literature (CAA, 2002; Erickson et al., 2005).

Castle Mountain Resort was affected by five snowstorm cycles and two major wind events during the 9-week study period. The first and major storm cycle was associated with a northern cold front and brought almost twice as much new snow to the Upper station than to the Base (Fig. 2). The last and second largest storm cycle occurred in March and was also associated with a cold front from the north. It brought six times as much snow to the Upper weather station than to the Base. The three other storms occurred between 11 and 26 February, and were associated with southwestern systems and temperature inversions bringing a cumulative total of 63 cm and 16 cm of new snow at the Upper and Base stations, respectively. Additionally, six snowdrift events were recorded at the Upper station during which no new snow was recorded at the Base. The first two wind drift events, in January, were related to southwestern winds while the 1 February event had northern winds and occurred during the first snowstorm cycle. The two late February snowdrift events resulted from southwestern winds while the 1 March event was attributed to the strong northern winds preceding the last snowstorm cycle. In total, 275 cm of new snow fell at the Upper and 126 cm at the Base station, while 23 cm of wind-drifted snow was accumulated at the Upper station.

STRATIGRAPHY

In the 2003–2004 winter season, 15 snow layers that formed at the Base and 11 that formed at the Upper station can be traced to specific events. Seven of the Base station's layers, including two rain crusts, developed prior to 9 January (Fig. 3a). The first rain crust is from a 19 November rain event percolating down the 17 November snow layers, forming a 23 cm thick ice/facet layer (visible in the 16 January profile: Fig. 3a). The rain crust is discontinuous, and can only be clearly seen from the 7 February profile onwards, when a distinct two-boundary layer developed above the 17 November snowpack. A second rain crust developed from rain on 3 December followed by 2 cm of snow on 5 December and another rain event on 6 December. The other four layers have no crusts. An additional eight layers accumulated after 9 January of which four are rain and/or melt crusts (15 and 30 January, 27 February, and 5 March). The remaining layers are from snow/drift accumulation events only. At the Upper station four layers developed prior to 9 January (Fig. 3b). Most of these, and subsequent eight layers, were affected by moderate to strong winds, based on our daily nominal wind speed and snow drift observations and extrapolation of wind measurements at the Pincher Creek weather station (Environment Canada, 2005). A total of nine common layers were found between the Base and

TABLE 3

Study period (9 Jan to 7 Mar) mode of daily maximum and minimum temperature lapse rates (LR_{Tmax} and LR_{Tmin}) for 1999–2004. Categories include overall, normal, dry, and wet normal lapse rates, and inversion lapse rates. Bold numbers represent the normal lapse rates (excluding inversion days) that we suggest are used in lapse rate extrapolation of local climate data. See explanation in text.

	LR_{Tmax} mode						LR_{Tmin} mode					
Year	*1999*	*2000*	*2001*	*2002*	*2003*	*2004*	*1999*	*2000*	*2001*	*2002*	*2003*	*2004*
All days	−9.5	−6.3	−7.9	−6.3	−6.3	−7.9	−9.5	−6.3	−7.9	−6.3	−6.3	−4.8
Excl. inversions	**−9.5**	**−6.3**	**−7.9**	**−6.3**	**−6.3**	**−7.9**	**−9.5**	**−6.3**	**−7.9**	**−6.3**	**−6.3**	**−4.8**
Wet	−9.5	−6.3	−7.9	−7.9	−6.3	−6.3	−9.5	−6.3	−6.3	−6.3	−3.2	−4.8
Dry	−9.5	−6.3	−9.5	−6.3	−6.3	−7.9	−7.9	−3.2	−6.3	−3.2	−6.3	−6.3
Inversions only	3.2	1.6	3.2	3.2	4.8	3.2	1.6	1.6	1.6	1.6	6.3	1.6

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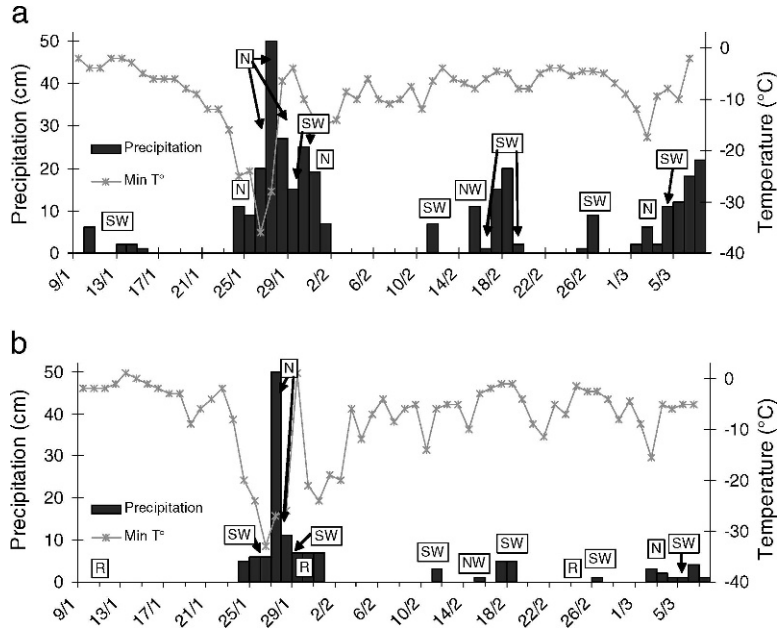

FIGURE 2. Study period storm cycles for Upper (a) and Base (b) weather stations with daily solid precipitation, wind direction, and minimum air temperature. R indicates a rain event.

Upper weather plots (17 and 29 November; 17 December; 8, 15, and 30 January; 11 and 19 February; and 7 March).

SNOWPACK SETTLEMENT

We calculated settlement rates (SR = compaction of individual layers), in percentage, for the first and second week after each snowfall event for all traceable layers in the Base and Upper plots. SR ranges from 0 to 100% (melt/rain crusts disappearing), with a majority in the 45–70% range. This variability can be related to the duration, amount, and type of snowfall per event, as well as number of days between snowfall and profile interpretation, temperature, and wind conditions during and after the snow event, and any snow and rain following it (Gray and Morland, 1995; Marshall et al., 1999). However, our method of profiling from week to week by digging 30 cm behind the previous pit introduces an additional layer thickness variability, which can be up to one-fourth of the individual layer thickness. This suggests that SR $< 25\%$ cannot be interpreted as absolute settlement. However, since all our thickness measurements, including where weekly changes are $<25\%$, show that thinning is progressive from week 1 to week 2, the majority of the observed thinning is considered to be due to settlement (compaction and metamorphosis) rather than through spatial snowpack thickness variation, which should be random at the small spatial scale of the individual plots.

