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1	Impact of windflow calculations on simulations of
2	alpine snow accumulation, redistribution and ablation

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8 ABSTRACT

- 9 Wind redistribution, radiation and turbulent heat fluxes determine seasonal snow accumulation
- 10 and melt patterns in alpine environments. Mathematical representations of windflow vary in
- complexity and introduce uncertainty to snow modelling. To characterize this uncertainty, a
 spatially distributed snow model that considers the physics of blowing snow transport and
- spatially distributed snow model that considers the physics of blowing snow transport and
 sublimation and the energy fluxes contributing to snowpack ablation was evaluated for its ability
- to simulate seasonal snow patterns around a windy alpine ridge in the Canadian Rockies. The
- 15 model was forced with output from three windflow models of varying computational complexity
- 16 and physical realism: i) a terrain-based empirical interpolation of station observations, ii) a
- simple turbulence model, and iii) a computational fluid dynamics model. Compared to wind
- 18 measurements, the windflow simulations produced similar and relatively accurate (biases lower
- than $\pm 1.1 \text{ m s}^{-1}$) wind speed estimates. However, the snow mass budget simulated by the snow
- 20 model was highly sensitive to the windflow simulation used. Compared to measurements,
- 21 distributed snow model depth and water equivalent errors were smallest using either of the two
- turbulence models, with the best representation of downwind drifts by the computational fluid
- 23 dynamics model. Sublimation was an important mass loss from the ridge and windflow model
- choice resulted in cumulative seasonal sublimation differences ranging from 10.5% to 19.0% of
- seasonal snowfall. When aggregated to larger scales, differences in cumulative snowmelt and
- snow transport were negligible but persistent differences in sublimation and snow-covered area
- suggest that windflow model choice can have significant implications at multiple scales.
- 28 Uncertainty can be reduced by using physically based windflow models to drive distributed snow
- 29 models.
- 30
- 31 Keywords: blowing snow, wind, windflow model, alpine snow, sublimation

32 1. INTRODUCTION

- 33 The evolution of an alpine snowpack is greatly influenced by wind patterns. During and after
- 34 snowfall events, wind can redistribute snow from exposed areas and deposit it in sheltered
- regions (Pomeroy *et al.*, 1997; Essery and Pomeroy, 2004). In the absence of vegetation,
- topography and cumulative synoptic wind patterns determine the formation and persistence of
- snow drifts in alpine environments (Greene *et al.*, 1999; Mott *et al.*, 2010; Schirmer *et al.*, 2011)
- 38 with important ecohydrological impacts (Williams and Melack, 1991; Brooks and Williams,
- 1999; Walker *et al.*, 2001; Wipf *et al.*, 2009). Particularly in mid-winter, turbulent energy
- 40 exchange at the snow surface can exceed radiation in importance and result in sublimation losses
- 41 (Marks and Dozier, 1992; Marks and Winstral, 2001). In cold, dry and windy environments, the
- 42 additional sublimation of blowing snow can be a substantial fraction of winter snowfall
- 43 (Pomeroy, 1989). During wind transport through an unsaturated atmosphere, snow particles are
- 44 well ventilated and undergo sublimation at rates exceeding that of the snow surface (Dyunin,
- 45 1959; Schmidt, 1972; Schmidt, 1986). Sublimation losses are important to consider in cold

regions hydrological models and estimation requires accurate windflow representation (Bowling *et al.*, 2004).

Windflow also has important effects on snowmelt rates. Wind affects the spatial patterns of meltwater availability indirectly through its influence on the end-of-winter snow distribution (Pomeroy *et al.*, 1998; Pomeroy *et al.*, 2003; Grunewald *et al.*, 2010; Schirmer *et al.*, 2011; Egli *et al.*, 2012) and directly through the turbulent exchange of temperature and water vapor between the snow surface and the overlying air (Male and Granger, 1981). Pohl *et al.* (2006) and Menard *et al.* (2014) have shown that variable wind exposure over complex terrain strongly influences turbulent transfer to snow and subsequent melt rates.

55 In mountainous terrain, windflow patterns exhibit complex variability at spatial scales that complicate efforts to map the influence of topography on wind speed and direction. Many 56 models rely on terrain-based empirical calibration on available measurements (e.g., Liston and 57 58 Sturm, 1998) or terrain shelter parameterizations based on assumed mean flow fields (e.g., 59 Winstral and Marks, 2002). Linearized turbulence models such as the MS3DJH/3R model (Walmsley et al., 1982; Taylor et al., 1983; Walmsley et al., 1986) have been used to drive a 60 distributed blowing snow model (Essery et al., 1999; Fang and Pomeroy, 2009). Linear 61 turbulence models represent windflow in a more physically realistic manner than the terrain-62 based methods, but the simplified physics limits application to gentle slopes. More recently, 63 computationally intensive nonlinear turbulence models with stronger physical realism have been 64 used to downscale windflow patterns simulated by atmospheric models to simulate snow-drift 65 processes in complex terrain (Lehning et al., 2008; Mott et al., 2008; Bernhardt et al., 2009; 66 Dadic et al., 2010; Mott and Lehning, 2010). The approaches highlight a disparity in model 67

68 complexity in how windflow is commonly calculated in distributed snow model studies.

The objective of this paper is to explore warranted model complexity (Dornes *et al.*,
2008) for calculating seasonal snowpack evolution around an alpine ridge and to examine how
different windflow representations can propagate errors when used to drive a distributed blowing

72 snow and energy balance model. The study examines the impact of windflow calculations on

rd simulations of alpine snow redistribution, sublimation and subsequent melt; however, as in most

energy balance snow models, the turbulent advection from heterogeneous surface heating is notconsidered. Specific research questions include: i) what is the relative accuracy of three

76 windflow models of varying computational complexity and physical realism? ii) how sensitive

are the snow mass balance calculations of a distributed blowing snow and energy balance model

to the representation of windflow? iii) do differences in snow dynamics calculated using

79 different windflow models persist as time and space scales increase?

80 2. METHODS

81 2.1. Study site and measurements

Fisera Ridge is an alpine study site in the Marmot Creek Research Basin (50°57'N; 115°12'W),

83 in the Canadian Rocky Mountains, Alberta, Canada (Figure 1). The site is located near treeline at

84 2320 m above sea level (asl) and the land cover is primarily bare soil and alpine grasses. The

ridge has an E-NE orientation and a generally perpendicular W-NW prevailing wind (Figure 2).

86 Any winter snow deposition on the windward (NW) slope is quickly wind-scoured and deposited

in a ~100 m zone on the leeward (SE) slope downwind of the ridge crest. The leeward and

88 windward slopes are $< 20^{\circ}$ and the ridge crest is rounded with a gradual change in slope (i.e.,

89 terrain curvature).

