Modelling the spatial–temporal variability of spring snowmelt in an arctic catchment

S. Pohl and P. Marsh*
National Water Research Institute, Saskatoon, SK, Canada

Abstract:
Arctic spring landscapes are usually characterized by a mosaic of coexisting snow-covered and bare ground patches. This phenomenon has major implications for hydrological processes, including meltwater production and runoff. Furthermore, as indicated by aircraft observations, it affects land-surface–atmosphere exchanges, leading to a high degree of variability in surface energy terms during melt. The heterogeneity and related differences when certain parts of the landscape become snow free also affects the length of the growing season and the carbon cycle.

Small-scale variability in arctic snowmelt is addressed here by combining a spatially distributed end-of-winter snow cover with simulations of variable snowmelt energy balance factors for the small arctic catchment of Trail Valley Creek (63 km²). Throughout the winter, snow in arctic tundra basins is redistributed by frequent blowing snow events. Areas of above- or below-average end-of-winter snow water equivalents were determined from land-cover classifications, topography, land-cover-based snow surveys, and distributed surface wind-field simulations. Topographic influences on major snowmelt energy balance factors (solar radiation and turbulent fluxes of sensible and latent heat) were modelled on a small-scale (40 m) basis. A spatially variable complete snowmelt energy balance was subsequently computed and applied to the distributed snow cover, allowing the simulation of the progress of melt throughout the basin. The emerging patterns compared very well visually to snow cover observations from satellite images and aerial photographs.

Results show the relative importance of variable end-of-winter snow cover, spatially distributed melt energy fluxes, and local advection processes for the development of a patchy snow cover. This illustrates that the consideration of these processes is crucial for an accurate determination of snow-covered areas, as well as the location, timing, and amount of meltwater release from arctic catchments, and should, therefore, be included in hydrological models. Furthermore, the study shows the need for a subgrid parameterization of these factors in the land surface schemes of larger scale climate models. Copyright © 2005 Crown in the right of Canada. Published by John Wiley & Sons, Ltd.

KEY WORDS modelling; small-scale spatial variability; snowmelt; arctic; energy balance

INTRODUCTION
The transition from winter to summer is a very important event for northern ecosystems, with approximately 30 to 60% of annual precipitation being released as runoff during a relatively short period. As a consequence, spring snowmelt is often the main runoff event of the year, leading to maximum annual discharge rates and water levels (Marsh et al., 2002). The removal of the snow cover also initiates the melting of river and lake ice, the thawing of the active layer, and marks the beginning of the evaporation season. The meltwater produced raises soil moisture and initiates or increases streamflow.

The decaying snowpack causes a major change in the surface energy balance of arctic regions. Snow affects net shortwave radiation due to its high albedo, whereas net longwave and turbulent fluxes of sensible and latent heat are mainly influenced by the surface temperature limitation of snow-covered areas (SCAs) to 0 °C. In contrast, adjacent snow-free areas experience surface temperatures of up to 40 °C (Kane et al., 1991; Liston, 1995; Marsh, 1999). The difference in surface temperatures leads to significantly different fluxes of longwave...
radiation and of sensible and latent heat. In addition, turbulent exchanges between the atmosphere and the land surface are often in different directions under patchy snow cover conditions (away from the snow-free surface and towards the SCAs) (Neumann and Marsh, 1998).

Several studies have shown that arctic end-of-winter snow covers are highly variable due to the redistribution of snow by wind during blowing snow events (Pomeroy et al., 1998; Essery et al., 1999; Marsh, 1999; Liston and Sturm, 2002). The combination of such a snow cover with spatially variable energy fluxes leads to the quick development of a mosaic pattern of coexisting snow-covered and snow-free patches. The patchy snow cover results in local-scale advection of sensible heat from snow-free areas to remaining snow patches, a small-scale process that increases the magnitude and spatial variability of turbulent fluxes transferred to the snow cover, while also affecting the average energy flux over the composite landscape (Marsh et al., 1997).

Aircraft observations carried out during the Mackenzie GEWEX Study (MAGS) have shown that the spring melt period with its patchy snow cover is characterized by an especially high spatial variability in surface energy fluxes (Brown-Mitic et al., 2001). It becomes clear that melt variability plays a key role in determining SCA, as well as volume and timing of meltwater release from arctic catchments.

Although spatially variable snow covers and melt energy terms have been studied in alpine regions (Bloeschl et al., 1991; Luce et al., 1998; Marks et al., 1999), the significance of the various controlling factors for low-relief arctic tundra regions is not well known. Related studies include those of Hinzman et al. (1992), Woo and Young (2003) and Déry et al. (2004, 2005) and the advection studies of Marsh et al. (1999). As a result of the lack of previous studies, it has not been possible to consider the relative importance of these processes to arctic snowmelt or to address them properly in applicable hydrologic or land-surface models.

This study considers the small-scale variability in arctic snowmelt over tundra surfaces by combining a spatially distributed snow cover with simulations of spatial variability in radiation and turbulent fluxes of sensible and latent heat at an arctic site in northwest Canada, and then comparing the simulated snowmelt patterns to observations. The goal of the study is to understand better the relative magnitude of these factors on the development of a patchy snow cover, the surface energy balance of arctic spring landscapes, and the timing and volume of meltwater runoff from arctic basins. Larger scale hydrologic and atmospheric models are needed to assess the impact of future climate scenarios on the timing and volume of arctic spring runoff and on the exchange of energy between the snow-covered terrain and the atmosphere. Future work will be concentrating on including the small-scale processes identified in this study into larger scale hydrologic and atmospheric models in order to estimate better the meltwater runoff for current and future climate conditions.