Of the nine common layers in the two weather plots, six can be reasonably correlated in terms of settlement, as the period between the pre-17 December snowfall and our first snow profile recording is too long to distinguish individual events. Further, only 4 layers of the Base, and 2 layers of the Upper snowpack could be traced for settlement up to two weeks after snowfall. SR is greatest during the first week, after which an additional 5–30% settlement was recorded. In Figure 4 initial layer thickness is plotted against percentage settlement for one week after a snowfall event. From these data, a number of conclusions about snow settlement at Castle Mountain Resort can be derived. Firstly, SR is generally greater at the Base weather plot. This could partly be due to a combination of warmer temperatures at lower elevations and increased wind loading at higher elevations, both of which are lapse rate dependent. Secondly, settlement is related to initial snow thickness. The only three exceptions are the two Upper plot layers of 53 and 73 cm with anomalously low SR ($<4\%$), and the 157 cm Upper plot layer with SR $< 40\%$. These layers accumulated during strong wind drift conditions and in a multi-day snow event. Generally, there is a weak significant linear correlation between the SR of the Base and the Upper plots ($R^2 = 0.52$), where most discrepancies can be attributed to the temperature and wind lapse rate related differences between the two plots.

Single day temperature lapse rates cannot easily be associated with individual layers, since snow events generally last several days and profiles are only interpreted once a week. However, when correlating LR_{Tmax} of the last day of snowfall with SR in the first week (Fig. 5), a strong negative correlation transpires (Upper $R^2 = 0.69$, Base $R^2 = 0.77$, $n = 5$), where days with temperature inversions (positive lapse rates) have low SR ($<25\%$), hence within the range of the spatial snowpack variability, and where days with large normal lapse rates have increasingly higher SR. More SR observations in relation to lapse rates are needed to verify these findings. It could be argued that inversions usually occur during cold, stable atmospheric conditions, and it would therefore be the colder temperatures that reduce settlement. Nevertheless, strong winds occasionally occur at the Upper plot during inversions. By pre-compacting the snow, these conditions, similarly, have a settlement-reducing effect.

SNOWPACK TEMPERATURE GRADIENT

For the Base station, snowpack temperature gradients (TG) are negative (warmer near the top) in the top 10–20 cm, except for the first and last weeks during which negative TG was up to 30 cm below the surface. This top "active layer" is affected by diurnal temperature variations, and the decrease in snowpack temperature is due to loss of heat to the overlying cold air, which becomes more pronounced during prolonged cold periods (McClung and Schaerer, 1993). Below this active layer, the average TG is relatively small and quite consistent amongst all profiles (0.46 ± 0.21°C 10 cm^{-1}). All TG $\geq$ 1°C 10 cm^{-1} occurred near the snowpack surface except for three minor instances in week 5 (1.1°C 10 cm^{-1} at 74–64 cm and 44–54 cm below the surface and 1.2°C 10 cm^{-1} near the ground) and one in week 6 (1.4°C 10 cm^{-1} near the ground). Since facets and depth hoar form in snowpack

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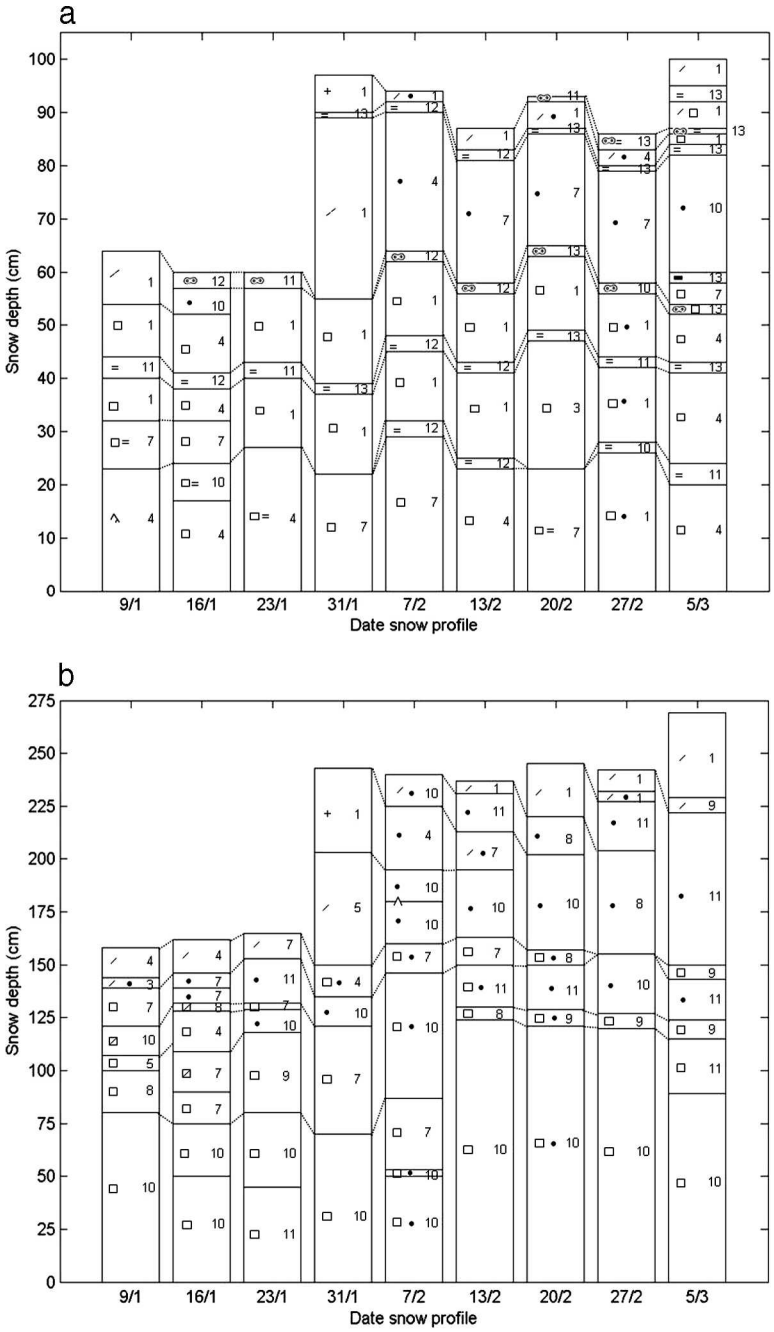

FIGURE 3. Snow profile stratigraphy for Base (a) and Upper (b) weather stations for 9 weekly snow pits. Symbols and numbers in each layer depict crystal form and layer resistance (CAA, 2002), respectively. Layer resistance is on a scale 1–14, with 1 = Fist and 14 = Ice. Profile depth is shown from ground.

zones where TG are $\geq 1°C\ 10\ cm^{-1}$ (Birkeland et al., 1998; Pfeffer and Mrugala, 2002), our region regularly depicts conditions appropriate for near-surface facet development. However, faceted snow layers are found throughout the Base profile (Fig. 3a). Due to the nature of our TG data, we cannot accurately assess whether the effects of diurnal cycles and solar radiation induced large TG and near-surface facets within the snowpack. An alternative explanation for the common presence of facets in our Base snowpack is the fact that faceted crystals are often observed above and below crusts (Colbeck and Jamieson, 2001). Since the Base profiles have between 1 and 7 crusts, we can assume that at least some of the faceted layers within our profile interpretation could be a result of latent heat release and large vapor pressure gradients between crusts and adjacent snow layers.