90 Three meteorological stations were located on the windward slope (windward station), the top of the ridge (ridgetop station), and the leeward slope (leeward station) over a ~160 m 91 linear distance (Figure 1). The ridgetop station was located midway between the two stations and 92 slightly offset down the ridge crest (Figure 1). The ridgetop station recorded 15-minute averages 93 94 of 10-second measurements of air temperature and relative humidity (Campbell Scientific[®] HMP45C212 probe with a Gill radiation shield at a height of 2.3 m), incoming shortwave and 95 longwave radiation (Kipp & Zonen[®] CNR1 net radiometer at a height of 1.4 m), snow depth 96 (Campbell Scientific[®] SR50-45 ultrasonic sensor), and wind speed and direction (R.M. Young[®] 97 98 05103AP at a height of 2.6 m). Snow depth (SR50-45) and wind speed were also recorded at the windward and leeward stations with Met One[®] 013 three-cup anemometers at heights of 2.4 m 99 (windward) and 3.2 m (leeward). The nearest precipitation measurement was from a shielded 100 Geonor T200B gauge two km away in a forest clearing at 1845 m asl. After the study, an 101 102 identical gauge was installed in a sheltered area near the ridgetop station. The relationship between precipitation values measured at the two locations for the 2009 water year was used to 103 estimate a multiplicative increase with elevation (1.86) to extrapolate measurements to the Fisera 104 Ridge study plot. Precipitation measurements were corrected for gauge under-catch as in 105 MacDonald *et al.* (2010). When air temperature was $\leq 0^{\circ}$ C, relative humidity was estimated with 106 107 respect to ice following Yang et al., (2010).

Thirteen manual snow surveys of depth and density were conducted between late-January 108 and May of 2008. Surveys consisted of two bisecting transects: a slope-parallel transect extended 109 from the windward station over the ridge and down beyond the leeward station and a shorter 110 111 ridge crest transect that extended below the ridgetop station (Figure 1). Snow depth was measured every 1 - 3 m and snow density was measured every fifth depth measurement using an 112 ESC-30 snow tube and handheld spring scale when snow depth permitted (~ 20 cm < depth < 113 ~120 cm). Otherwise, depth-integrated density measurements (1000 cm³) were made at snowpits 114 115 near the automated stations. Snow density values from the nearest measurement location were used to estimate SWE from survey depth measurements. 116

Airborne light detection and ranging (LiDAR) mapping was conducted in August, 2007 (snow-free) and again on 28 March, 2008 (snow-covered). A digital elevation model (DEM) and a snow depth model at one-metre resolution were created from the data (Hopkinson *et al.*, 2012). The aerodynamic surface roughness length estimated from LiDAR-derived vegetation height and land surface classification is provided in Figure 1 (see Section 2.3.3). Note that the ridge and immediate slopes are unvegetated to sparsely-vegetated.

123

124 **2.2.** Snow model

125 Meteorological observations from the ridgetop station were used to force a physically based

- snow redistribution, mass and energy balance model at 8 m grid spacing over a 1.024 km by
- 127 1.024 km model domain centred on the Fisera Ridge study area (Figure 1). The average slope
- within the domain was $22^{\circ} \pm 7^{\circ}$ with a maximum slope value of 52°. The Distributed Snow Model (DSM) is a multi-layer soil and three-layer snow model that considers blowing snow and
- Model (DSM) is a multi-layer soil and three-layer snow model that considers blowing snow and in-transit sublimation based on a simplified version (Essery et al., 1999) of the Prairie Blowing
- 131 Snow Model (PBSM) (Pomeroy *et al.*, 1993; Pomeroy and Li, 2000). The snowpack compaction
- and thermodynamic routines are based on the JULES land surface model (Best *et al.*, 2011). The
- soil routine is described in Ménard et al. (2014). Meteorological observations other than wind
- speed and slope-projected shortwave radiation were assumed to be homogeneous. The windflow
 and blowing snow models were not fully coupled in that surface roughness (0.005 m) did not
 change with snow depth.
- Wind speed variation due to topography was estimated with three different windflow models of varying computational complexity and physical realism (see Section 2.3). The windflow models produced maps of wind speed normalized by the ridgetop station values for eight wind directions. For each direction, normalized windflow maps were provided as a library to DSM to estimate wind speed over the domain from the measured wind speed and direction at the ridgetop station.
- 143

144 **2.3.** Windflow models

145 2.3.1. 'Liston-Sturm' empirical windflow model

The simplest of the three windflow models evaluated, an empirical model by Liston and Sturm 146 147 (1998) (hereafter LS) was used with point wind speed and direction observed at the ridgetop station in conjunction with wind-topography relationships to extrapolate wind speed to grid cells. 148 While the full LS model includes a diverting algorithm (Ryan, 1977) to estimate terrain-induced 149 wind direction, the wind direction measured at the ridgetop station was uniformly applied to all 150 151 grid cells for consistency with DSM assumptions. Terrain curvature, slope, and aspect were computed from the DEM following Liston and Sturm (1998). The average terrain curvature in 152 four directions was computed with a 50 m length scale; estimated to be the average distance 153 between the ridge crest and the middle of the two slopes, or approximately half the wavelength 154 155 of Fisera Ridge. The upwind slope was computed for eight primary wind directions. For each grid cell (i,j) and wind direction (θ) a wind weighting factor, $W_{w_{ij\theta}}$, used to modify the 156 measured wind speed, was estimated from the upwind slope $(\Omega_{s_{i,i,\theta}})$ and curvature $(\Omega_{c_{i,i}})$ terrain 157 parameters, both scaled such that $-0.5 \le \Omega_{s,c} \le 0.5$, as in (Liston and Sturm, 1998): 158

159
$$W_{w_{i,j,\theta}} = 1 + \gamma_s \Omega_{s_{i,j,\theta}} + \gamma_c \Omega_c \tag{1}$$

- 160 where the additional upwind slope and curvature weighting factors (γ_s and γ_c) with a range of
- 161 [0,1] were specified as 0.5 to equally weight the importance of the two terrain parameters in
- determining the local windflow around Fisera Ridge; this value is close to that determined

163 empirically in Liston and Elder (2006). The eight wind weight maps were provided as input to

164 DSM as described in Section 2.2.

165 2.3.2. Mason-Sykes turbulence windflow model

The second windflow model evaluated was a simple linear turbulence model developed from the 166 two-dimensional theoretical work of Jackson and Hunt (1975) by Walmsley et al. (1986) and 167 168 applied to three-dimensional (3-D) topography as in Mason and Sykes (1979) (hereafter, MS). It 169 solves linearized momentum equations using Fourier transforms of the topography. The model offers more theoretical and physical realism than the empirical LS model, but does makes a 170 number of simplifying assumptions, including neutral stratification, and as a result it is only valid 171 over low hills (slopes < 25%). The MS model was run over the domain with a constant 172 roughness length of 0.005 m as in Essery et al. (1999). The model produced normalized wind 173