STUDY AREA

This study was conducted during the spring of 1999 as part of the MAGS–Canadian GEWEX Enhanced Study (CAGES) programme. A digital elevation model (DEM) showing an area of 14 km × 12 km around the National Water Research Institute research basin of Trail Valley Creek (TVC) was used for the simulations. The complete DEM consisted of 105 000 grid cells with a resolution of 40 m × 40 m and was obtained by digitizing 1:50 000 National Topographic Survey maps.

TVC is located at 68°45'N, 133°30'W, and lies approximately 55 km northeast of Inuvik in the Northwest Territories. The area has fairly low relief and is characterized by gently rolling hills with some deeply incised river valleys. Elevation ranges from 40 to 187 m a.s.l., with an average elevation of 99 m a.s.l. The mean slope is 3°, with the maximum gradient reaching 33°. The region is underlain by continuous permafrost. It lies at the northern edge of the forest–tundra transition zone, with tundra vegetation dominating much of the upland areas; some shrub tundra and sparse black spruce (Picea mariana) forest can be found on hillslopes and in the valley bottoms (Neumann and Marsh, 1998).

TVC experiences extensive redistribution of snow due to blowing snow events during the winter period. Studies have shown that open tundra areas act primarily as sources of blowing snow, and riparian shrub and forest areas act as traps. In addition, snow tends to accumulate in snowdrifts on steep slopes, lake margins,
and in the actual stream channels. The land cover of the area was classified into four major categories: tundra, shrub tundra, forest and drift. The respective land cover class for individual grid cells of the DEM was determined using a midsummer Thematic Mapper image and PCI software for image classification (Marsh and Pomeroy, 1996). The resulting land cover classification map for the study area is shown in Figure 1. Drifts are defined as areas in which snow accumulates during blowing snow events mainly located on steeper leeward slopes and around lake margins. Drift areas were identified from the DEM and spring satellite images.

Snowmelt in the region usually occurs during May and June. In the spring of 1999, snowmelt started around 5 May as daytime temperature first rose above freezing point. To cover the entire snowmelt, a study period lasting from 5 May to 10 June 1999 was chosen. Land-cover-based snow surveys were conducted in the basin prior to melt to establish end-of-winter snow water equivalents (SWEs), and meteorological conditions throughout the model period were recorded at two permanent and two temporary observing stations situated in the basin. The SCA for TVC and the surrounding area was determined from three SPOT satellite images taken during the melt period, with SCA declining from 79% on 23 May to 39% on 28 May and 2.5% on 10 June. Aerial photographs showing the progress of melt in the area were taken on several days throughout the period.

METHODOLOGY

End-of-winter snow cover

The end-of-winter snow cover in arctic basins generally has an SCA of 100%, while being highly variable in its SWE due to redistribution by blowing snow events. Arctic environments are especially vulnerable to blowing snow events due to long exposed fetches and the absence of snow-cover-stabilizing freeze–melt
cycles or rain-on-snow events during the winter. Studies in TVC have shown that end-of-winter SWE in individual landscape units typically varies from 50% (of measured snowfall) in open tundra areas to 400% in snowdrift areas (Pomeroy et al., 1997). Essery et al. (1999) note that there is also considerable variation in SWE within the land cover classes, with coefficients of variation varying from 0-42 in exposed locations to 0-08 for more sheltered areas. Much of the variability in open areas can be attributed to topographic factors. During blowing snow events, snow is scoured preferentially from slopes facing the predominant winter wind directions due to the accelerated winds on these slopes, whereas flow separation in the lee of topographic features causes the cessation of transport and subsequent deposition. Kane et al. (1991) found up to 65% more snow accumulation on the leeward hillslopes of a small Alaskan watershed, even if these slopes only had a gradient of 2–3°. Liston (1995) notes that, in general, snow drifts tend to form in the same locations every winter, even in years of relatively low snow transport rates, since drifts occupy a small area compared with their source area. Once snow is trapped in a drift, very little additional sublimation occurs, whereas above-average sublimation losses are observed on windward slopes, further adding to the variability of the end-of-winter snow cover. In TVC, snowdrifts develop mostly along the steep slopes of the river valleys (slope drifts), as well as along the margins of lakes and in the actual river channels themselves (channel drifts) (Figure 1). Drift areas cover approximately 8% of TVC and approximately 5% of entire study area. Snowdrifts are crucial for the hydrology of the area, since they may hold up to 33% of the entire end-of-winter snow and can persist well into the summer (Marsh and Pomeroy, 1996).

As a result of the observed relationship between snow cover distribution and topography, for this study it was decided to subdivide exposed, open tundra and drift areas according to slope and aspect of each grid cell of the DEM in relation to the dominant northerly to westerly winter winds of 1998–99. Grid cells were designated as windswept if their simulated wind speeds (from the surface wind model (see ‘Snowmelt energy balance’ section)) were more than 5% higher than the area-wide average values, as windward if oriented between west and north and with a slope larger than 3°, and as leeward if their aspects pointed between south and east and with a slope larger than 3°. Grid cells not falling into these categories were assumed to be neutral. The windswept designation was only used for open tundra areas. Appropriate weighting factors relating SWE for the new classes to observed SWE from snow surveys were determined based on results from Essery et al. (1999). Table I shows the final land cover/topography classes with their respective area, weighting factors, and final SWE’s for each class. A similar approach was used by Woo and Young (2003) to model a spatially distributed end-of-winter snow cover for an arctic basin. The study then simulated snowmelt at a 1 km resolution by calculating a snowmelt energy balance from constant shortwave radiation, humidity, and precipitation; other meteorological variables, like temperature and wind speed, were related to elevation with empirical formulae.

