MELTWATER FLUXES AT AN ARCTIC FOREST-TUN德拉 SITE

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ABSTRACT

Models of surface energy balance and snow metamorphism are utilized to predict the energy and meltwater fluxes at an Arctic site in the forest-tundra transition zone of north-western Canada. The surface energy balance during the melt period is modelled using an hourly bulk aerodynamic approach. Once a snowcover becomes patchy, advection from the bare patches to the snow-covered areas results in a large spatial variation in basin snowmelt. In order to illustrate the importance of small-scale, horizontal advection, a simple parameterization scheme using sensible heat fluxes from snow free areas was tested. This scheme estimates the maximum horizontal advection of sensible heat from the bare patches to the snow-covered areas. Calculated melt was routed through the measured snowcover in each landscape type using a variable flow path, meltwater percolation model. This allowed the determination of the spatial variability in the timing and magnitude of meltwater release for runoff. Model results indicate that the initial release of meltwater first occurred on the shallow upland tundra sites, but meltwater release did not occur until nearly two weeks later on the deep drift snowcovers. During these early periods of melt, not all meltwater is available for runoff. Instead, there is a period when some snowpacks are only partially contributing to runoff, and the spatial variation of runoff contribution corresponds to landscape type. Comparisons of melt with and without advection suggests that advection is an important process controlling the timing of basin snowmelt.

KEY WORDS snowmelt; surface energy fluxes; patchy snowcovers; Arctic

INTRODUCTION

Snow plays an important role in the water and energy fluxes of arctic regions. During the 6–10 months that snow is stored on the ground, its high albedo strongly controls the surface energy balance. In the spring, there is a rapid and dramatic change in surface energy fluxes, with the resulting melt releasing 35–60% of annual precipitation. The resulting runoff often accounts for over 50% of the annual flow. This runoff has important implications to northern ecosystems, and it also plays a role in controlling circulation patterns in the Arctic Ocean (Aagaard and Carmack, 1989). It is critical for hydrological, weather and climate predictions to estimate properly the surface energy balance during the spring melt, and the resulting timing, volume and spatial distribution of meltwater release.

Early in the spring melt period, snowcover in arctic areas is continuous, and the surface melt rate is controlled primarily by net radiation and the turbulent exchanges of sensible and latent heat. However, since arctic snowcovers are highly variable in depth at the end of water (Woo et al., 1983), those areas with a thin snowcover quickly become snow free, while areas with deep snow retain their snowcover into early summer in the subarctic, and perennially in the high arctic (Marsh and Woo, 1981). During the period when the snowcover is patchy (Figure 1), which comprises most of the snowmelt period, the bare ground warms, owing to the low albedo, and there is a significant transfer of sensible heat from the bare ground to the overlying air. This energy is then transported downwind, where it may be transferred to the snowcover (Weisman, 1977; Liston, 1995; Shook, 1995) and results in increased melt around the edges of the snow patches. The relative importance of this horizontal advection of energy from bare areas to snow patches, and methods to estimate its magnitude, have not been well documented.
In the Arctic, meltwater typically enters into a snowcover which is many degrees below freezing. As a result, much of the initial meltwater freezes within the snowcover, with the amount of freezing determined by the snow depth and temperature, and the soil heat flux. Because of cold soil temperatures and the existence of permafrost, this soil heat flux is always negative (i.e. from snow to the soil) during the melt period, with reported magnitudes as large as 80 W/m² (Marsh and Woo, 1987). In temperate areas the ground heat flux is not as important, and as a result is often ignored in snowmelt runoff models. An additional complication in determining when meltwater reaches the base of the snowcover is that wetting fronts do not move uniformly through the snowcover. Instead, preferential flow paths or ‘flow fingers’ develop at the leading edge of the wetting fronts, allowing meltwater to move very quickly through portions of the snow, while much of the snowpack remains dry and therefore does not transport meltwater. This process plays an important role as a heterogeneous flux of latent energy within the snowpack and initiates runoff much earlier than would otherwise be expected. The objective of this paper is to demonstrate the application of physically based models of snowmelt and water flux through cold snowcovers over an Arctic basin and to demonstrate the spatial and temporal variation in meltwater release from Arctic snowcovers.