For the Upper station, the negative TG generally occurs within the top 10–40 cm but weeks 2 and 4 are entirely positive. When removing the active layer, the average TG at the Upper station is also relatively small ($0.30 \pm 0.12°C\ 10\ cm^{-1}$), and not statistically different from the Base profile TGs. These smaller TGs could be consistent with the greater occurrence of rounded type snow crystals observed in the Upper station's profiles as well as with the absence of crusts (Fig. 3b).

TEN CENTIMETER SNOWPACK TEMPERATURE

Linear regressions of daily $Ts_{10\ cm}$ snowpack data and daily air T_{min}, T_{max}, and T_{pres} for the Base and Upper stations yield the strongest positive correlations when associating $Ts_{10\ cm}$ with T_{pres}. Base T_{min}, T_{max}, and T_{pres} results (Fig. 6b) yield R^2 of 0.65, 0.64, and 0.69 while the Upper weather station (Fig. 6a) has R^2 of 0.47, 0.35, and 0.67, respectively. Introducing a continuous four day time lag did not produce any significant correlations. This suggests that heat loss or gain in the top 10 cm of the snowpack at Castle Mountain Resort is within hours, which could be an effect

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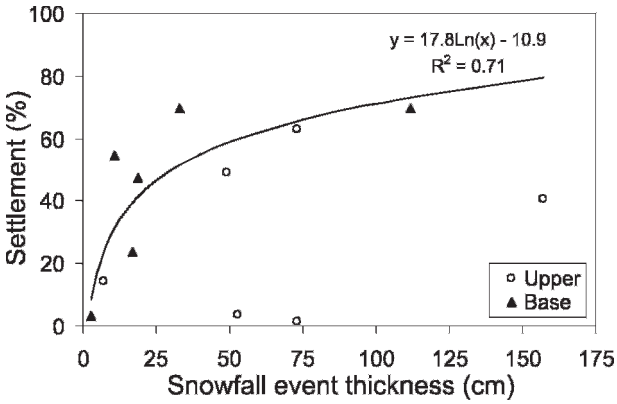

FIGURE 4. Initial snowfall event thickness related to Base and Upper plot settlement rates for the first week after snowfall. Logarithmic trend line fitted through the Base plot data as well as three of the Upper plot data points ($n = 9$). The three Upper plot outliers with minimal settlement are explained in the text.

of the relatively strong winds, enhancing the turbulent flux at this site.

COLD WAVE PENETRATION

Weekly snow minimum temperatures (Ts_{min}) of the Base and Upper profiles varied from −6.5 to −1.0°C and −10.5 to −5.5°C, respectively. All Ts_{min} were found within 40 cm of the snow surface at the Base station and within 50 cm at the Upper station, except for 31 January during which the Upper Ts_{min} was 80 cm below the surface. This deep Ts_{min} can be associated with a prolonged cold spell followed by a rapid warming (see Fig. 6a). No correlation can be found between Ts_{min} depth and any of the four previous night's air T_{min} for the Base plot. However, a correlation was found between Upper snowpack Ts_{min} depth and T_{min} of four nights previous ($R^2 = 0.78$, $n = 9$) and T_{range} ($R^2 = 0.84$, $n = 9$). This is most likely the result of the prolonged late January cold spell.

In order to investigate the influence and time lag of heat loss at the surface of the snowpack as a result of cold surface air, Ts_{min} were correlated to T_{min} of the four previous nights, T_{max} of the previous 4 afternoons, and T_{range} of the previous 24 hours. Of the 9 weekly snow profiles at the Base station, 9 and 31 January showed little correlation with the previous night T_{min} due to the occurrence of low T_{min} and large T_{range} within several nights of profile interpretation. The 5 March Ts_{min} showed no correlation with any of the four previous night's T_{min}. This profile, as well as that of 31 January, was recorded at 14:00, which generally corresponds to T_{max} occurrences. All other profiles were interpreted before 12:30. Linear regression of T_{min} and Ts_{min} shows a weak positive correlation ($R^2 = 0.44$, $n = 9$), but reveals a stronger positive correlation ($R^2 = 0.88$, $n = 6$) when the 9 January, 31 January, and 5 March profiles are removed.

The Upper station snowpack T_{min} is also affected by the previous night's T_{min}, in relation to the previous afternoon T_{max} range. Within the 9 snow profiles of the study period, 2 (9 January and 13 February) show little correlation with the previous night T_{min}. These differences are also due to the occurrence of low T_{min} and large T_{range}. Persistent cold T_{min} can be observed in the snowpack for several days after the air T_{min} has warmed up. The 9 January snowpack showed Ts_{min} as low as −9°C, even with the previous two nights T_{min} being warm (−2°C and −4°C). In this case, three to seven nights pre-profile had extremely low T_{min} (see

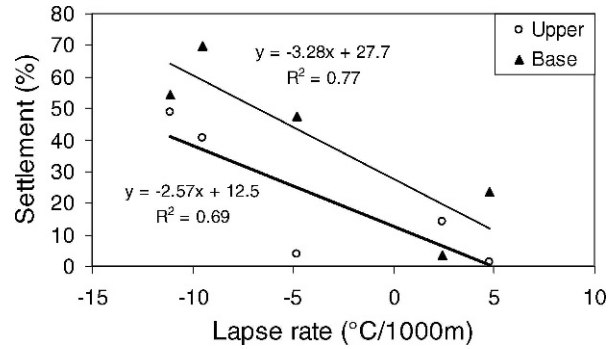

FIGURE 5. Maximum temperature lapse rate (LR_{max}) related to settlement rate in the first week for Base and Upper stations. Linear trend lines fitted for Base (thin) and Upper (thick) station data.

Fig. 6a). As for 13 February, a similar colder T_{min} trend can be observed several days prior to snowpack interpretation. Linear regression of previous night air T_{min} and Ts_{min}, including all 9 weeks, shows no correlation ($R^2 = 0.00$, $n = 9$) but reveals a strong positive correlation ($R^2 = 0.75$, $n = 7$) when the 9 January and 13 February profiles are omitted. For the Upper station profiles, no correlation can be observed between Ts_{min} and pre-profile T_{min}, not even when 9 January and 13 February are removed.

Hence, it appears that cold wave penetration time lag varies with temperature and T_{range}. For profiles interpreted during small T_{range} (≤6°C) periods, Ts_{min} is correlated with T_{min} of the previous 24 hours, but not with T_{min} 2 to 4 nights pre-profile. For profiles interpreted during large T_{range} (≥18°C) periods, a strong discordance can be observed between Ts_{min}, T_{min}, and T_{max}. Intermediate T_{range} periods give very weak or no correlation with T_{min} of 1–4 days previous. No correlation was found between Ts_{min} and any of the 4 preceding afternoon's T_{max}. These results emphasize the importance of T_{min} and prolonged cold spells on the winter snowpack TG.