<table>
<thead>
<tr>
<th>Land cover</th>
<th>Wind regime</th>
<th>Area (%)</th>
<th>Weighting factor</th>
<th>SWE (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open tundra</td>
<td>Windswept</td>
<td>9</td>
<td>0.35</td>
<td>41</td>
</tr>
<tr>
<td></td>
<td>Windward</td>
<td>10</td>
<td>0.95</td>
<td>114</td>
</tr>
<tr>
<td></td>
<td>Neutral</td>
<td>34</td>
<td>1.11</td>
<td>129</td>
</tr>
<tr>
<td></td>
<td>Leeward</td>
<td>12</td>
<td>1.23</td>
<td>143</td>
</tr>
<tr>
<td>Shrub tundra</td>
<td>Windward</td>
<td>28</td>
<td>1.00</td>
<td>197</td>
</tr>
<tr>
<td></td>
<td>Neutral</td>
<td>2</td>
<td>1.00</td>
<td>110</td>
</tr>
<tr>
<td></td>
<td>Leeward</td>
<td>1</td>
<td>1.80</td>
<td>822</td>
</tr>
<tr>
<td>Forest</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drift</td>
<td>Windward</td>
<td>2</td>
<td>0.70</td>
<td>320</td>
</tr>
<tr>
<td></td>
<td>Neutral</td>
<td>2</td>
<td>1.00</td>
<td>457</td>
</tr>
<tr>
<td></td>
<td>Leeward</td>
<td>1</td>
<td>1.80</td>
<td>822</td>
</tr>
</tbody>
</table>
Snowmelt energy balance

The surface energy balance for a continuous snow cover where local advection is not important can be written as

\[ Q_M = Q_D + Q_N + Q_H + Q_E + Q_G + Q_P \]  

(1)

where \( Q_M \) is the energy available to warm and subsequently melt the snow cover, \( Q_D \) is the initial snow cover energy deficit, \( Q_N \) is net radiation, which can be further subdivided into a shortwave (solar) and a longwave (thermal) component, \( Q_H \) and \( Q_E \) are the turbulent fluxes of sensible and latent heat, \( Q_G \) is the ground heat flux, and \( Q_P \) is the flux of heat from precipitation (Luce et al., 1998). Previous work has shown that several of these energy terms are greatly affected by the local topography, even in the relatively gentle terrain of the study area (Pohl et al., 2006a,b). An additional factor, termed local advection, is introduced as the snow cover becomes discontinuous.

Topographically induced, small-scale variability of incoming solar radiation was modelled by Pohl et al. (2006a) utilizing an approach outlined by Ranzi and Rosso (1991). The following is a brief outline of the methods used in that study. The solar model calculates hourly, theoretical clear-sky global radiation values for a horizontal surface. A cloudiness index is then computed by comparing the calculated with measured global radiation data, and is used to partition the measured global radiation into direct and diffuse components. The topographic effect on incoming direct radiation is assessed by calculating local illumination angles from the solar zenith and azimuth angles and terrain slope and aspect angles using standard methods. Diffuse radiation is assumed to be isotropic and uniformly distributed over the entire area. The model also checks each grid cell for shadowing effects, with shadowed cells receiving only diffuse radiation. The model was validated against measured direct and diffuse radiation. A time-dependent albedo decay function (Gray and Landine, 1987) was applied to model outputs to compute net solar radiation over SCAs. Albedo was reset to 0-8 after new snowfall was observed.

Incoming longwave radiation fluxes for the present study were calculated on an hourly basis using an equation introduced by Satterlund (1974). This formula has been shown to work well in conditions around 0°C (Male and Granger, 1981), which were observed through much of the melt period. The contribution of cloud cover to atmospheric longwave emission was quantified via a nonlinear function of fractional cloud cover (Kustas et al., 1994; Brutsaert, 1982). The cloudiness index calculated by the shortwave radiation model was used to obtain hourly fractional cloud cover values. Outgoing longwave radiation was calculated from the Stefan–Boltzmann equation, which relates longwave emission from a surface to the fourth power of its temperature. An emissivity of 0.985 for the snowpack was chosen (Marks and Dozier, 1992). A variable snow-surface temperature was estimated by assuming a simple relationship between snow surface and air temperature and including a term for radiative cooling of the snow surface especially at night (Jordan, 1991; Marsh and Pomeroy, 1996). Pohl et al. (2006b) note that computed snow-surface temperatures compare well to measured values.

A field experiment of wind flow around a low (maximum elevation difference 54 m) conical hill in TVC showed that there is considerable topographic influence on surface wind fields, even in the gently undulating terrain typical of much of the basin. Differences in wind speed of up to 39% were observed between the upper convex reaches of the windward slopes, including the actual hilltop, and the lower concave leeward slopes (Pohl et al., 2006b). Such topographic modifications of surface winds result in small-scale variations in the turbulent fluxes of sensible and latent heat. Pohl et al. (2006b) present a method for determining this spatial variability, which is described briefly in the following. A simple model (Liston and Sturm, 1998) was employed to simulate topographic effects on surface wind flow. The model calculates wind weighting indices from slope and curvature factors for each grid cell of the DEM in relation to observed wind directions, which are then used to distribute wind speed throughout the basin, accelerating winds on convex, windward slopes and decreasing wind speeds in concave, leeward areas. The model was able to reproduce results of the field experiment adequately, with \( R^2 \) values comparing predicted with observed wind speed ranging from 0.82 to 0.97. Hourly sensible and latent heat fluxes were calculated from data collected at the meteorological stations.
of TVC using a bulk aerodynamic approach. These were then distributed throughout the model domain using the simulated wind fields. Based on observations at four different locations in the basin and the low elevation differences within the basin, air temperature and relative humidity were assumed to be constant over the study area.

Ground heat flux was measured with heat flux plates at three locations within the study area. The measured values were fairly small, averaging less than 4 W m$^{-2}$, as long as the locations were snow covered; this is very close to values reported in the literature (Pomeroy et al., 1998). Since this ground heat flux was very small compared with other melt energy factors, it was disregarded in this study in the surface energy balance calculations of snow-covered ground. Furthermore, only trace amounts of precipitation (mainly falling as snow) were observed during the study period and $Q_P$ was, therefore, not considered in the snowmelt energy balance simulations.