STUDY AREA AND FIELD METHODS

During the spring of 1993, field studies were carried out at Trail Valley Creek (TVC; 68°45’N, 133°30’W), located approximately 40 km north-east of Inuvik, NWT (Figure 2) in the forest–tundra transition zone (Bliss and Matveyeva, 1992). Trail Valley Creek is approximately 63 km² in area with elevations ranging from 60 to 190 m a.s.l. The climate is characterized by short, cool summers, long, cold winters and low precipitation, much of which occurs as snow (Table 1). Trail Valley Creek is in the zone of continuous permafrost (Heginbottom and Radburn, 1992), with taliks only occurring beneath lakes, active layer depths from 0.3 to 1.0 m, mean annual ground temperature of −7°C (Natural Resources Canada 1995) and maximum permafrost thickness estimated at 350 to >575 m (Natural Resources Canada, 1995). The ground surface is dominated by numerous periglacial features such as ice wedges, earth hummocks and thermokarst phenomena.
Figure 2. Map showing the major landscape classes in Trail Valley Creek, with 20 m contour interval. The inset shows the location of the Trail Valley Creek study area and the approximate location of the meteorological station.
Table I. Climatic normals for 1951–1980 in the study region (Atmospheric Environment Service, 1982a, b). Tuktoyaktuk is located on the Beaufort Sea coast (approximately 80 km north of Trail Valley Creek), while Inuvik is approximately 40 km south of Trail Valley Creek (Figure 2).

<table>
<thead>
<tr>
<th>Location</th>
<th>Mean Daily Temperature (°C)</th>
<th>Precipitation (mm)</th>
<th>Rain (mm)</th>
<th>Snow as a % of total precip.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Annual (D,J,F)</td>
<td>(J,J,A)</td>
<td>Annual October–April</td>
<td>May–September</td>
</tr>
<tr>
<td>Inuvik</td>
<td>-9.8</td>
<td>-28.6</td>
<td>11.5</td>
<td>266.1</td>
</tr>
<tr>
<td>Tuktoyaktuk</td>
<td>-10.9</td>
<td>-27.6</td>
<td>8.2</td>
<td>137.6</td>
</tr>
</tbody>
</table>

Table II. Percentage of Trail Valley Creek basin covered by each of the landscape types shown in Figure 2. For the purposes of snow accumulation and snowmelt estimates, landscape types were grouped into four combined landscape types.

<table>
<thead>
<tr>
<th>Landscape type</th>
<th>Area (%)</th>
<th>Combined types</th>
<th>Area (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Exposed soil</td>
<td>0.50</td>
<td>Tundra</td>
<td>69.76</td>
</tr>
<tr>
<td>Lake</td>
<td>0.85</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tundra</td>
<td>49.42</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sparse shrub</td>
<td>18.99</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Open shrub</td>
<td>20.35</td>
<td>Shrub tundra</td>
<td>21.52</td>
</tr>
<tr>
<td>Closed shrub</td>
<td>1.17</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope drift</td>
<td>5.89</td>
<td>Drift</td>
<td>8.21</td>
</tr>
<tr>
<td>Channel drift</td>
<td>2.32</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sparse forest</td>
<td>0.50</td>
<td>Sparse forest</td>
<td>0.50</td>
</tr>
<tr>
<td>Basin</td>
<td>99.99</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The landscape in this area is comprised of surfaces of water, tundra, shrub tundra, forest and zones that accumulate deep snow drifts. Landscapes classes were mapped for the Trail Valley Creek area using a digital elevation model, a midsummer Landsat Thematic Mapper image and PCI software for image classification. The red and three, near-infrared bands from the Landsat image were used for the vegetation classification. The landscape classification was compared to aerial photographs and ground observations and found sufficiently accurate for modelling the spatial variation of snowmelt processes. The resulting landscape classification map is shown in Figure 2 for the Trail Valley Creek area, and the percentage cover of each landscape type is given in Table II. Tundra areas are those with a plant cover dominated by lichens, mosses, herbs and low shrubs. Shrub tundra areas are those dominated by deciduous shrubs (Alnus, Betula, Salix) from 0.5 to 3 m in height, with the ground cover varying from approximately 10–25% for sparse shrub, 25–60% for open shrub and >60% for closed shrub (Strong et al., 1990). Sparse forest are those areas with coniferous trees (Picea) covering 25–60% of the area. Drift areas are defined as those areas with a slope greater than 9°, plus stream channels and lake shorelines. Also shown are lakes and areas of exposed soil. For the purpose of snow process modelling, these landscape types were combined to form four landscape classes: tundra, shrub tundra, drift areas and sparse forest (Table II). Given the dominance of the tundra, shrub tundra and drift areas, covering over 99% of the total basin area, this study will only present results from these areas. No attempt is made to estimate surface energy balance, melt or runoff from the sparse forest areas.