SNOWPACK SPATIAL VARIABILITY

The seven Candy Cane snow profiles recorded on 25 January vary in snowpack thickness (HS) as well as number and thickness of individual layers, but the main accumulation layers can be easily correlated. HS varied from 241 to 195 cm with an average of 218 ± 18 cm. Between 5 and 8 major layers were identified in each profile (Fig. 7). The snowpack TG, from the temperature inflection point (below the top "active layer," which is affected by diurnal temperature variations and where temperature decreases with depth), were statistically similar for all profiles (0.44 ± 0.14, 0.37 ± 0.11, 0.34 ± 0.10, 0.56 ± 0.15, 0.55 ± 0.16, 0.63 ± 0.18, 0.47 ± 0.15 °C 10 cm^{-1}) with no TG > 1°C 10 cm^{-1} below 80 cm from the surface. Seven layers in the Candy Cane profiles could be correlated with snowfall and snow drift events, while six of these could be cross correlated with layers in the 23 and 31 January Upper and/or Base weather plot profiles (Fig. 8). Generally, the Candy Cane snowfall amount per event is of the same order as that at the Upper station and 2–4 times more than at the Base, which is not surprising as Candy Cane is within 10–30 m elevation and within 400 m distance of the Upper station. At the Base, 89 cm of snow was recorded by 17 November and 65 cm between 18 and 29 November, but snow recording had not yet started for the Upper plot because the ski hill did not open until 9 December. These two early season events are reflected as 25 and 14 cm thick layers in the late January Base profiles, and as

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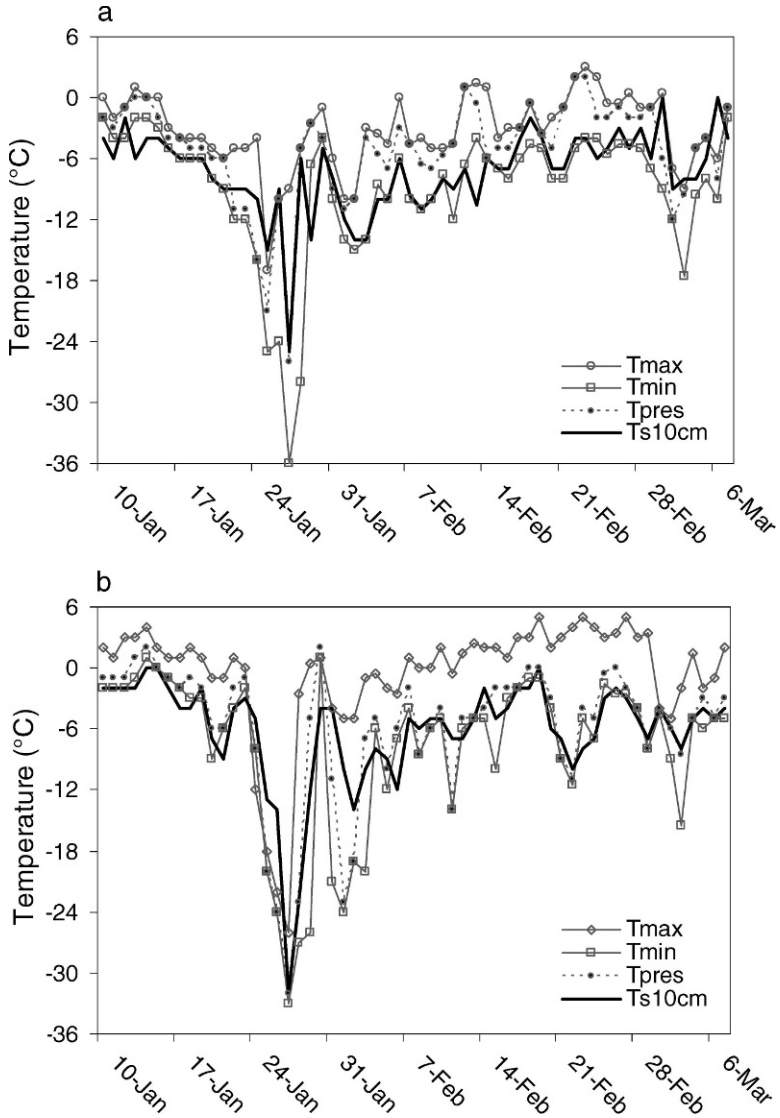

FIGURE 6. Upper (a) and Base (b) weather station daily snow temperatures 10 cm below the surface ($Ts_{10\ cm}$) and associated daily minimum (T_{min}) and maximum (T_{max}) air temperatures, and air temperature at time of snow profile observation (T_{pres}).

approximately three times thicker correlating layers in the Upper and Candy Cane profiles (Fig. 8). On 15 January snow accumulated at Candy Cane and the Upper station, but only a rain and melt crust had developed at the Base. This crust is undetectable in the 31 January Base layer, but is consistent in all other stratigraphic profiles (Fig. 3a). Conversely, Candy Cane appears to be more affected by snowdrift accumulation than the Upper station site, resulting in 5–10 cm more snow on 17 December, 8 and 18 January at Candy Cane. The variability of layer thickness between the Candy Cane plots (standard deviations of up to 25%; Fig. 8), as well as that between the 23 and 31 January layers in the Upper and Base plots, again, give rise to caution in the interpretation of settlement rates that are less than 25% of the original snow thickness.

SNOWPACK STABILITY FROM COMPRESSION TESTS

For the 9-week study period, there were 18 compression test failure results at the Upper station (0–5 each week) and 26 at the Base station (2–4 each week). The seven Candy Cane snow pits recorded on one day displayed 14 failures (0–4 per snow pit), of which the most prominent medium force failure occurred at the same layer interface between 30 and 41 cm from the surface in 5 of the 7 pits. A failure of similar force and type occurs at the same depth in the 31 January profile of the Upper station. The difference of ease of failure between the Base and Upper stations is remarkable, with distributions of low force (1–10 light taps), medium force (11–20 medium taps), and high force (21–30 hard taps) for Base (31, 61, 8%) and Upper (5, 39, 56%), respectively, while Candy Cane (21, 50, 29%) is intermediate. This suggests that the Base snowpack is less stable than the Upper station's, as 92% of failures at the Base required low to medium force, while just 44% of the Upper did so. However, the depth of the failure plane is also critical for the eventual avalanche danger (Chalmers and Jamieson, 2001) and 7 out of 8 Base low force failures occurred within 15 cm of the top of the snowpack, while only one occurred at 40 cm. All of these Base station failures occurred at the transition between rain crusts and regular snow layers. The only low force failure at the Upper station similarly occurred within 15 cm of the top, but none of its low or medium force failures were on rain crusts. The depth distribution of medium force failures for Base and Upper is also similar and ranges between 11 and 75 cm. The only ground level failure occurred at the Base station with a total snow depth of only 60 cm; neither a steep temperature gradient nor a depth hoar layer were detected here. The Candy Cane low force failures occur slightly deeper, with 3 low force failures at 20–33 cm from the top, and 6 medium force failures between 18 and 60 cm. All high force failures at the Base, Upper,