Numerous studies have shown that an additional energy term is introduced into the snowmelt energy balance once the snow cover has become discontinuous, as sensible heat is transferred horizontally from snow-free areas to the remaining snow patches (Liston, 1995; Marsh et al., 1997; Shook and Gray, 1997). This energy term was included in the snowmelt energy calculations of this study as soon as the snow cover became discontinuous. Neumann and Marsh (1998) introduced an advection efficiency term $F_s$ that quantifies the fraction of sensible heat originating from bare ground areas that is actually advected to adjacent snow patches. For the present study, a regression equation relating $F_s$ to percentage snow-free area was determined from a combination of field measurements (Neumann and Marsh 1998) and model data (Liston, 1995; Marsh et al., 1999) (Figure 2). Model data included as field measurements from Neumann and Marsh (1998) were only available up to 88% snow-free area.

To calculate the local advection the sensible heat flux over snow-free ground had to be calculated. This was computed as the residual of the bare ground energy balance (Marsh and Pomeroy, 1996). Net radiation from a meteorological site was linked to incoming solar radiation measured at the same site and found to be highly correlated ($R^2 = 0.94$) once the underlying snow cover had melted. Since the meteorological site did not become snow free until well into the melt period, this relationship was used to calculate net radiation for bare-ground locations up to that point. Latent heat flux was computed using the Priestley–Taylor method, which has been applied successfully in a variety of northern environments (e.g. Mendez et al., 1998; McFadden et al., 1998). Marsh et al. (1994) showed that the Priestley–Taylor evaporation values were within 5% of those derived from the water balance method for a sub-basin of TVC. An $\alpha$ (evaporation coefficient relating potential to actual evaporation in the Priestley–Taylor equation) of 0.69 was determined for the study area from lysimeter measurements in a previous study (Neumann and Marsh, 1998). Ground heat flux for

![Figure 2. Advection efficiency versus snow-free area. Simulated data are from Liston (1995) and Marsh et al. (1999)](image-url)
snow-free areas was estimated as a constant portion (18%) of net radiation (Eaton et al., 2001). The extent of snow-free areas was determined on a daily basis and the results were used to obtain the appropriate F_s for that day from the regression curve in Figure 2. Subsequently, residual sensible heat fluxes from the bare-ground energy balance were multiplied by the respective F_s and the resulting advected energy was uniformly distributed over the remaining snow patches.

RESULTS

Distributed snow cover

In order to map variations in SWE for the end-of-winter snow cover 1999, land-cover-based snow surveys were conducted in April 1999. The surveys showed an average SWE of 117 mm for open tundra, 110 mm for forest, 197 mm for high shrub vegetation, and 457 mm for drift areas. Within-land-cover variations were similar to those noted by Pomeroy et al. (1998) and reproduced by Essery et al. (1999) for the same study area (Table II). These SWEs were multiplied by the respective weighting factors for the landscape classes (Table I), resulting in SWEs ranging from 41 mm for windswept open tundra areas to 822 mm for leeward drift locations. Figure 3 shows the simulated spatially distributed snow cover over the study region. The lowest SWE can be seen on exposed west- to northwest-facing tundra slopes (e.g. in the northwestern part of the study area), whereas snowdrifts are evident primarily along the steep slopes of the river valleys, in the stream channels, and on lake margins. The basin-wide mean of simulated SWE was 159 mm. It should be noted that the snow cover distribution in Figure 3 is based solely on the nine identified landscape classes, each of which has a uniform SWE. It can be expected that the SWE within these classes varies, making the actual snow cover even more variable than the simulated one.

Distributed snowmelt energy balance

The snowmelt energy balance was determined from the simulated spatially distributed net solar radiation and turbulent fluxes combined with the spatially uniform longwave and local advection values on an hourly basis for 37 days from 5 May to 10 June 1999. Figure 4 shows the area-wide daily averages of the snowmelt energy terms. In the melt period, 6 of the first 8 days showed positive average melt energy balance values, followed by a cold spell of 5 days with negative melt energy balance values (14–18 May). The bulk of the snowmelt occurred after that period. Table III lists accumulated energy balance factors over the model period, including some information on the simulated variability for solar radiation and turbulent fluxes. Note that positive values indicate a gain of energy by the snow cover.

The spatially distributed hourly values of snowmelt energy were accumulated to obtain daily values, as well as values for the entire study period. An analysis of the daily maps shows that maximum overall values of

<table>
<thead>
<tr>
<th>Vegetation class</th>
<th>Coefficient of variation</th>
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</thead>
<tbody>
<tr>
<td>Pomeroy et al.</td>
<td>Snow survey</td>
</tr>
<tr>
<td>(1998)</td>
<td>1999</td>
</tr>
<tr>
<td>Open tundra</td>
<td>0.31</td>
</tr>
<tr>
<td>Shrub tundra</td>
<td>0.19</td>
</tr>
<tr>
<td>Sparse forest</td>
<td>0.16</td>
</tr>
<tr>
<td>Drift</td>
<td>0.34</td>
</tr>
</tbody>
</table>
Figure 3. Distributed TVC end-of-winter snow cover for 1999

Figure 4. Accumulated daily snowmelt energy balance averaged over study area

Snowmelt energy were observed on clear, warm days with strong southerly winds. The amount of variability present in the energy balance depended mainly on the amount of cloud cover and on wind direction. Increasing cloud cover tends to decrease variability, as incoming shortwave radiation becomes more diffuse and, therefore, less variable. Southerly winds increased the variability of snowmelt energy, since, under those conditions, areas
Table III. Accumulated totals of snowmelt energy balance components for entire melt period