The methodology for measuring snow water equivalent distribution is described by Pomeroy et al.
P. MARSH AND J. W. POMEROY (1993). Aerial photos of the basin were obtained on 17 May (JD 137), 22 May (JD 142), 25 May (JD 145), 30 May (JD 150) and 3 June (JD 154). The snow-covered areas was obtained by scanning these photos and selecting the snow-covered area using computer software. The snow-covered area between these dates was estimated assuming a linear change in snow-covered area with time, and based on field observations that the basin was virtually snow free by 23 June (JD 173).

Data from a meteorological station installed in the Trail Valley Creek basin were used for estimating the surface energy balance for snow-covered surface early in the melt and snow free areas later in the melt period. This station is located on a tundra plateau in the lower portion of the basin (Figure 1 and 2) (elevation of approximately 85 m a.s.l.). The parameters measured and recorded half hourly by a Campbell 21x recorder include: air temperature, relative humidity, wind speed and direction, net radiation, incoming and outgoing solar radiation, air pressure and temperature at two levels within the snowpack and at three levels within the soil down to a depth of 0.75 m. Wind speed was measured by an NRG type 40 anemometer field calibrated against a Qualimetrics micro-response 2032 anemometer, which had been calibrated in a wind tunnel. Net and solar radiation were measured by a Middleton CN-2 net radiometer and Epply black and white model 8-48 pyranometers, respectively, while a Setra barometer measured air pressure. Once the site was snow free, the depth to the frost table was measured every few days. Late winter snow depth at this site was approximately 0.4 m and, as a result, the site became snow free relatively early in the melt period (28 May, JD 148). To supplement this record, a portable meteorological station was installed over a deeper snow patch (0.6 m) which retained its snowcover until approximately 7 June (JD 158). At this site, air temperature, relative humidity and net radiation were recorded by a Campbell 21x recorder. Air temperature was measured with a 76 μm diameter thermocouple to avoid solar heating (Blanford and Gay, 1992) and relative humidity was estimated from an aspirated Vaisala HMP35CF capacity sensor. Calibration showed that it was within the error limits of ±3% suggested by the manufacturer.

Owing to instrumentation problems and difficulties in measuring radiation over bare areas early in the melt period, and over snow later in the melt period, continuous records of net radiation over both snow and bare ground are not available. To fill the periods of missing data, net radiation \( Q_* \) was estimated from

\[
Q_{ss} = (1 - \alpha_s)K_D + [Q_{sB} - (1 - \alpha_B)K_D]
\]

\[
Q_{sB} = (1 - \alpha_B)K_D + [Q_{sS} - (1 - \alpha_S)K_D]
\]

where \( K_D \) is the incoming short-wave radiation and \( \alpha \) is the surface albedo. Subscripts \( S \) and \( B \) refer to snow and bare ground surfaces, respectively. Measured average albedos were 0.7 and 0.2 for snow and bare ground areas, respectively. The relatively low snow albedo is due to the impact of tundra vegetation sticking through the shallow snowcover. Comparison of predicted \( Q_{ss} \) and \( Q_{sB} \) with observed values shows that the predicted closely follows the observed values with \( R^2 \) of 0.94 and 0.99 for snow and bare ground, respectively.

Water flux through the snowpack was measured daily with a 0.25 m² meltwater collection lysimeter, and hourly with a 1.0 m² meltwater collection lysimeter. In addition, a multi-compartment lysimeter was used to determine the variability of flow over small areas. These lysimeters were located near the base of a deep drift (1.7 m) approximately 100 m north of the meteorological station in Trail Valley Creek. As a result of refreezing of meltwater within the snowpack, accurate data on meltwater flow could only be obtained after the entire snowpack had been wetted. At this site, this occurred on 30 May (JD 150). By 4 June (JD 155) sufficient snowmelt had occurred so that the lysimeters were too close to the snow surface to provide reliable data. Snow properties, including location of wetting fronts, grain size, snow temperature, and density were determined at snowpits located in this drift area.