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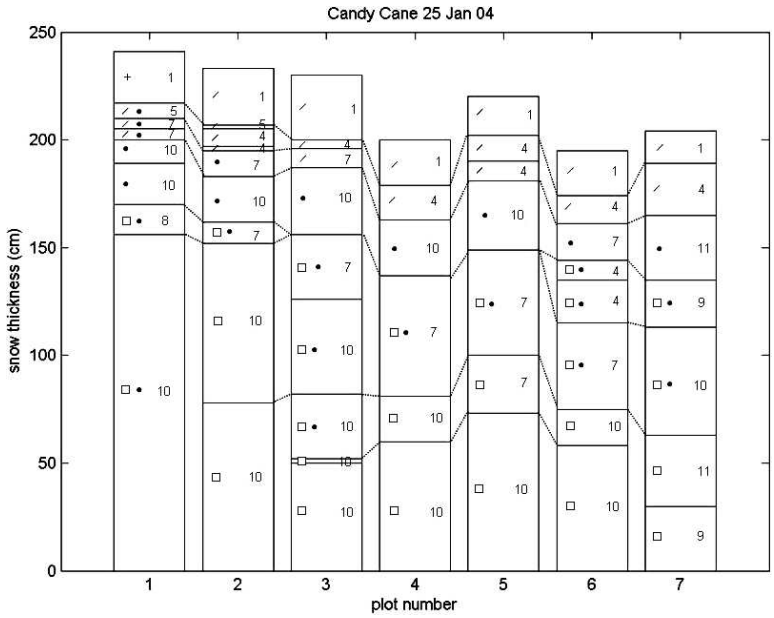

FIGURE 7. Snow profile stratigraphy for seven Candy Cane snow profiles taken on 25 January 2004. Stratigraphy symbols include crystal form and layer resistance (CAA, 2002) for individual layers. Layer resistance is on a scale 1–14, with 1 = Fist and 14 = Ice. Profile depth is shown from ground.

and Candy Cane sites occurred between 30 and 110 cm. Hence, overall, a higher force is needed to fail layers lower in the snowpack, which is visualized by plotting failure force versus failure layer depth below the surface (Fig. 9). This clearly shows an exponential decline of ease of failure with depth (R^2 for the 3 sites is in the range of 0.56–0.66, while R^2 is 0.59 for the entire data set, $n = 58$).

We tested for the correlation between force needed for failure and (1) layer hardness (McClung and Schaerer, 1993), (2) hardness and (3) grain size difference across the failure interface (Schweizer and Jamieson, 2002), (4) temperature gradient (Pfeffer and Mrugala, 2002), and (5) age of the snow layer (Landry et al., 2004). Of these, only age of snow layer (in weeks since snowfall) had a significant positive correlation with force, where older snow layers need higher force and are thus less likely to fail (Fig. 10). However, when crusts are failure horizons, failure force does not decrease with age. We tested both linear and logarithmic correlations between force and layer age, giving $0.43 < R^2 < 0.58$, and found that there was only a marginal and non-systematic difference between the two.

The difference in slope angle between the Base (~0°), Upper (14°), and Candy Cane (31°) locations could have had a confounding effect on our stability test results. However, (1) elevation is the primary factor determining rain and sun crust occurrence (though rain crusts might be thinner on steeper and/or windward slopes: Jamieson, 2004), (2) no systematic decrease in stability was found with increasing slope, and (3) stability results in the Upper and Candy Cane profiles in the same week returned similar failures at similar depth. We therefore conclude that our compression test results were primarily related to elevation

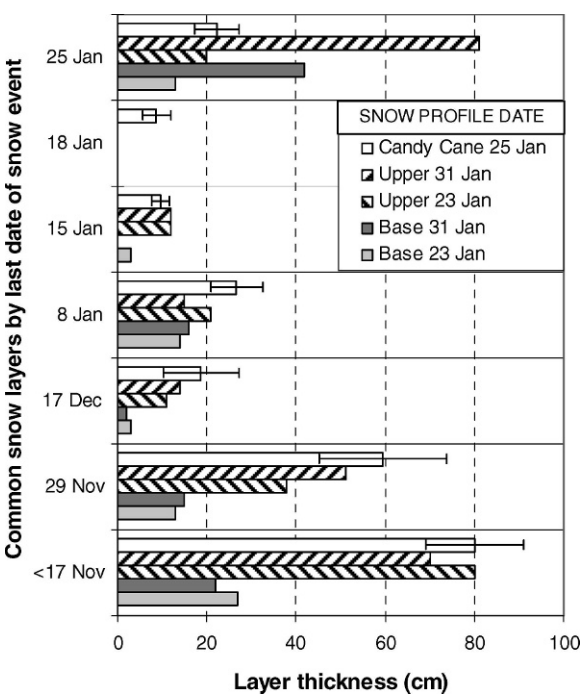

FIGURE 8. Layer thickness of seven common snowfall and/or snowdrift events at the Base (gray bars) and Upper (hatched bars) stations observed on 23 and 31 January, and mean and standard deviation of seven Candy Cane plots (white bars) observed on 25 January. The "25 January" Upper and Base layer thickness is based on snowfall recorded on 24 and 25 January, and the 31 January layer thickness is measured in the 31 January snow profile. Note the additional accumulation between 25 and 31 January.

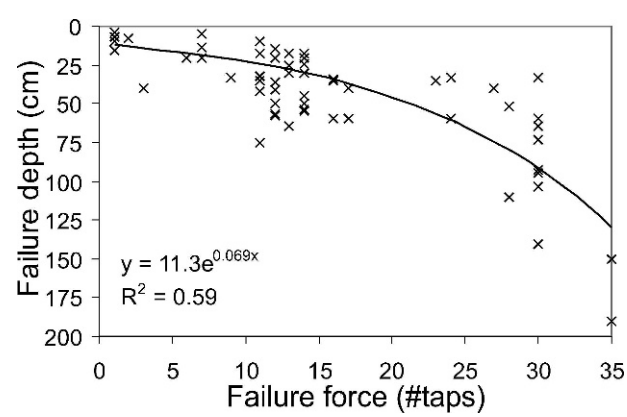

FIGURE 9. Failure force versus failure depth of all failures at the Base, Upper, and Candy Cane snow profiles. An exponential trend line is fitted through entire data set ($n = 58$).