<table>
<thead>
<tr>
<th>Component</th>
<th>Mean (MJ m(^{-2}))</th>
<th>Range (MJ m(^{-2}))</th>
<th>S (MJ m(^{-2}))</th>
<th>2S (MJ m(^{-2}))</th>
<th>2S (mm SWE)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Net solar radiation</td>
<td>195</td>
<td>52.8</td>
<td>2.8</td>
<td>11.1</td>
<td>33</td>
</tr>
<tr>
<td>Net longwave radiation</td>
<td>−135</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Turbulent flux</td>
<td>79</td>
<td>71.8</td>
<td>2.1</td>
<td>8.4</td>
<td>25</td>
</tr>
<tr>
<td>Local advection</td>
<td>55</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Melt energy balance</td>
<td>−194</td>
<td>−102.4</td>
<td>−2.7</td>
<td>−10.8</td>
<td>−32</td>
</tr>
</tbody>
</table>

of above- and below-average values of solar and turbulent melt energy fluxes are co-located on south- and north-facing slopes respectively. Days with persistent northerly winds show considerably decreased variability, as the topographical influences on solar radiation and turbulent fluxes counteract each other. Figure 5a and b shows the distributed energy balance of the study area for two consecutive days of the melt period with clear skies and fairly similar melt rates. The main difference was that southerly winds dominated on 23 May and winds came primarily from northerly directions on 24 May. This resulted in large differences of variability simulated for the study area, with the standard deviation \( S \) on 23 May (0.28 MJ m\(^{-2}\)) being almost three times as high as on 24 May (0.11 MJ m\(^{-2}\)). The differences in variability are also evident in the histograms of distributed energy balance values over the study area for the 2 days (Figure 6a and b).

Figure 7 shows the accumulated distributed snowmelt energy balance for the entire study period. It should be noted that most of the area did become snow free at some point during the modelling period; therefore, this map is not a representation of actual surface energy balance values, but is an estimate of maximum potential amounts of melt. The overall variability of the snowmelt energy balance was decreased due to the predominantly northerly spring winds in the study area (similar to 24 May; see Figure 5b). This could explain why the assumption of uniform melt rates over longer time steps may adequately reproduce average basin-wide melt rates and snow cover disappearance, as shown in previous studies in the area (Marsh and Pomeroy, 1996). The results of this study, however, suggest that considering small-scale variabilities is crucial when simulating snowmelt and surface atmosphere exchange processes at higher spatial and temporal resolutions.

Over the entire period, an area-wide average of 194 MJ m\(^{-2}\) was available for snowmelt, with \( S = 2.7 \text{ MJ m}^{-2} \) (Table III). South-facing slopes showed above-average energy totals, since solar radiation contributed more energy than turbulent fluxes to the overall melt energy balance, whereas below-average melt energy values are simulated for north-facing slopes and especially along the bottoms of the river valleys. Here, a positive interaction between solar radiation and turbulent fluxes could be observed, as valley bottoms receive less solar radiation due to frequent shadowing at lower solar elevation angles, and below-average turbulent fluxes as a result of decreased wind speeds in the concave river valleys. To assess the importance of the simulated melt energy variability better, it is useful to convert it to differences in resulting potential snowmelt amounts, assuming that all the available energy was used for snowmelt after the snowpack had reached an isothermal state. The simulation shows that in 1999 most of the area’s snow cover became isothermal on the second day of the study period (6 May), supporting this assumption. The relatively flat upland tundra areas of TVC show potential snowmelt values within one standard deviation, with a maximum difference in melt of about 8 mm. Most of the steeper slopes along the river valley and on the prominent hills are much closer to the \( 2S \) around the mean, resulting in differences of potential snowmelt amounts of as much as 32 mm SWE. Considering the usually low end-of-winter SWE for open tundra areas (typically 50 to 120 mm), it is evident that the simulated variability in the energy terms plays an important role in the melting of the snow cover and the formation of a patchy snow cover.

Figure 8 shows a north–south valley cross-section located near the basin outlet. The snow cover along this transect, showing a typical, average open tundra snowpack, was obtained from snow survey data. Snowdrifts can be seen in the actual river channel and on the upper reaches of the north-facing slope. Figure 8 also shows
continuous three-grid-cell means of spatially variable snowmelt energy balance fluxes across the transect expressed as difference from the basin-wide mean values (see Table III). As expected, solar radiation shows maximum contributions to snowmelt on the steepest parts of the south-facing slope, and turbulent fluxes show maximum values on the north-facing slope. The below-average energy balance values in the valley bottom.
simulated in other parts of the basin are not as evident in this cross-section, since the valley is much wider than the narrow river valleys in the upstream portions of the basin. The potential snowmelt values differ by as much as 31 mm SWE between the south- and the north-facing slopes.

**Snowmelt simulation**

The model simulation, combining the variable snow cover with the distributed energy balance, was initiated on 5 May. Snow temperature, determined from snow pit data, was −11°C and the snow cover cold content for each grid cell of the DEM was determined from this temperature, the SWE, and the specific heat of ice (Gray and Landine, 1988; Tarboton et al., 1995). The melt energy obtained was initially used to bring the snow cover to an isothermal state at 0°C. Any additional positive surface energy was used to melt the snow, whereas negative surface energy balance values resulted in the generation of a new energy deficit and a lowering of the snow temperature below 0°C. This new energy deficit had to be overcome before additional melt was simulated for the particular location. A large portion (approximately 90%) of the snow cover in the study area became isothermal for the first time on 6 May. This day had a large, positive melt energy with an area-wide average of 7.3 MJ m⁻² (see Figure 4). Most of the deeper snowpack in drift areas, however, did not become isothermal until 10 May, with some drift areas remaining below 0°C as late as 22 May.