**METHODOLOGY**

*Surface energy balance*

*Complete snowcover*. The surface energy balance at a point on a continuous snowcover where local advection is not important can be written as:

\[
Q_{M_S} = Q_{ss} + Q_{Hs} + Q_{Es} + Q_{R_s}
\]
where $Q_{Ms}$ is the energy available for melt, $Q_{gs}$ is the net all-wave radiation, $Q_{Hs}$ and $Q_{Es}$ are the sensible and latent heat fluxes, respectively, and $Q_{Rs}$ is the flux of heat from rain. In all cases the subscript $S$ refers to fluxes over a snowcover, and positive fluxes indicate an energy gain by the surface. This equation applies only to an isothermal surface layer. Changes in snowpack internal energy and soil heat flux are described in a subsequent section on percolation of meltwater into the snowcover. $Q_{gs}$ and $Q_{Rs}$ for the snowcover are measured directly, while a bulk transfer method (Dunne et al., 1976; Moore, 1983) is used to estimate $Q_{Hs}$ and $Q_{Es}$ for the snowcover. For these calculations, air temperature and humidity were measured at 1 m height, a surface roughness ($z_0$), as estimated from wind profiles, of 0.002 m was used and the drag coefficient was modified to account for stable and unstable conditions (Dunne et al., 1976). The snow surface vapour pressure was assumed to be saturated at the snow surface temperature. Snow surface temperature was estimated by: (a) assuming a simple relationship between snow surface and air temperature (Jordan, 1991), and (b) calculating radiative cooling of the snow surface (Jordan, 1991). The snow surface temperature ($t_s$) was then determined as:

(a) when $t_a > 0^\circ C$, $t_s = 0^\circ C$
(b) when $t_a < 0^\circ C$ and $Q_{gs} > 0$, $t_s = t_a 0.5$
(c) when $t_a < 0^\circ C$ and $Q_{gs} < 0$, then $t_s = t_a 0.5 - [(Q_{gs} t)/(C_i \rho_S d)]$

where $t_a$ is the air temperature, $t_a 0.5$ is the air temperature during the previous half hour, $Q_{gs}$ is the net radiation over the snow surface, $t$ is the time step, $C_i$ is the specific heat of ice, $\rho_S$ is the snow density and $d$ is the depth of a thin surface layer (0.015 m).

Discontinuous snowcover. When the snowcover becomes discontinuous, with patches of snow and bare ground (Figure 1), the snow free areas have a lower albedo, higher net radiation and, therefore, higher surface temperature than the remaining snowcover. The resulting sensible heat flux from the bare areas ($Q_{Hb}$) warms the overlying air. This energy is transferred downwind, where it may result in increased melt at the leading edge of the snow patches.

Equation (2) represents the energy balance of a point in a large snowpatch, unaffected by local advection; it will therefore underestimate the average melt for the entire patch. In order to estimate average melt for a patch, an advective energy term ($Q_A$) must be added to Equation (1) as follows:

$$Q_{Msp} = Q_{Ms} + Q_A$$

where $Q_{Msp}$ is the average melt for the snow patch, $Q_{Ms}$ is the melt calculated by Equation (2) for a large patch where melt is not affected by local advection and $Q_A$ is local advection of sensible heat from bare patches to snowcover.

Shook (1995) described methods to estimate the advective heat flux ($Q_A$) and its spatial variation over snow patches. In this paper a simple parameterization scheme, assuming that all, or a portion of, the sensible heat flux from the bare areas ($Q_{Hb}$) is transferred from the bare areas to the snow patches, so that:

$$Q_A A_S = Q_{Hb} A_B$$

where $A_S$ and $A_B$ are the areas of snow patches and bare areas, respectively. The energy advected (per unit snow-covered area) to the snow patches ($Q_A$) may then be given as

$$Q_A = \left[\frac{Q_{Hb} P_B}{P_S}\right] [F_S]$$

where $P_B = A_B / A_T$ and $P_S = A_S$ are the fraction of the basin snow free and snow covered, respectively, $A_T$ is the total area, $Q_{Hb}$ is the sensible heat flux from the portions of the basin which were snow free as estimated from Equation (6) and $F_S$ is a function expressing the portion of the bare area $Q_{Hb}$ which is advected to the snow patches. $F_S$ would have a value between 0 and 1 depending on the relative portion of bare versus snow covered area, wind speed, terrain roughness, upwind bare fetch length and the perimeter to area ratio of snow patches. Equation (5) shows that the total energy available to be advected (assuming constant $F_S$) increases with increasing area of bare ground.
A variety of methods are available for calculating the sensible heat flux \( (Q_{Hb}) \) per unit area of the bare patches. Aerodynamic methods would be preferable, but the lack of two levels of air temperature, or a bare ground surface temperature, precludes its use in this study. Instead, \( Q_{Hb} \) for bare areas was estimated from the residual of the energy balance for bare ground as:

\[
Q_{Hb} + e = -Q_{ea} - Q_{Ea} - Q_{Gb}
\]