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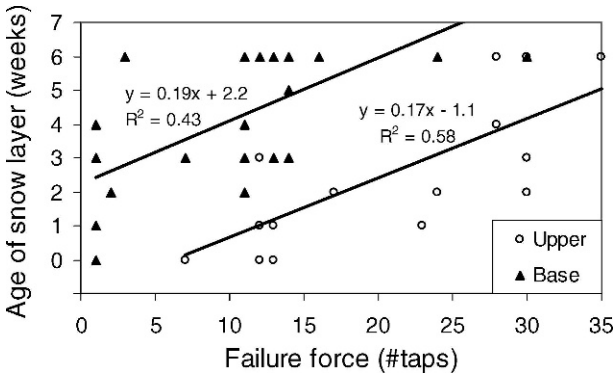

FIGURE 10. Failure force versus age of snow layer for all failures occurring in the 9 weeks of snow profile stability tests at the Base and Upper stations. Linear trend lines are fitted. All failure forces <5 taps occurred on rain crusts at the Base station.

differences, but acknowledge that slope might have had a minor influence.

Discussion

TEMPERATURE LAPSE RATES

Pielke and Mehring (1977) were one of the first to quantify the importance of elevation and terrain in climatological data. For their 1958–1973 monthly mean lapse rate observations in Virginia, they concluded that linear regressions were appropriate to obtain temperature lapse rates but identified limitations affecting their use in regions with frequent inversions. Pielke and Mehring's (1977) January, February, and March lapse rates averaged −5.6, −6.0, and −6.4°C km^{-1}. These values closely resemble our normal January–March −6.2°C km^{-1} LR_{Tmax} but do not fit the −4.1°C km^{-1} LR_{Tmax} including inversion days. By separating inversion from non-inversion days, the use of linear lapse rates in mountainous regions seems to yield more appropriate results. Conversely, Thayyen et al. (2005) found non-linear average monthly lapse rates among three high- and low-elevation weather station pairs in the Garhwal Himalaya, India, where low and high pairs gave markedly different lapse rates in all seasons. These non-linear lapse rates were attributed to the altitudinal differences in snow cover. Overall lapse rates averaged −5.9°C km^{-1} for the study period (May–November 1998–2000) when using the lowest (2540 m) and highest (3763 m) stations as a pair, but during the monsoon months (July–August) the average lapse rate was 1–3°C km^{-1} lower. Thayyen et al. (2005) observed inversion days only during October and November, suggesting the importance of inversions during the winter months rather than the summer, and emphasized the similarity between Din Gad catchment's non-inversion lapse rates and the normal environmental lapse rate of −6.5°C km^{-1}.

Geographically closest to our study region, Shea et al. (2004) used linear regression analysis of monthly average temperature lapse rates from 1961 to 1990 in the Columbia Mountains, British Columbia, and grouped their data into four seasonal bins. Their average winter (November–February) lapse rate including inversion was −4.9°C km^{-1}, and similar to our LR_{Tmax} including inversions (−4.1°C km^{-1}). In contrast, their average March–May, and June–August lapse rates were −6.0°C km^{-1} and −5.3°C km^{-1} respectively, while September–October was −4.6°C km^{-1}. When eliminating inversion days, our winter LR_{Tmax} of −6.2°C km^{-1} is closer to Shea et al.'s (2004) spring and summer lapse rates. Rolland (2003), who did not exclude inversion days, also found a strong seasonal lapse rate pattern in the European Alps, with consistently higher summer values. These findings suggest, again, that inversions affect winter lapse rates considerably and should be taken into account in snow hazard forecasts.

Our results further show that wet LR_T is often higher than dry LR_T, albeit marginally. This, at first instance, seems counterintuitive. However, saturated adiabatic lapse rate varies with temperature and elevation (our elevation range is small enough to assume no steepening with elevation) and increases for our elevation range from about 4°C km^{-1} for high temperatures to close to 10°C km^{-1} for temperatures below about −15°C (Stull, 2000). The larger (steeper) our yearly average wet LR_{Tmax} and wet LR_{Tmin}, the larger the ratio of cold days (below −15°C) versus warmer days with precipitation in that year (e.g. 1999: Table 2). Other research (e.g. Bolstad et al., 1998) further shows that valley to ridge stations lapse rates (−6.5 to −7.0°C km^{-1}) can be significantly smaller than their higher elevation side slopes to ridge station lapse rates (−8.0 to −9.0°C km^{-1}), which could also be a reflection of this temperature dependence of the saturated adiabatic lapse rate. However, the fact that dry LR_T for all our years was much smaller than the dry adiabatic lapse rate (10°C km^{-1}) suggests either that our measurements are affected by radiation from the ground, or that even on dry days the lifting condensation level (Stull, 2000) is reached, and thus, that days that we categorize as "dry" are more a reflection of the true environmental lapse rates in the region.

Bolstad et al. (1998) tested the accuracy of temperature lapse rates generated from 10 years of data using regional regression models, kriging, and local models from 13 local and 35 regional National Climatic Data Center (NCDC) stations in the southern Appalachians. They found regional regression models to yield more accurate estimates of station temperatures when using the NCDC stations but found no significant differences in daily temperature predictions between regional regressions and local lapse models when compared to an independent data set. Bolstad et al.'s (1998) monthly average LR_{Tmin} ranged from −3.8 to −5.8°C km^{-1} and was consistently smaller than LR_{Tmax}, ranging from −4.0 to −10.0°C km^{-1}, regardless of the method used. The smaller LR_{Tmin} values were attributed to cold air drainage in valleys, which occurred in over half their study period. However, there were some local and regional differences; January to February LR_{Tmin} ranged from −2.0 to −1.0°C km^{-1} for valley to ridge stations, −2.0 to −2.5°C km^{-1} for side slopes to ridge stations, and −3.8 to −4.0°C km^{-1} for their regional average. This may be a reflection of local extrapolations which could be emphasizing local inversion occurrences that are not accentuated in regional lapse rate values. Yet again, it seems that removing inversion days from local data gives more appropriate local lapse rate values which may or may not be well represented by regional lapse rates, depending on the area's topographic character. However, our 2004 LR_{Tmin} values obtained from a valley and slope station also showed much lower values (−1.1°C km^{-1} including, and −4.9°C km^{-1} excluding inversions) than LR_{Tmax} (−4.1°C km^{-1} including, and −6.2 excluding inversions). Indeed, most years show a significantly higher LR_{Tmax} than LR_{Tmin} with or without inversions, but more so for the averages than for the modes. This lower lapse rate for temperature minima than for maxima is in agreement with others who researched lapse rates at the small scale (Lookingbill and Urban, 2003) as well as at larger scale (Thornton et al., 1997), and most studies still show lower LR_{Tmin} than LR_{Tmax} values, even when removing inversions. Additionally, Dodson and Marks (1997) found LR_{Tmax} to be more spatially stable than LR_{Tmin} because of cold air drainage in the latter. This leads us to believe that lower LR_{Tmin} values can also be

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partly attributed to other phenomena, such as the effect of daytime heating on LR_{Tmax}. Moreover, this lapse rate phenomenon suggests that snowpack properties are affected more dissimilarly during the day than at night, and further, that nighttime snow fall might result in more similar snowpack characteristics for high and low elevation sites than daytime snow fall. However, Rolland (2003) found that LR_{Tmin} and LR_{Tmax} values were similar in an analysis of 640 climate stations in the Italian and Austrian Alps. This discrepancy may be due to the regional characteristic of Rolland's (2003) study area, yet may also reflect a more fundamental difference in physical environment and climate patterns (i.e. different continents, latitude). Nevertheless, Rolland's (2003) January to February LR_{Tmin} (-4.0 to -5.6°C km^{-1}) and LR_{Tmax} (-4.0 to 5.1°C km^{-1}) values are consistent with our average LR_{Tmin}, and might be a little less steep than our lapse rates because of the milder winter temperatures and more humid conditions in the more maritime climate of the European Alps.