Figure 9 illustrates the simulated progress of the melt throughout the study area and the respective satellite and aircraft observations. It becomes evident that the model captures the emerging patterns of snow-covered and snow-free areas very well. The first areas to become bare are mainly located on northwest- to southwest-facing slopes due to the combination of eroded snow cover and above-average melt energy fluxes at those locations. Previous studies (Hinzman et al., 1992; Pohl et al., 2006a) have shown that west-facing slopes have higher melt rates as a result of receiving maximum solar radiation during the afternoon hours, coinciding with the highest air temperatures and a ripe snow cover that has recovered from any energy deficit that might have
be incurred during the previous night. The model predicted that the bulk of the open upland tundra areas would become snow free between 27 and 29 May, as was observed from aerial photography (Figure 9).
Figure 9. Comparison of modelled melt patterns with satellite images and aerial photographs.
Most of the north-facing open tundra slopes and shrub tundra areas were predicted to become snow free over the next 2 days (30–31 May). Satellite images indicate that shrub tundra areas showed a higher variability than was indicated by the model, with most of the shrub tundra areas melting prior to or simultaneously with open tundra sites. This discrepancy was not unexpected, since the model was mostly set up for open, vegetation-free areas (especially through assumptions made about albedo and surface roughness) that were not corrected for vegetation influence. Furthermore, the end-of-winter snow cover of shrub tundra areas was not further subdivided, since there seemed to be no physical foundations for such a subdivision, as the snowpack properties of vegetated areas have been shown to be largely independent of wind directions and topography (Essery, 2001). Clearly, more research concerning distributed snow deposition and distributed energy balance factors in this important tundra land cover class is needed.

By 3 June the simulation showed that only drift areas along the sides of the river valley, around the lake margins, and in the actual channels itself remained snow covered. This model prediction is validated by aerial photographs for that day. These remaining snow patches can be attributed to a combination of higher than average end-of-winter SWE in the designated channel and lake drift areas (see ‘Distributed snow cover’ section) and on below-average snowmelt energy amounts, especially on steep, north-facing slopes and in the bottom of the river valleys (see ‘Distributed snowmelt energy balance’ section).

Kane et al. (1991) note that the snow remaining in the actual river channels can play an important hydrological role, causing a lag of 1–3 days in meltwater runoff due to snow damming. Until the end of the study period on 10 June, most drifts in river channels and along western valley slopes were predicted to disappear, leaving only late-lying snowdrifts mainly on southerly, easterly, and some northerly slopes. The same pattern can be observed on the satellite image of the area for that day (Figure 9). Late-lying snowdrifts have a considerable impact on the late spring to early summer runoff of northern rivers (Marsh and Woo, 1981) and rivers in more temperate regions (Luce et al., 1998). Melting of late-lying snowfields also affects the runoff-producing processes in the area, since studies have shown that hydrological conductivities and, therefore, response times drop dramatically as the water table subsides into the lower peat layer of the permafrost soil (Quinton and Marsh, 1999). Meltwater originating from the late-lying snowdrifts keeps the water table down slope from their locations near the surface and, therefore, hydrologically connected to the stream network. It becomes evident that identifying the correct location and extent of late-lying snowfields is crucial for hydrological modelling of meltwater runoff from arctic catchments.

To compare further the simulated and observed snowmelt patterns obtained, a geospatial analysis was performed using the geographical information system package PCI (PCI Geomatics Inc.). For the analysis, one cover type had to be considered the surface matrix while the other was designated as the patch type. Snow-free patches (for 23 May) and snow-covered patches (28 May and 11 June) from the simulated snow cover maps and from the satellite images were converted to polygons and their spatial statistics were calculated (Table IV). The analysis shows that the simulated patches exhibited a larger average area and a longer average perimeter length. These differences indicate that, although the overall SCA was predicted fairly accurately, the simulated remaining snow patches tended to be more continuous and the observations showed a more fragmented snow cover. This can be partly attributed to the larger resolution of the satellite image (20 m) compared with the

| Table IV. Geospatial statistics for the snowmelt patterns obtained |
|------------------------|------------------------|------------------------|------------------------|
|                        | Modelled               | Observed               |                        |
|                        | 23 May   | 28 May   | 11 June  | 23 May    | 28 May    | 11 June    |
| SCA (%)                | 91       | 37-8     | 2-2      | 79        | 38        | 2-5        |
| Patch type             | Snow free | Snow     | Snow     | Snow free | Snow     | Snow       |
| Average patch area (m²)  | 33 620  | 80 500  | 90 500  | 20 400    | 51 600    | 60 20      |
| Average perimeter length (m) | 736      | 1335     | 366      | 627       | 1160      | 325        |
model (40 m). However, it also indicates that, although the model simulation captures much of the small-scale variability, natural conditions are still more variable, leading to the observed more fragmented, smaller snow patches. As discussed in the ‘Distributed snowmelt energy balance’ section, the natural spatial variability in the end-of-winter SWE, in particular, is likely to be higher than in the simulated snow cover, as uniform SWE values were assumed within the nine land cover classes identified in the model. The model resolution of 40 m was largely dictated by the resolution of the DEM available for the study area. A larger resolution might be advantageous, owing to the high heterogeneity of arctic snowmelt processes. However, these advantages have to be weighed against the vastly decreased computational efficiency, particularly in the solar radiation and surface wind flow model. Future studies will attempt using a higher resolution simulation.

Figure 10 shows the snow cover depletion curves (SDCs) for a variety of scenarios, from the most complex (combining a spatially variable snow cover in conjunction with a distributed energy balance used in this study) to the most simple (uniform snowpack, uniform melt energy), still employed by some land surface schemes of larger scale climate and weather prediction models. Also shown are the SCAs determined from the three satellite images available for the study area over the spring of 1999.