- speed tables for the primary wind directions.
- 175

176 2.3.3. Windsim® windflow model

177 The third and most physically based windflow model examined was the commercial Windsim[®] computational fluid dynamics (CFD) package (http://windsim.com) designed for the assessment 178 179 of wind energy resources in complex terrain. The CFD windflow model (hereafter WS), is based 180 on a 3-D Reynolds Averaged Navier Stokes (RANS) solver and uses a k - ε turbulence closure scheme (Launder and Spalding, 1974). By solving the non-linear transport equations for mass. 181 182 momentum and energy, WS offers more theoretical and physical realism than the (linear) MS 183 turbulence model and may therefore be a more suitable windflow model in mountainous terrain. A nesting technique was used to define the lateral boundary conditions of the (inner) 1.024 km x 184 1.024 km model domain. A 24 km x 24 km (outer) domain at 120 m horizontal resolution was 185 186 defined (Figure 1). The upper boundary conditions for both domains were specified with the 'constant pressure' boundary option in WS, described to be most suitable for complex terrain. 187 The lateral boundary conditions of the outer domain were specified with a logarithmic velocity 188 profile < 500 m above the terrain; above this height a constant wind profile and 20 m s⁻¹ 189 geostrophic wind speed was specified. The surface roughness of the outer domain was estimated 190 191 as a function of terrain elevation (Gravdahl and Vargvei, 1998). The logarithmic profile

- assumption is only valid over flat terrain, which is violated here, but was only used to specify the
- 193 lateral boundary conditions of the outer domain to estimate the inner domain wind profile. The
- nested domains were vertically discretized into 50 layers of 10 m thickness except for the lowest
- 195 layer, which was prescribed a 6 m thickness extending to 4 m above the DEM surface.
- 196 Experiments conducted with minimum heights < 4 m produced physically inconsistent values
- 197 indicative of numerical solution issues (not shown). Surface roughness lengths over the inner
- domain (Figure 1) were estimated from vegetation height, h, derived from LiDAR
- 199 measurements. Roughness lengths for the inner WS simulations were specified as 0.5h for $h \ge 2$
- 200 m, 0.4*h* for $0.4 \le h \le 2.0$, and a minimum of 0.005 or 0.1*h* for $h \le 0.4$ (Wallace and Hobbs,
- 201 2006). The specification of LiDAR-derived roughness lengths might be expected to improve

windflow performance over the two simpler models that either did not consider terrain roughness(LS) or that considered the roughness length to be constant (MS).

The WS windflow model produced orthogonal u, v, w wind speed vector components for each primary wind direction and specified height. Results from a height of 4 m above the inner domain (snow-free) surface were used. For each wind direction, the horizontal wind speed was calculated and the resulting wind field was normalized by the wind speed simulated at the pixel corresponding to the location of the ridgetop station.

209 2.4. Experimental design

210 **2.4.1.** Windflow model evaluation against measurements

The three windflow models were evaluated for their relative skill at simulating the observed wind speed on opposing slopes of the Fisera Ridge site. Windflow model accuracy was evaluated against 15-minute data (n=57,441) for the October, 2008 to September, 2010 period when wind data were available from all three stations. For each time step and windflow model, the ridgetop station wind direction was used to reference the corresponding windflow map. The simulated (normalized) wind speed values at locations of the windward and leeward stations were then multiplied by the wind speed measured at the ridgetop station. The model root mean squared

- error (RMSE) and bias values were computed. In addition, the modelled and measured wind
- speed values were evaluated for time steps when the wind was out of the prevailing W-NW
- 220 direction, or roughly perpendicular to Fisera Ridge.
- 221 **2.4.2.** Assessment of the impact of windflow calculation on simulated snowpack states
- Snow depth and SWE estimates from DSM forced by output from the three windflow models
- 223 were evaluated against multi-scale snowpack measurements. At the point-scale, simulated (daily)
- snow depth values at the locations of the three stations were compared to automated
- measurements. Modelled SWE was evaluated against field-based estimates derived from thirteen
- 226 (manual) snow density measurements and coincident (automated) snow depth measurements. At
- the slope-scale, model simulations of SWE along the 'T-shaped' survey transect were evaluated
- against survey measurements using nearest-neighbor averaging. Results for each windflow
- 229 model and for the respective transect-slope (i.e., windward, ridgetop, and leeward) are reported
- in terms of the mean and standard deviation of the SWE error ('modelled minus measured'). In
- addition, spatially explicit snow depth simulations for 28 March, 2008 were qualitatively
- compared to LiDAR-derived snow depth resampled from 1 m to 8 m grid spacing. Finally, the
- impacts of the three windflow calculations on both the magnitude and timing of slope-averaged
- simulated snow mass fluxes were evaluated. Simulated snow-covered area (SCA) and
 cumulative seasonal snow transport, surface and blowing snow sublimation, and melt fluxes
- were compared amongst the three windflow models. To evaluate whether different windflow
- 237 calculations impact the relative timing of simulated snow transport and sublimation, the
- normalized and cumulative frequency of the hourly fluxes were binned into 12-hours periods
- relative to the last precipitation event and the distributions were compared.

240 **3. Results**

241 The perpendicular orientation of Fisera Ridge to the prevailing wind direction (290°; Figure 2)

resulted in high measured wind speeds at the exposed windward and ridgetop station locations

with lower wind speeds on the sheltered leeward side. The average and standard deviation of the

- 15-minute wind speed measured at the windward, ridgetop, and leeward stations between 1 October, 2007 and 30 September, 2010 were $3.1 \pm 2.6 \text{ m s}^{-1}$, $2.3 \pm 2.2 \text{ m s}^{-1}$, and $2.3 \pm 1.5 \text{ m s}^{-1}$,
- respectively. The pronounced wind speed variability over relatively short distances (~ 100 m) is
- typical of windflow patterns in complex alpine terrain.

The three windflow models used to simulate wind speed on the opposing slopes produced 248 reasonable results compared to one year of measured wind speed (Figure 3). The RMSE and bias 249 values for all models were < 1.7 m s⁻¹ and better than ± 1.1 m s⁻¹, respectively (Figure 4). Model 250 errors were generally similar as indexed by the correlation coefficients (Figure 3) and RMSE 251 252 values (Figure 4). The LS model slightly overestimated wind speed on both the leeward and 253 windward slopes (Figures 3 and 4). The MS model also underestimated wind speed on both slopes and was the only model with negative wind speed biases (Figures 3 and 4). The WS 254 model exhibited near-zero mean model biases (Figure 4), but was prone to overestimating high 255 wind speeds (Figure 3). 256

Automated and manual snow measurements indicated that both the windward and ridgetop sites were largely wind-scoured with seasonal average snow depths around 20 cm (Figure 5). The deepest snowpack accumulated on the wind-exposed slopes during a series of spring snowfall events when wet snow conditions restricted wind erosion (April - June). In contrast to the wind-scoured slopes, a large drift accumulated on the leeward slope where snow depths ranged between 100 and 180 cm and SWE exceeded 600 mm (Figure 5). At all sites, maximum SWE occurred in early-May.