Lapse rates are generally found to decrease with latitude, partly through latitudinal dependence of temperature at sea level and related moisture content. Comparing non-inversion lapse rates from the above studies (where separation of inversion days was not possible, then summer lapse rates were used as non-inversion rate proxies), no clear latitudinal trend transpires between 31° and 56°N. Ranges in lapse rates from south to north are: -5.9 to -6.5°C km^{-1} (30°50′N: Thayyen et al., 2005); -4 to -7°C km^{-1} (35°–36°30′N: Bolstad et al., 1998); -5.6 to -6.4°C km^{-1} (36°–40°N; Pielke and Mehring, 1977); -6.3 to -6.6°C km^{-1} (43°50′N: Rolland, 2003); -5.4 to -6.9°C km^{-1} (49°19′N: this study); -5.3 to -6.0°C km^{-1} (49°–56°N: Shea et al., 2004). These overlapping ranges and the fact that latitudinal lapse rate differences including inversion days are found to be seasonally dependent and larger in winter (Rolland, 2003), when more inversions occur, suggest that—for this latitude range—differences in lapse rates could partly result from confounding latitudinal/terrain dependent factors influencing frequency of inversion days. Hence, if inversion days are removed, then lapse rates appear independent of latitude, as previously demonstrated by Moore (1956).

SNOWPACK PROPERTIES VERSUS LAPSE RATES

Total snowfall per day, or per storm cycle, was always larger at higher elevations, but no significant differences in relative amounts were found with different lapse rate conditions. This could be an artifact of our sampling scheme, which only allowed weekly observation of snowfall, and whereby multiple day layers nor wind loading effects could be separated from single day lapse rates. Nevertheless, we found a significant correlation between LR averages as well as modes, and several other snowpack properties. Settlement rate for a week after snowfall appears directly correlated to temperature lapse rate, whereby large normal lapse rates resulted in higher settlement rates, and inverse lapse rates resulted in minimal settlement. This suggests that snowpack properties change faster with steeper lapse rates, and do not change significantly during inversions. Inversions generally occur during clear skies (Barry, 1981) and hence with generally colder conditions, and this therefore suggests that colder air temperatures do not rapidly change the properties of the snowpack. Generally, cold temperatures are found to harden and strengthen a snow layer (Jamieson and Johnston, 1999), unless a large TG promotes faceting, while sudden warmer temperatures soften the layer and are more often associated with avalanches (McClung and Schaerer, 1993). However, Kozak et al. (2003) suggested that while higher temperatures decrease snow hardness and hence resistance to failure in the short term, in the long term, higher temperatures increase snow hardness.

Rain crusts only occurred at the Base station, and boundaries between these and adjacent snow layers are locations of faceted crystal and depth hoar growth and were found to be the most likely low force failure planes. This suggests that detailed knowledge of local lapse rates might help predict the highest elevation for rain occurrence (freezing level) as well as air temperatures influencing snow metamorphosis. Since higher elevations subsequently have larger snowfall amounts, rain crusts will generally occur at greater depths below the surface and are thus likely to increase the frequency as well as the magnitude of slab avalanches at higher elevations. Also, existing rain crusts (formed in early winter) will progressively occur deeper in the snow profile, potentially increasing the avalanche hazard throughout the winter season. November rain events in the Columbia Mountains have resulted in increasing occurrence of rain crusts since 1995, and some of these formed the base of slab avalanches (Jamieson, 2004). Therefore, increased knowledge of small scale effects on lapse rates and accurate detection of freezing levels might help increase the accuracy of avalanche danger forecasting.

Cold wave penetration analysis indicated that previous night's T_{min} is a good indication of snowpack temperatures unless persistent cold air temperatures occurred several days previous. Very cold temperatures cause large TG, destabilizing the snowpack by the formation of faceted layers (McClung and Schaerer, 1993). T_{max} does not have a significant influence on Ts_{min}. Therefore, if LR_{Tmin} are extrapolated for a region, it should be taken into account that about 25% of the night temperatures have inverse lapse rates.

SNOWPACK SPATIAL VARIATION

Variability in snow depth is a result of the interaction of local and regional weather (wind, temperature, snowfall rate, solar radiation), during and after deposition, with terrain heterogeneity (slope, aspect, substrate, vegetation) (Sturm and Benson, 2004). Erickson et al. (2005) suggested that a large portion of the variability can be attributed to rough topography and related wind redistribution in areas above the tree line, and that an index of wind sheltering has the greatest effect on snow depth. However, even in relatively flat terrain such as the Arctic, wind variation appears to have the greatest effect on distribution of snow within the 10–20 m scale, where it is related to the distribution of vegetation and snow dunes (Sturm and Benson, 2004). Comparison of individual snow layers as well as of total snow depth between the seven snow pits at Candy Cane showed that spatial variability is generally 25% of layer thickness as well as of total snowpack thickness. Since compaction rates are often less than 25%, one must be careful in interpreting compaction rates from snow pits that are dug adjacent to each other in subsequent days or weeks. Since the terrain characteristics and weather conditions in this small plot are uniform, thus have a constant overall wind shelter index, it suggests that either small terrain differences can have some wind index effect, or that the snow is affected by upwind disturbances (e.g. terrain or vegetation). Small terrain effects could be small surface undulations (<5 cm), mimicking bottom topography and causing large enough spatial variability in near-surface wind conditions to cause a snow dune effect of differential snow deposition and/or compaction. However, at our site, it is perhaps more likely that the variability is not affected by the *in situ*

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conditions, but by those at some distance upwind. The Candy Cane plots are within 20–30 m of tree stands at the same elevation contour. Perhaps wind funneling through the tree canopy can cause considerable changes in the local wind regime over the Candy Cane plots. If this were the case, then Erickson et al.'s (2005) suggestion that "once intense sampling at a site has established the effect of topographic parameters on snow properties, predictions of the spatial distribution of snow in other years could be made without the need for intense sampling …" might only hold while the alteration of the local wind field by the canopy remains constant.