A comparison of simulated and observed SCAs shows that the model slightly underestimates the snow-free area early in the melt season, indicating that the area of extremely eroded snow cover (the windswept tundra land class) is probably slightly larger than assumed in the present study. The other two SCAs for the middle and latter stages of the melt are very well predicted by the model. The SDC produced by a spatially distributed snow cover and a uniform melt seems to follow the fully distributed SDC fairly closely, even though the first snow-free areas in this scenario appear with a 2-day delay. A uniform snow cover (the basin-wide mean of 159 mm was used) combined with a variable melt energy balance would melt within 6 days, compared with the minimum 31 days (not all snow has melted by the end of the model period) predicted by the fully distributed simulation. As expected, the uniform snow cover in combination with a uniform energy balance does a very poor job in replicating the simulated SDC, leading to potentially large errors if those SCAs are used to compute the surface energy balance for the area.

Local advection proved to be a very important factor for the snowmelt energy balance, contributing about 28% to the overall melt energy. Figure 4 shows that it was often the most important melt energy balance factor in the latter stages of melt, a phenomenon also documented by Marsh and Pomeroy (1996). Figure 11 shows that including local advection in the melt energy balance improves the comparison of simulated and observed values of SCA. It also indicates that considerable changes in SCA of up to 20% (30 May to 3 June)

![Graph](image-url)
can result from local advection. Local advection contributed as much as 27 mm SWE to daily melt rates of the study area, with an average of 7.5 mm (20 May to 10 June).

Hourly snowmelt amounts were determined by identifying areas that remained snow covered and averaging the snowmelt energy values for those areas. Overall, the simulated energy balance melted 155 mm SWE from the study area. This is very close to the 159 mm that was determined as average end-of-winter snow cover for the area, especially when considering that some snow remained in drift areas at the end of the model period. The first significant melt in the area was simulated for 9 May, with a total daily melt of 9.8 mm SWE. The first runoff at the TVC outlet, however, was not recorded until 21 May, indicating a lag time of about 12 days between the onset of melt and the initiation of runoff. This delay can be attributed mainly to meltwater percolation through the snow cover, to infiltration into the frozen soil at the base of the snow cover, and to the formation of a basal ice layer at the ground surface (Kane et al., 1991; Marsh and Woo, 1984). Marsh and Pomeroy (1996) simulated a lag of 6–9 days due to meltwater percolation for a similar tundra snow cover in TVC for the spring of 1993. The lag time of the present study was likely somewhat increased by the 5-day cold spell (14–18 May) that led to the refreezing of liquid water in the snow cover. Maximum daily melt rates were simulated between 23 and 29 May, and maximum runoff was measured from 29 May to 4 June, representing a delay of about 6 days between modelled peak melt and measured runoff values for TVC.

Overall, a cumulative runoff at the end of the modelling period (11 June) of 52 mm was observed at the TVC outlet, compared with a simulated melt of 155 mm. These numbers are very close to values reported (SWE 144 mm, cumulative runoff 54 mm) for the spring of 1994 (Marsh et al., 2002). That study showed that the remaining meltwater went into basin storage and was gradually released to the stream or evaporated over the summer period.

The results of this study should be applicable for a wide variety of snow-covered open landscapes, where blowing snow processes and spatially variable snowmelt are important. Such areas include arctic tundra, grasslands, and prairie environments.

**DISCUSSION**

Detailed hydrological models simulating the percolation of meltwater through the snow cover to the ground surface, and eventually on to the stream network, rely heavily on short-term, daily or even hourly melt rates.
Marsh and Woo (1984) developed a model to compute the advance of a faster moving meltwater finger front and a slower moving background front through a cold snow cover. The progress of these wetting fronts can be used to determine areas within catchments that are not, partially (only the finger front has reached the ground), or fully (both fronts have reached the ground surface) contributing to runoff (Marsh and Pomeroy, 1996). The model needs surface melt rates and snow depth as inputs. This study shows that small-scale variability causes considerable differences in short-term melt rates across the basin. The biggest relative variability in daily melt rates occurred on 25 May, when areas within 2s of the mean mostly located on opposite sides of the river valleys differ in their melt by up to 61% of the basin-wide mean (3 mm SWE). Absolute variability reached a maximum on 6 June, when strong southerly winds coincided with warm temperatures and relatively clear skies, creating a 2s range in daily melt energy of 1.9 MJ m⁻², or about 6 mm of melt.

The variable snow cover is important, as it dictates the snow depth through which the wetting fronts have to penetrate and, therefore, it affects the lag time between the initiation of melt at the top of the snow cover and the release of the meltwater produced at the base of the snowpack. Furthermore, the model simulations show that the variable SWE in the study area leads to considerable differences in the timing of snow cover ripening. A difference of 16 days was simulated between the snow covers in tundra, forest, and shrub tundra areas (which mostly became isothermal on 6 May) and the snowpack in drift areas (which remained below 0 °C as late as 22 May). It becomes evident that only a combination of variable snow cover and variable energy balance terms ensures an accurate simulation of meltwater percolation through the snowpack and subsequent magnitude and timing of meltwater release from a basin at any particular point in time during the melt period. This is especially important for hydrological runoff models working on the principle of variable-runoff source areas.

Observations within TVC have shown that the first melt usually occurs on steep west- to southwest-facing valley slopes in the middle part of the basin. The present study shows that this can be attributed to the combination of an eroded snow cover with above-average melt energy terms simulated for west to southwesterly slopes. Owing to the proximity of these locations to the main channel, this is often the area where sufficient meltwater accumulates in the channel to initiate streamflow. As shown in Figure 9, these areas are also among the first to become snow free and, therefore, cease to contribute to the meltwater runoff. If uniform values of snow cover and melt energy were used, then it would be difficult to simulate the location or timing of streamflow initiation.