DSM forced by the three windflow models produced distinct differences in the seasonal 264 evolution, magnitude and location of simulated snow drifts (Figure 5). All DSM runs simulated 265 266 the mid-winter scour of the windward slope quite well, although the late-spring accumulation events were uniformly overestimated. Compared to depth and SWE measurements at the 267 windward station, the MS turbulence model resulted in the lowest RMSE and bias values while 268 the empirical LS model, and particularly the CFD WS windflow model, caused overestimated 269 270 accumulation on the wind-exposed slope (see Table 1). At the ridgetop station, greater differences in depth and SWE were simulated amongst the three model runs (Figure 5). As on 271 272 the windward slope, the MS-driven DSM best represented the frequent wind-scour of snow at the ridgetop station with small depth and SWE biases of +3.5 cm and +11 mm, respectively (Table 273 274 1). Conversely, the LS-driven DSM erroneously simulated a large drift near the ridgetop station with large depth and SWE biases of +73.9 cm and +419 mm, respectively (Figure 5 and Table 275 1). All DSM runs simulated drift formation on the leeward slope, but generally underestimated 276 277 the magnitude. The WS-driven DSM was closest to accurately simulating the leeward drift, with 278 depth and SWE biases of -10 cm and -66 mm, respectively (Figure 5 and Table 1). The MS- and 279 LS-driven DSM runs significantly underestimated the leeward SWE with mean biases of -157

280 mm and -259 mm, respectively. In general, improved SWE estimation was obtained with the281 more physically based windflow models (MS and WS).

To better understand the cause of the simulated snowpack differences as determined at 282 the individual stations, the following metrics were evaluated along the 160 m linear transect 283 284 between the windward and leeward stations: 1) modelled wind speeds in the prevailing wind direction (290°) relative to that measured at the ridgetop station (Figure 6a), 2) the change in the 285 modelled wind speed with distance du/dx (Figure 6b) and 3) the simulated SWE (Figure 6c) 286 over the ridge transect elevation profile (Figure 6d). The WS and LS models simulated a 287 288 decrease in wind speed from the windward to leeward sides while the MS model simulated wind speeds on the ridgetop and windward slope, but a greater leeward decline in wind speed than the 289 other two models. Comparatively, the LS model simulated a relatively smooth wind speed 290 transition from the windward to leeward slopes. Breaks in the wind speed slope were greater in 291 the two turbulence models than the LS windflow model, but were simulated in different locations 292 293 along the ridge transect (Figure 6b). DSM modelled SWE (Figure 6c) varied significantly along the ridge transect and that variability was windflow-model dependent. In general, DSM forced 294 by the two turbulence models simulated the greatest SWE on the leeward slope with DSM forced 295 by the WS model simulating the drift slightly closer to the ridgetop on the leeward side than the 296 MS-driven model. DSM forced by the empirical LS model erroneously simulated this drift 297 slightly to the windward side of the ridge. 298

Compared to the 28 March LiDAR snow depth estimates, the greatest differences in the 299 snow depth patterns from DSM forced by the three windflow models were amongst the empirical 300 LS model and the two turbulence models (Figure 7). The LS model resulted in a smoothly 301 302 varying snow-cover, deepest in proximity to the ridge crest and shallowest on both the windward and leeward slopes. This is in contrast to the general understanding of snow accumulation around 303 alpine ridges (Pomeroy and Gray, 1995). By comparison, the two turbulence models resulted in 304 snow-cover patterns that were similar to the LiDAR derived snow cover with shallow snow and 305 snow-free areas on the windward and ridgetop zones, and deep and spatially heterogeneous drifts 306 covering much of the leeward slope. DSM forced by the two turbulence models simulated the 307 deepest snowpack (> 200 cm) in roughly similar locations, with the WS-driven DSM simulated 308 drift forming slightly closer to the ridge crest than the MS-driven model as described in the 309 310 transect evaluation. Note that the LiDAR product indicates deep drifts around small trees in the southern- and eastern-most parts of the domain (see roughness heights in Figure 1); these areas 311 are included in the LiDAR depth map for completeness, but the inclusion of sparsely vegetated 312 areas prevents direct quantitative comparison of the measured and modelled products because the 313 314 DSM does not include vegetation roughness impacts on snowpack distribution and ablation. Time-series of the seasonal evolution of simulated SWE is provided in Figure 8. Notably, the 315 29 April snow-cover extent is greater than the mid-winter distributions as a result of wetter 316 spring snow conditions and an associated lower likelihood of wind transport (Li and Pomeroy, 317 1997); this dynamic is recorded in the observations (Figure 5) and is generally captured by DSM 318 regardless of the windflow model. 319

320 Slope- and windflow model-specific SWE errors, computed as the seasonal average error

- against data from the 13 snow surveys, show the general overestimation of SWE on the
- windward slope and ridgetop by the LS-driven DSM (299±135 mm and 311±123 mm,
- respectively) and, to a lesser extent, by the WS-driven DSM (138±98 mm and 142±91 mm,
- respectively) (Figure 9). DSM forced by the MS turbulence model outperformed SWE estimated
- by DSM forced by the other two windflow models at the two wind-exposed areas $(35\pm59 \text{ mm})$
- and -23 ± 75 mm, respectively). On the leeward side of Fisera Ridge, DSM forced by any
- 327 windflow model underestimated SWE, but the WS model had significantly reduced errors (-
- 28±91 mm) relative to DSM driven by the LS (-114±97 mm) and MS (-131±86 mm) windflow
 models (Figure 9).

330 Differences in the impact of windflow calculations on snow regime estimation (i.e., depth and SWE) were largely manifested in how the windflow models impacted the calculation of 331 332 seasonal snow fluxes including transport and sublimation. The greatest concurrence in simulated 333 transport, sublimation, melt, and SCA amongst the simulations forced by the three windflow models occurred for the leeward slope (Figure 10), where wind speeds were lowest by all 334 estimates (Figure 2). The greatest deviation in cumulative blowing snow transport and 335 sublimation due to the windflow model occurred at the ridgetop station, where the MS-driven 336 337 DSM, found to be most accurate in terms of depth and SWE, generated the greatest snow transport (out) and the highest sublimation fluxes. The WS- and LS-driven DSM simulated 338 ~50% and ~25%, respectively, of the cumulative seasonal (total) sublimation losses calculated 339 by the MS-driven DSM. Only the LS model at the ridgetop station resulted in cumulative 340 transport estimates that differed in sign from the other models in that snow accumulated at the 341 342 ridgetop; the other model runs transported the snow off the ridgetop to the leeward slope.