FAILURE AND SNOWPACK PROPERTIES

The loading force of skiers generally dissipates below 0.5–0.8 m (Schweizer and Camponovo, 2001). Although trigger zones occur generally in weak layers within that depth, fracture planes found in snow pits at depths ≤1 m can also be indicative of potential failure because of the shape of the base of slab avalanches (McClung and Schaerer, 1993). In our snow profiles, 88% of failures occurred within 0.8 m while 91% occurred within a depth of 1 m. The force needed for failure exponentially increased with depth, and low force failures only occurred in the top 40 cm. This suggests that, in 2004, skier-triggered large slab avalanches would have been rare. However, since snowmobilers frequent the Castle Mountain region as well, their higher loading force might trigger these deeper-rooted avalanches (c.f. Stethem et al., 2003).

Spatial variability in the Candy Cane compression test results revealed a prominent shallow low to medium force failure in 5 of the 7 pits, suggesting that 2/7 (28%) of the snow pit compression test results are not representative of the stability of the slope. This percentage concurs with findings of Landry et al. (2004), who used data from 54 pits. In contrast to Landry et al. (2004) we found a correlation between age and force needed for failure, but with only 43–58% of the variance explained. The fact that we could not detect a correlation between hardness difference and failure might be because weak layers can be very thin (in the order of mm) and might be missed in manual hardness tests (McClung and Schaerer, 1993). With our manual hardness test, 19% of our failures showed no hardness difference between the layers on either side of the failure plane. Had our density data been usable, we would have had similar problems with detecting similar small scale changes in layer density.

FORECASTING SNOWPACK PROPERTIES

Efforts towards avalanche forecasting and modeling have increased dramatically since the 1970s due to the rapid growth of winter backcountry recreation and its associated fatalities (McClung and Schaerer, 1993). Even with the tremendous efforts directed towards avalanche forecasting, actual avalanche observations are still the strongest indicators of immediate snowpack instabilities (Jamieson et al., 2001). When no recent avalanche occurrences are noticeable, snow profiles and stability tests are used to extrapolate snowpack instabilities. However, Landry et al. (2004) established that only 48% of quantified loaded column stability tests were representative of actual slope stability, Birkeland and Chabot (2006) found that 1 out of 10 stability tests give "false-stable" results, while our Candy Cane results suggest that 5/7 (71%) might be representative. These new findings imply that (1) stability tests are not as reliable as previously thought, (2) some of the null-results might represent spatial variability, and multiple profiles need to be assessed to account for this, and (3) new prediction methods are necessary.

Models such as SNOWPACK, SAFRAN-CROCUS-MEPRA, and SNTHERM (Jordan, 1991; Brun et al., 1992; Bartelt and Lehning, 2002; Durand et al., 2003) have been built in an attempt to facilitate avalanche forecasting in remote areas. These models are constantly being improved and tested but are, as of yet, too costly and/or broad-scaled for small recreational operations and recreational backcountry users. Input data of snowpack models use common meteorological parameters as such as wind velocity and direction, solar radiation, temperature, relative humidity, and precipitation. Accurate input data are essential for proper model prediction. However, model inputs are often based on point meteorological data which necessitates spatial interpolation in order to attain desired prediction accuracies. Because forecasting accuracy is dependent on the scale relationship of input data and model constraints (McClung, 2000), appropriate extrapolation of local lapse rates is an important component of making accurate local snowpack predictions.

Although spatial variability of snowpack properties (Landry et al., 2004; Kozak et al., 2003; and this study) generally poses a difficult obstacle for model-based predictions, recent models have started to include spatially variable topographic parameters (Erickson et al., 2005) and snowpack spatial structures (Kronholm and Birkeland, 2005). In these models, indexes for wind sheltering and wind drift, as well as elevation, slope, and potential radiation were found to be significant predictors of snow depth, while large spatial continuity in weak layers were found to promote propagation of fracturing over larger scales and hence cause more devastating avalanches. Kronholm and Birkeland (2005) therefore suggested that disrupting the spatial structure of the snowpack should inhibit avalanche formation.

Conclusions

The lapse rate and snowpack data reported in this paper represent the first data from the southeastern slopes of the Canadian Rocky Mountains, and highlight somewhat different controls on snowpack characteristics and spatial variability than in other more intensively studied regions. We reaffirm snowpack spatial variability and the consequent complexities in finding representative snow profile locations for stability evaluation. Only when the influence of mountain terrain lapse rates on this snowpack variability are truly understood, avalanche forecasting models could be used with less labor-intensive, and occasionally dangerous, wide-scale snowpack sampling. Our data analysis allows us to derive the following conclusions on lapse rates, snowpack properties and spatial variability and its relation to weather:

(1) Inversion events occur mainly in winter and at night, and affect about 25% of T_{min} and 10% of T_{max}.

(2) Lapse rate averages including inversion days are not representative of actual lapse rates in mountainous regions. Failure to separate inversion from non-inversion days yields physically meaningless values. If it is unclear whether there are inversion days in a region, the multi-year lapse rate mode provides a better representation of the actual lapse rate than the lapse rate average.

(3) When inversion days are removed from the Castle Mountain Resort lapse rate data, the average LR_{Tmax} is $-6.1 \pm 3.2°C\ km^{-1}$ and the average LR_{Tmin} $-5.9 \pm 3.4°C\ km^{-1}$. This is similar to the theoretical environmental lapse rate and lapse rates found in other mountainous regions. LR_{Tmax} and LR_{Tmin} modes

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were consistent, and a value of 6.3°C km^{-1} could be used as a fixed lapse rate value where data are unavailable.

(4) Minimum snowpack temperatures are positively correlated with T_{min} of the previous 24 hours, unless a prolonged cold spell has occurred, when Ts_{min} is most strongly correlated with T_{min} of the previous 4 days. No correlation between Ts_{min} depth and present or pre-profile air temperatures was found.

(5) Heat loss in the upper 10 cm of the snowpack occurs within hours, and heat loss up to 40 cm within 24 hours. Frequent strong winds in the study region might enhance turbulent heat exchange and affect upper snowpack temperatures more quickly than in other regions.

(6) Snowpack spatial variability indicates that layer thickness varies by about 25% between pits dug 5 m apart, but generally less between pits dug 30 cm apart. Settlement <25% is therefore difficult to separate from spatial variability.

(7) In the absence of terrain heterogeneity, wind is considered to be the main factor influencing microscale snowpack spatial variability.

(8) Compression test results show that failure force decreased exponentially with depth, and that stability increased with age of snow layers, unless rain crusts are present in the snowpack.

(9) Rain and sun crusts are semi-permanent low-force failure horizons. Environmental conditions related to their formation as well as to subsequent snowfall distribution should be the focus of further studies. When rainfall occurs at lower elevations, good knowledge of lapse rate conditions might allow for determination of the freezing level, which is crucial for snow stability forecasts at higher elevations.

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