Figure 10 seems to indicate that the distributed end-of-winter snow cover is relatively more important when trying to predict SCA accurately throughout the melt period. Much of this can be attributed to the properties of the study area, combining an extremely variable snow cover with a relatively gentle terrain that somewhat limits the spatial variability of the energy fluxes. Additionally, as was discussed before, the predominant northerly winds during the spring melt period of 1999 reduced the variability of the overall energy balance even further, as areas of above-average solar radiation received less turbulent fluxes, and vice versa. It can be expected that the distributed energy balance will have a much greater impact on SCA in more mountainous regions or in areas where the dominating spring wind directions lead to a positive feedback between the variable solar radiation and turbulent fluxes (southerly winds for basins in the Northern Hemisphere). Such a scenario was described by Pomeroy et al. (2003), who observed a large spatial variability in melt energy in a subarctic mountain catchment dominated by southerly spring winds.

Regional climate and numerical weather prediction models often operate at scales that would treat the model area of this study as a single grid cell. Furthermore, the snow component of land surface schemes used by these models is often fairly simple (Lynch-Stieglitz, 1994; Marshall and Oglesby, 1994; Essery, 1997). Figure 10 indicates that the use of a uniform snow cover and a uniform energy balance will lead to considerable errors in the computed SCA over much of the melt period, resulting in inaccurate surface energy balance values for open, northern environments. Considering the vast areas of circumpolar regions that exhibit prolonged (often up to 4 weeks) patchy snowmelt conditions, these errors in the surface energy balance terms seriously degrade the ability of these models to simulate atmospheric conditions. At the very least, a subgrid parameterization of SCA should be included. A common approach is to determine the energy...
balance and areal coverage of snow-covered and snow-free areas separately and, subsequently, calculate a weighted average for the full grid box (Claussen, 1991). This ‘tile model approach’ improves simulations, but it is unable to account for processes at the edges of the separate land classes (Essery, 1997), meaning that the process of local advection cannot be considered by this approach. The results of this study show that small-scale local advection plays a very important role in arctic snowmelt and affects the resulting SCAs strongly. Studies have also indicated that local advection affects the overall surface energy balance of the composite landscape (Lhomme et al., 1994). Land surface schemes should, therefore, be modified to include algorithms quantifying the contributions of local advection to the snowmelt energy balance.

Snow cover and snowmelt energy heterogeneity can be handled using a wide variety of methods. A study by Déry et al. (2005) uses snow-cover areal depletion curves derived from satellite observations to distribute the SWE in a study area in Alaska. The depletion curves are used to compute fractional areas of shallow and deep snow covers. A blowing snow model is then used to calculate precipitation factors and subsequently SWE for the shallow and deep snow packs (Déry et al., 2004). Despite its simplicity, the method has some limitations. As it relies on visible and infrared satellite observations, frequent episodes of low-level clouds may reduce the temporal resolution and, therefore, the accuracy of the snow areal depletion curves. Late-season snowfalls and exposed vegetation further complicate the derivation of these depletion curves (Déry et al., 2005). Furthermore, the resolution of the satellite images has a lower limit of 500 m. The areal fraction of SCA below this resolution has to be computed from an empirical equation. The method produces a snow cover with two distinct snow cover classes, shallow and deep, compared with the nine land cover classes identified in the current study. However, the approach of distributing the snow cover from remotely sensed data can be easily automated and, therefore, applied to a multitude of snow-covered basins, whereas the method used in our study is very work intensive, as land-cover-based end-of-winter snow surveys have to be conducted in every study area. It seems that the method presented by Déry et al. (2004, 2005) is better suited for larger scale studies involving multiple basins, whereas the method presented in this paper might be favourable for smaller scale simulations and process-oriented studies. Furthermore, by including vegetation, the approach used in this paper can be applied better to studies of the impact of climate change on hydrologic and atmospheric processes in arctic tundra regions, especially since an advance of shrub tundra is predicted as a result of a warmer climate (Liston et al., 2002).

In a comparative study, Déry et al. (2004, 2005) combine the distributed snow cover with a uniform and a distributed snowmelt energy balance to simulate the progress of melt and the runoff of the meltwater. In contrast to this paper, the distributed energy balance does not account for topographic influences on the surface wind field and the related turbulent fluxes of sensible and latent heat, and no local advection was considered.

**CONCLUSIONS**

The present study shows that, even in the relatively gentle terrain of the arctic coastal plains, topographic influences on solar radiation and wind speed lead to considerable small-scale variability in the snowmelt energy balance. The amount of variability depended mainly on cloud cover and wind direction. The use of evenly distributed snowmelt energy balance values would lead to substantial errors in determining the rate of snowmelt in different parts of the basin and, consequently, in the calculation of the surface energy balance of the composite landscape. Differences of around 11 MJ m\(^{-2}\), or 32 mm of potential snowmelt, were found between north- and south-facing slopes, depending on their steepness. The simulation results suggest that the variability could be even greater in more mountainous environments or regions where northerly winds are not dominant during the spring. The study also shows that the variable end-of-winter snow cover in arctic regions plays a crucial role in determining the emerging patterns of a patchy snowpack and the decrease in SCA throughout the melt period. Compared with a calculation using an evenly distributed snow cover and snowmelt energy balance terms, the first areas of bare ground appear 18 days earlier in the small-scale simulation, whereas some snowdrifts persist at least 13 days longer.
Consequently, the study demonstrates the need to simulate the spring snowmelt of arctic basins at small spatial and temporal scales using a variable end-of-winter snow cover in conjunction with spatially distributed energy balance terms, in order to predict accurately the snow-covered and snow-free areas, the timing and amount of meltwater release from different parts of the basin, and the overall surface energy balance for the composite landscape.

Additional research is needed in the very complex shrub tundra land class in order to understand snow accumulation patterns and spatial variability better in the melt energy terms. Future work will attempt to use the results of this study to develop simple relationships between topography, vegetation, and small-scale spatial variability in both snow cover and energy fluxes. Such relationships are required for accurate simulations of hydrological processes and land-surface–atmosphere energy exchanges needed for reliable predictions of hydrology, weather, and climate.

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