The location of the greatest (total) sublimation losses was windflow model-dependent: 343 sublimation was highest on the windward slope with the LS- and WS-driven DSM, but on the 344 ridgetop with the MS-driven DSM (Figure 10). On average, cumulative surface sublimation 345 346 losses were approximately 50% of the cumulative blowing snow sublimation losses. Blowing snow sublimation, reported as a percentage of cumulative seasonal snowfall, ranged from 8% 347 (leeward station) to 20% (windward and ridgetop stations). On average across the three slopes 348 (windward, ridgetop, and leeward rectangles in Figure 1), blowing snow sublimation losses with 349 350 the MS and WS models were 19% and 17.5% of cumulative seasonal snowfall, respectively, while the average loss with the LS windflow model was only 10.5%. The sublimation source 351 also exhibited seasonality; blowing snow sublimation generally ceased at the beginning of 352 March, while most of the seasonal surface sublimation occurred from March through July 353 354 (Figure 10). Blowing snow sublimation estimated by DSM forced with the two turbulence windflow models were similar to those in MacDonald et al. (2010) (19%) using the Cold 355 Regions Hydrological Model for the same year and at the same site but forced by measured 356 rather than simulated wind speeds. 357

To put the DSM results into context with those of model studies that treat blowing snow sublimation as a self-limiting mechanism, the meteorological observations and DSM blowing 360 snow sublimation estimates from the largest blowing snow event of the 2008 winter are provided (Figure 11). The event substantially redistributed alpine snow as is evident in the before and after 361 photographs. For simplicity, only results from DSM forced with the WS windflow model are 362 included in Figure 11. Following a period of light snow, low temperatures (-5° to -10°C), low 363 364 wind speeds (1 m/s), and saturated relative humidity with respect to ice (100%) on the morning of Feb. 28, the snowfall stopped, air temperature plateaued at -4°C, relative humidity dropped to 365 ~60%, and wind speed steadily increased (Figure 11). Two (hourly-average) wind speed maxima 366 were measured on Feb. 29: one at 01:00 (15 m/s) and the other at 07:00 (19.5 m/s). The DSM 367 simulated minor blowing snow fluxes (< 3.3 mm/hr; < 2 hrs.) corresponding to the timing of the 368 first wind speed maxima before a more substantial blowing snow event lasting ~4 hrs. with 369 maximum sublimation estimates of 13.3 mm/hr, 5.8 mm/hr, and 1.9 mm/hr on the windward, 370 ridgetop, and leeward sides, respectively occurring at 08:00 on Feb. 29 (Figure 11). Simulated 371 372 blowing snow sublimation stopped after four hours (10:00) and the wind speed dropped below 373 15 m/s. The air temperature measured at the ridgetop station steadily increased from -4.4°C at the beginning of the large blowing snow event (05:00 Feb. 29) to -1.7°C (10:00) and the relative 374 humidity dropped slightly from 64% (05:00) to a minimum of 55% during the simulated blowing 375 snow maximum (08:00) and increased to 59% by the end of the event (10:00) (Figure 11). 376 377 Early in the melt period, cumulative snowmelt was insensitive to windflow representation,

and only became sensitive late in the season as differences in SCA depletion among the models
dictated meltwater availability (Figure 10). The leeward slope generally had the greatest SCA
with the latest snow-cover depletion, while the wind-scoured windward slope sustained an
intermittent snow-cover (Figure 10). The LS model resulted in the smoothest and most
homogeneous snow-cover (Figure 7) as well as the greatest SCA and latest snow-cover depletion
on all slopes. In contrast, the MS model resulted in the most variable SCA and the WS model
caused a gradual SCA change from intermittent (windward) to complete (leeward).

The results show that the windflow model choice can have significant implications for snow 385 386 regimes and snow fluxes at point- to slope-scales. When averaged over the full model domain the differences in transport and melt were subtle to negligible; however, more appreciable 387 differences in sublimation and snow-cover depletion suggest that windflow model choice can 388 have important implications at multiple scales (Figure 10; right-most column). The windflow 389 390 model choice not only influenced the magnitude of seasonal blowing snow transport and sublimation fluxes, but also the timing of these fluxes relative to snowfall events. In general, 391 DSM simulated a large majority of seasonal (hourly) blowing snow transport to occur between 392 13 and 24 hours after a snow event (Figure 12). On average, this trend was consistent across the 393 three slopes; however, depending on the windflow model, the fraction of seasonal blowing snow 394 transport during this 12 hour period varied by as much as 20%. Conversely, less than 1% of the 395 cumulative seasonal snow transport was simulated to occur more than 72-hours after a snowfall. 396 The windflow model choice had a lesser impact on the timing of sublimation losses. It is 397 398 interesting to note that >90% of the seasonal blowing snow sublimation losses and <55% of the

surface sublimation losses were simulated to occur within 36 hours of snowfall (Figure 12), withthe most surface sublimation occurring during the melt season (>72 hours).

401 **4. Discussion**

When forced with ridgetop windflow observations, all three windflow models adequately 402 captured the general pattern of high wind speeds on the exposed windward side of the alpine 403 ridge and lower wind speeds on the protected leeward side. The perpendicular nature of the 404 prevailing wind direction recorded at Fisera Ridge was remarkably persistent (Figure 2) as a 405 406 combined result of local terrain orientation and regional flow patterns. The slope-parallel 407 windflow persistence likely facilitated model accuracy by placing less emphasis on model skill at simulating windflow direction relative to the reference station, and more emphasis on wind 408 velocity representation. As such, the model comparison represents a 'best-case' scenario that 409 410 provides important insight into the impacts of windflow calculations on simulations of alpine snow redistribution and ablation. 411

Compared to measurements, the MS turbulence model had the greatest bias on both slopes 412 and highest RMSE on the windward slope (Figure 4). As previously noted, the empirical LS 413 windflow model weighting factors upwind slope and curvature were not determined from local 414 calibration, but specified as in previous empirical studies to be more consistent with how an 415 empirical windflow model might be applied to complex terrain. Despite the lack of local 416 calibration, when compared to measured wind speed on the two slopes the empirical LS model 417 performed as well as the WS model (in terms of the RMSE values) and better than the MS 418 model. For example, the MS simple turbulence model had the greatest average wind speed bias 419 of -0.95 m s⁻¹ compared to the relatively smaller biases of the LS (0.25 m s⁻¹) and WS (0.05 m s⁻¹) 420 421 ¹) windflow models (Figure 4). However, the windflow model evaluation against windward and 422 leeward slope wind speeds was a poor indicator of how wind speed errors might propagate into DSM snow state errors and flux differences. 423

The three windflow models used to force DSM had appreciable and varying impact on the 424 calculation of seasonal snow mass balance (i.e., depth, SWE) and fluxes (i.e., transport, 425 426 sublimation and melt). The two turbulence models resulted in the deepest snowpack (> 200 cm) 427 in terrain-sheltered locations downwind of the ridgetop (Figures 7 and 8). By comparison, the LS-driven DSM simulated a smoothly varying snow-cover, deepest in proximity to the ridgetop 428 and shallowest on both the windward and leeward slopes. The results suggest that improved 429 430 performance of the empirical LS windflow model might have been obtained from reducing the distribution of weight on the curvature parameter and increasing the weight on the upwind slope 431 parameter; however, there is no guarantee that calibration of LS against wind speed alone would 432 have improved its performance in simulating the spatial distribution of SWE. The MS model 433 resulted in the lowest snowpack depth and SWE errors on the windward slope and ridgetop and 434 435 WS resulted in the lowest errors on the leeward slope (Figures 5 and 9; Table 1). These results contrast with the evaluation of wind speed simulations discussed previously and imply that, 436

particularly in high-wind environments such as the ridgetop and windward slope where MS was
not the most accurate wind speed model, the representation of precisely how much the snow
transport wind speed threshold was exceeded may be of secondary importance for snow transport
calculations to the representation of wind speed spatial variability.

441 Modelled wind speed acceleration or deceleration indicated by positive and negative du/dx442 values, respectively, (Figure 6b) determines whether snow simulated at a grid element is scoured or deposited. The variation in the sign, magnitude, and spatial location of the simulated breaks in 443 wind speed among the three models indicate substantial fine-scale differences in windflow 444 representation (Figure 6b) that contribute to differences in the snow depth and SWE estimates 445 (Figure 6c). The smoothly varying snow-cover simulated by the LS-driven DSM is attributed to 446 447 the low variation and small (absolute) values of the du/dx values estimated by the LS windflow model. By comparison, substantial variation in du/dx values simulated by the two turbulence 448 windflow models resulted in higher variability in simulated SWE (Figure 6c). The results suggest 449 450 that the turbulence models can represent windflow (and SWE) variability at two scales: i) slopescale terrain effects such as the windward and leeward sides of a ridge, and ii) small-scale (i.e., < 451 10 m) effects of slight terrain undulations. Differences in the exact position of wind speed breaks 452 over the ridge simulated by the turbulence models are likely due to structural disparities between 453 the linear (MS) and nonlinear (WS) windflow models. The empirical LS model may have been 454 455 able to capture these micro-scale wind speed variations with a smaller length-scale curvature parameter, but such a parameter change may come at the expense of reduced slope-scale 456 accuracy, i.e., the curvature metric would then be more sensitive to small-scale terrain features 457 458 than slope-scale features such as the ridge. While overall errors in estimating snow depth and 459 SWE were generally smallest using either of the two turbulence windflow models compared to the empirical model (Table 1), the ability of WS to estimate the leeward slope drift is notable for 460 two reasons: the snowpack mass balance at Fisera Ridge is dependent upon accurately simulating 461 upwind snow transport and in-transit sublimation; and the estimation of hydrologically important 462 463 leeward drifts is one of the main reasons to run a blowing snow model.

464 The models evaluated here assume that the wind direction is constant for all grid elements for 465 a given time step and do not consider terrain-induced alterations to the windflow direction. In locations where the wind direction varies little and topography is simple, such as Fisera Ridge, 466 the computational efficiency of assuming a constant wind direction may outweigh potential 467 deficiencies in model performance due to the assumption. When wind direction over a domain is 468 469 unknown and terrain is more complex, then windflow patterns should be estimated based on terrain characteristics (e.g., Ryan, 1977) or within a turbulence (e.g., Essery et al., 1999) or 470 atmospheric (e.g., Mott et al., 2014) model. Errors in the simulated drift formation compared to 471 measurements can accrue from the steady state assumption of the blowing snow model which 472 does not include a realistic temporal and spatial lag in the formation of snow deposition features 473 474 after a drop in wind speed on a lee slope. Non-steady-state blowing snow models are in their 475 infancy due to an incomplete understanding of turbulent snow particle interactions in complex

terrain. Despite these challenges, for the general application to areas of limited terrain

- complexity such as presented here, the DSM results suggest that the more physically realistic
- 478 turbulence models are an example of warranted model complexity over the empirical LS470 windflow model
- 479 windflow model.

480 It is shown that cumulative seasonal snow transport and sublimation losses can be 481 significant and are sensitive to the windflow characterization. When averaged over the ridge, the 482 cumulative seasonal blowing snow sublimation losses relative to seasonal snowfall simulated by DSM when forced with the MS (19%) and WS (17.5%) windflow models were similar to 483 estimates in MacDonald et al. (2010) (19%); note that the empirical LS windflow model caused 484 substantially lower estimates of blowing snow losses (10.5% of seasonal snowfall). The 485 486 differences imply that the windflow model choice can have significant implications on slope-487 scale hydrology, ecology, and land surface representation; topics that require accurate 488 characterization of snow-cover duration and snow drift magnitude.

The largest blowing snow event of the 2008 winter was accompanied by increases in 489 490 both the 2.3 m air temperature and saturation deficit (Figure 11). The observations support previous multi-height measurements made at a Canadian Prairie site (Pomeroy (1988) as 491 reported in Pomerov and Li (2000)) where the process was attributed to dry air advection that 492 resulted from the mixing of initially stable boundary layers. The field examples suggest that 493 atmospheric boundary layer models must consider more thermodynamic phenomena than the 494 495 negative feedback process (Pomeroy and Li, 2000), particularly in wind-prone complex terrain 496 such as the Canadian Rocky Mountains. Future development of fully coupled atmospheric and blowing snow models, validated by multi-height field observation, may provide useful insight 497 into the relative and often compensatory roles of blowing snow sublimation, moisture and 498 499 temperature feedback, and dry air advection mechanisms.

500 DSM estimated that the majority of cumulative seasonal snow transport and blowing snow sublimation occurred in the 13 - 24 hour period after a storm event (Figure 12), illustrating the 501 502 importance of considering blowing snow threshold conditions and in-transit sublimation in 503 calculating snow redistribution. The results raise questions about how simple snow redistribution models that immediately reallocate snowfall (e.g., Winstral and Marks, 2002) without 504 considering in-transit sublimation might result in the propagation of SCA and sublimation errors. 505 The accurate characterization of SCA is required to simulate the surface albedo, temperature, and 506 energy balance that are important for models that simulate atmospheric and hydrological 507 dynamics (Shook et al., 1993; Pomeroy et al., 1998). For example, the windflow model used to 508

- force DSM impacted the simulation of late-lying snow patches known to enhance alpine albedo and provide meltwater to alpine and subalpine lakes, wetlands and streams (Elder *et al.*, 1991).
- 511 During the spring and summer, water availability in alpine landscapes is influenced by winter
- snow drift patterns, which in turn critically impacts vegetation distribution (Billings and Bliss,
- 513 1959; Walker et al., 2001), soil moisture (Taylor and Seastedt, 1994), contaminant loading
- 514 (Pomeroy *et al.*, 1991) and nutrient cycling (Williams and Melack, 1991). At the slope-scale in

complex terrain, distributed blowing snow models require realistic windflow models toaccurately simulate these ecohydrological processes.

517 Finally, when DSM snow mass fluxes were spatially aggregated to include a larger area (~1 km²), which included less wind-prone areas, the windflow model-related differences in the time 518 519 evolution of aggregated snow transport and melt were subtle to negligible; however, there were 520 appreciable differences in sublimation and snow-cover depletion. The low sensitivity of 521 simulated spring melt fluxes to the windflow calculations may be underestimated but the error is 522 difficult to quantify due to the uncertainty in advection parameterisations for complex terrain snowmelt calculations. While not considered here, turbulent advection of sensible heat can 523 influence snow-cover depletion rates (Shook et al., 1993; Mott et al., 2014). Turbulent advection 524 525 on Fisera Ridge is considered to be relatively small because while snow-cover is rapidly depleted on the windward slope and ridgetop it persists in a large, continuous drift on the leeward slope 526 527 leading to one large snow patch with one leading edge. Thus, the spring snow-cover depletion patterns at Fisera Ridge differ from the patchy snowpack with a wide distribution of snow patch 528 529 sizes and fetch lengths that have been studied in the Canadian Prairies or Arctic (Shook et al., 1993; Granger et al., 2002). The driving meteorological data for DSM was collected at a ridgetop 530 531 station that would be over snowcover when the entire domain was snow-covered and mostly snow-free when only the leeward slope snowcover remained and so may have inherently 532 included some advected energy. While not explicitly considered, any additional turbulent energy 533 534 from advection may have propagated the reported differences in the estimated end-of-winter 535 SWE distribution amongst the windflow model-forced snow simulations due to inherent feedback processes between SCA and the advection of sensible heat (Marsh and Pomeroy, 536 1996). Therefore, windflow model choice may have more influence on late-spring snow-cover 537 538 depletion rates and the time evolution of spatially aggregated spring snowmelt than reported 539 here.

The results suggest that the issue of warranted model complexity should be weighed in careful consideration of the processes of interest, the model used, and the modelling objectives. The variability of aggregated snow states, mass fluxes and SCA amongst windflow model-driven DSM runs over landscape units corresponding to the windward, ridgetop and leeward slopes in Figure 10 is substantial and suggests that improved simulations at the landscape unit scale can be gained by using turbulence-based windflow models.

546 **5.** Conclusions

Compared to automated and manual measurements made on opposing sides of an alpine ridge,
DSM forced by the three windflow models produced distinct differences in the seasonal
evolution, magnitude and location of simulated snow drifts. The empirical LS-driven DSM
simulated a smoothly varying snow-cover, deepest in close proximity to the ridge crest and
shallowest on both the windward and leeward slopes. This was in contrast to the general
understanding of snow accumulation around alpine ridges. By comparison, the two turbulence
windflow model-driven DSM runs simulated snow-cover patterns that were similar to the

- LiDAR-derived snow-cover with shallower snow and snow-free areas on the windward and
- ridgetop zones, and a deeper drift covering much of the leeward slope. DSM forced by the two
- turbulence models simulated the deepest snowpack (> 200 cm) in roughly similar locations. The
- 557 WS-driven DSM provided the most accurate snow simulation on the leeward slopes where large
- 558 drifts accumulate due to snow transport from upwind slopes. On average, cumulative surface
- sublimation losses were approximately 50% of the cumulative blowing snow sublimation losses,
 which were 19% and 17.5% of the cumulative seasonal snowfall with the MS and WS turbulence
- 561 models, but only 10.5% with the LS empirical windflow model. Strong seasonality was detected
- 562 in the sublimation source; blowing snow sublimation generally ceased at the beginning of March,
- while most of the seasonal surface sublimation occurred from March through July. The location
- of the greatest (total) sublimation losses was windflow model-dependent: sublimation was
- highest on the windward slope with the LS- and WS-driven DSM, but on the ridgetop with the
- 566 MS-driven DSM. The results show that the windflow model choice can have significant
- 567 implications for calculating snow regimes and all snow mass fluxes at point- to slope-scales that
- are important for alpine ecology and at landscape scales relevant to hydrological and climate
- 569 models that consider sub-grid or sub-basin variability.

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Table 1. Snow depth and SWE errors for the snow simulations forced by the three windflow

- models as evaluated against snow observations made at the three stations. The shaded cells
- 711 indicate the windflow model that produced the lowest error values for each station. Note that the
- 712 depth errors were calculated from mean daily automated measurements while the SWE errors
- were the average error values computed on manual observations at near each station during 13
- 714 repeated snow surveys.

		RMSE			bias	
	windward	ridgetop	leeward	windward	ridgetop	leeward
_f LS	22.1	83.4	53.7	16.6	73.9	-38.9
SM E df	14.4	9.3	38.1	7.5	3.5	-22.7
\cap WS	29.3	25.7	27.0	21.4	20.9	-10.0
LS	91	429	270	70	419	-259
B a a b a b a b a b a b a b a b a b a b	57	38	170	37	11	-157
∽ ¯ WS	131	110	84	97	101	-66

716 FIGURE CAPTIONS

- Figure 1: Study site map showing the nested model domains (24 km x 24 km with 150 m
- elevation contour lines; 1.024 km x 1.024 km with 10 m elevation contour lines) centred on the

719 locations of three meteorological stations on the alpine Fisera Ridge in the Marmot Creek

Research Basin, Alberta, Canada (location indicated by the star in the upper-right panel). The

- small maps at right (20 m elevation contour lines) indicate the (top): LiDAR-derived roughness
- length values over the inner domain and (bottom): the locations of the snow survey transects
- relative to the three meteorological stations. Rectangular domains used to compare spatially
- averaged simulated fluxes representative of the windward, ridgetop, and leeward parts of Fisera
- 725 Ridge are shown.
- Figure 2: Wind roses including the mean and maxima wind speeds for the windward, ridgetop
- and leeward stations from 15-minute averaged data collected from October, 2008 to September,
- 2010 (n=57,441). Analysis was limited to time-steps when data were available from all three
- stations. Note that wind direction was only measured at the ridgetop station and was assumed
- representative of the two other stations for the purposes of the wind rose comparison.
- Figure 3: Scatter plots of wind speed comparing modelled values (y-axes) from each of the three
- windflow models (panel rows) to measured values (x-axes) at the windward (left panels) and
- regression fits, coefficients of leeward (right panels) automated weather stations. The (linear) regression fits, coefficients of
- 734 determination (R^2) , and correlation coefficients (r) are indicated.
- Figure 4: Mean (bars) and standard deviation (lines) of modelled wind speed RMSE (top) and
- bias (bottom) for the three windflow models compared to measurements at the windward and
- 737 leeward automated weather stations.
- Figure 5: Measured snow depth (top) and SWE (bottom) compared to simulated values from the
- three windflow models (lines) at the windward (left panel column), ridgetop (centre panel
- column), and leeward (right panel column) stations.
- Figure 6: Modelled a) seasonal mean wind speed normalized by the ridgetop station
- observations, b) the change in mean wind speed with distance (du/dx), and c) snow water
- requivalent (SWE) presented as the pixel-wise nearest-neighbor mean (lines) and standard
- deviation (shading) near the time of seasonal maximum accumulation (May 1, 2008) along d) a
- 160 m linear (12 m vertical) transect from the windward to leeward sides of the alpine ridge.
- Figure 7: The LiDAR-measured snow depth on Fisera Ridge on 28 March, 2008 (left) compared
- to that simulated by DSM forced by wind speed output from the Liston-Sturm (LS), Mason-
- 748 Sykes (MS) and Windsim (WS) windflow models. The location of the windward (red marker),
- ridgetop (black marker), and leeward (blue marker) stations are indicated. The elevation contour
- 750 lines are included.

- Figure 8: Distributed maps of SWE (color scale) near the Fisera Ridge stations (markers) on the
- dates of select snow surveys (panel rows) as simulated by DSM forced with output from the
 three windflow models (panel columns). Elevation contour lines are included.
- Figure 9: Model SWE error computed as the mean (bars) and standard deviation (lines) of
- ⁷⁵⁵ 'modelled measured' SWE averaged along the snow survey transects for 13 surveys on the
- vindward (left), ridgetop (centre), and leeward (right) sides of Fisera Ridge for the DSM model
- forced by wind speed output from the Liston-Sturm (LS), Mason-Sykes (MS) and Windsim
- 758 (WS) windflow models.
- Figure 10: Cumulative fluxes of snow transport, sublimation (total, surface and blowing snow
- losses), melt, and snow covered area averaged within domains centered on the windward,
- ridgetop, leeward, and the entire domain as simulated by DSM forced with wind speed output
- 762 from the Liston-Sturm (LS), Mason-Sykes (MS) and Windsim (WS) windflow models for the
- 763 2007-2008 snow season.
- Figure 11: Measured values of wind speed, air temperature, precipitation (left axis) and relative
- humidity (right axis) during a blowing snow event on February 29, 2008 at the Fisera Ridge
- (ridgetop) station. Blowing snow sublimation rates estimated by DSM forced with windflow
- output from the Windsim model at the locations of the windward, ridgetop and leeward stations
- are included. Photographs from a field camera mounted on the ridgetop station looking northwest
- toward the windward slope show snow cover before (15:00 Feb. 28) and after (12:00 Feb. 29) the
- 770 blowing snow event.
- Figure 12: The timing of normalized (left y-axes) and cumulative (right y-axes) hourly seasonal
- snow transport and sublimation (total, surface and blowing snow) fluxes, binned in 12-hour
- intervals since the last snowfall (x-axes), as simulated by DSM forced by wind speed output
- from the Liston-Sturm (LS), Mason-Sykes (MS) and Windsim (WS) windflow models for the
- 775 2007-2008 snow season.

Table 1. Snow depth and SWE errors for the snow simulations forced by the three windflow

- models as evaluated against snow observations made at the three stations. The shaded cells
- indicate the windflow model that produced the lowest error values for each station. Note that the
- 780 depth errors were calculated from mean daily automated measurements while the SWE errors
- were the average error values computed on manual observations at near each station during 13
- 782 repeated snow surveys.

	RMSE			bias		
	windward	ridgetop	leeward	windward	ridgetop	leeward
_ LS	22.1	83.4	53.7	16.6	73.9	-38.9
MS CH bf	14.4	9.3	38.1	7.5	3.5	-22.7
$^{\Box}$ WS	29.3	25.7	27.0	21.4	20.9	-10.0
LS	91	429	270	70	419	-259
∦a ∰ MS	57	38	170	37	11	-157
∞ ⊓ WS	131	110	84	97	101	-66





Figure 1: Study site map showing the nested model domains (24 km x 24 km with 150 m

elevation contour lines; 1.024 km x 1.024 km with 10 m elevation contour lines) centred on the

787 locations of three meteorological stations on the alpine Fisera Ridge in the Marmot Creek

Research Basin, Alberta, Canada (location indicated by the star in the upper-right panel). The

small maps at right (20 m elevation contour lines) indicate the (top): LiDAR-derived roughness

result result in the second result is the location of the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result in the second result is the second result in the second result

relative to the three meteorological stations. Rectangular domains used to compare spatially

averaged simulated fluxes representative of the windward, ridgetop, and leeward parts of Fisera

793 Ridge are shown.



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Figure 2: Wind roses including the mean and maxima wind speeds for the windward, ridgetop

and leeward stations from 15-minute averaged data collected from October, 2008 to September,

2010 (n=57,441). Analysis was limited to time-steps when data were available from all three

stations. Note that wind direction was only measured at the ridgetop station and was assumed

representative of the two other stations for the purposes of the wind rose comparison.



Figure 3: Scatter plots of wind speed comparing modelled values (y-axes) from each of the three
windflow models (panel rows) to measured values (x-axes) at the windward (left panels) and
leeward (right panels) automated weather stations. The (linear) regression fits, coefficients of

804 determination (R^2), and correlation coefficients (r) are indicated.



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Figure 4: Mean (bars) and standard deviation (lines) of modelled wind speed RMSE (top) andbias (bottom) for the three windflow models compared to measurements at the windward and

807 bias (bottom) for the three windflow808 leeward automated weather stations.





Figure 5: Measured snow depth (top) and SWE (bottom) compared to simulated values from the

- three windflow models (lines) at the windward (left panel column), ridgetop (centre panel
- column), and leeward (right panel column) stations.





observations, b) the change in mean wind speed with distance (du/dx), and c) snow water

equivalent (SWE) presented as the pixel-wise nearest-neighbor mean (lines) and standard

deviation (shading) near the time of seasonal maximum accumulation (May 1, 2008) along d) a

820 160 m linear (12 m vertical) transect from the windward to leeward sides of the alpine ridge.



Figure 7: The LiDAR-measured snow depth on Fisera Ridge on 28 March, 2008 (left) compared

to that simulated by DSM forced by wind speed output from the Liston-Sturm (LS), Mason-

- 824 Sykes (MS) and Windsim (WS) windflow models. The location of the windward (red marker),
- ridgetop (black marker), and leeward (blue marker) stations are indicated. The elevation contour
- 826 lines are included.





- Figure 8: Distributed maps of SWE (color scale) near the Fisera Ridge stations (markers) on the
- dates of select snow surveys (panel rows) as simulated by DSM forced with output from the
- three windflow models (panel columns). Elevation contour lines are included.



Figure 9: Model SWE error computed as the mean (bars) and standard deviation (lines) of

⁸³⁴ 'modelled - measured' SWE averaged along the snow survey transects for 13 surveys on the

835 windward (left), ridgetop (centre), and leeward (right) sides of Fisera Ridge for the DSM model

836 forced by wind speed output from the Liston-Sturm (LS), Mason-Sykes (MS) and Windsim

837 (WS) windflow models.



Figure 10: Cumulative fluxes of snow transport, sublimation (total, surface and blowing snow
losses), melt, and snow covered area averaged within domains centered on the windward,
ridgetop, leeward, and the entire domain as simulated by DSM forced with wind speed output

from the Liston-Sturm (LS), Mason-Sykes (MS) and Windsim (WS) windflow models for the

843 2007-2008 snow season.



Figure 11: Measured values of wind speed, air temperature, precipitation (left axis) and relative

humidity (right axis) during a blowing snow event on February 29, 2008 at the Fisera Ridge

848 (ridgetop) station. Blowing snow sublimation rates estimated by DSM forced with windflow

849 output from the Windsim model at the locations of the windward, ridgetop and leeward stations

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toward the windward slope show snow cover before (15:00 Feb. 28) and after (12:00 Feb. 29) the

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Figure 12: The timing of normalized (left y-axes) and cumulative (right y-axes) hourly seasonal
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858 2007-2008 snow season.