where \( e \) is an error term due to inaccuracies in estimating the other variables, \( Q_{ea} \) is the net all-wave radiation, \( Q_{Gb} \) is the ground heat flux, \( Q_{Ea} \) is the latent heat flux and in all cases the subscript \( B \) refers to fluxes over a bare area free of any snow. \( Q_{Ea} \) for the bare areas is calculated from the Priestley–Taylor method (Priestley and Taylor, 1972; Rouse et al., 1987), while \( Q_{Gb} \) for the bare areas is estimated from changes in measured soil temperature plus the energy used to deepen the active layer (Rouse, 1984). The \( \alpha \) parameter relating equilibrium evaporation to actual evaporation in the Priestley–Taylor equation was estimated from nine small evaporation lysimeters (47.8 cm² in area) located in representative vegetation and covering a range of surface wetnesses. A mean \( \alpha \) of 0.6 was used. Basin water balance studies in the Trail Valley Creek basin (Marsh et al., 1995) have shown that this method accurately estimates basin evaporation.

Equations (3)–(6) will be tested to estimate maximum possible advection from bare areas to the snow patches. There are two practical problems with utilizing these equations for calculating average snow-patch melt: estimating \( Q_A \) and determining \( Q_{Mb} \) independently of the influence of local advection. Early in the melt period, the bare areas are small in size and it is likely that all \( Q_{Hb} \) from the bare areas is advected to the snow-covered area, therefore \( F_S = 1 \). At this time local advection will increase melt for only a short distance from the bare areas (Liston, 1995). Since the sensors at the meteorological station were far from major bare areas, it is likely that the estimates of \( Q_{Mb} \) were independent of the influence of local advection. As the bare areas increase in size, the total \( Q_{Hb} \) from the bare areas increases greatly, and it seems likely that a gradually decreasing portion of \( Q_{Hb} \) from the bare areas is advected to the snow patches (i.e. \( F_S \) decreases with time), with the remaining energy being used to warm a thicker layer of the atmosphere. Unfortunately, this change in \( F_S \) is not yet known. As the total energy available for local advection increases, it results in increased melt over an ever larger portion of the snow-covered area. Eventually, all of the snow covered area is influenced by local advection, and any estimation of \( Q_{Mb} \) using measurements in the lower few metres of the atmosphere will be at least partially influenced by local advection. Given that values for \( F_S \) are unknown, and in order to ensure that \( Q_{Mb} \) is relatively free from the influence of local advection, this paper will assume: (i) that \( F_S = 1 \) (all of the bare area \( Q_{Hb} \) is advected to the snowcover) and that estimated \( Q_{Mb} \) is free from the influence of local advection up until the time when the snow free area is reduced in size to only the isolated drift areas (approximately 10% of the basin), and (ii) after that time the warming of the air completely accounts for the bare ground \( Q_{Hb} \), and \( F_S \) must be relatively small, and therefore it is assumed that \( F_S = 0 \). In addition, at this time it is assumed that the estimated \( Q_{Mb} \) accounts for the influence of local advection. Considerably more research is required to account properly for variations in these factors.

This approach is anticipated to account properly for \( Q_A \) and \( Q_{Mb} \) early in the melt period, to overestimate \( Q_A \) and \( Q_{Mb} \) during the middle portions of the melt period, and to underestimate \( Q_A \) later in the melt period. As such it provides an envelope for the behaviour of the advective term during melt of Arctic snowcovers until more sophisticated methods become usable (Shook, 1995). Additional research is also required to determine the impact of snow surface temperatures of 0°C and bare ground surface temperatures greater than 0°C on turbulent mixing in the atmosphere and its impact on surface fluxes during periods with patchy snow.

**Meltwater percolation**

Percolation of surface melt through the snowpack was estimated from a variable flow path, meltwater percolation model, with the melt flux applied to mean snowcover depth and density in each landscape type. This allowed the determination of the spatial variability in the timing and magnitude of meltwater released for runoff. This percolation model, described in Marsh and Woo (1984) and Marsh (1991), parameterizes the percolation processes as follows:
Table III. Mean snow depths, densities and snow water equivalent (SWE) for each landscape type in Trail Valley Creek basin as determined from snow surveys. Note that for snow accumulation purposes, tundra includes bare areas, sparse shrubs and lakes.

<table>
<thead>
<tr>
<th>Combined landscape type</th>
<th>Area (cm²)</th>
<th>Depth (cm)</th>
<th>Snow density (kg/m³)</th>
<th>Water equivalent (mm)</th>
<th>Snow water equivalent per unit basin area (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tundra</td>
<td>0.6976</td>
<td>45</td>
<td>150</td>
<td>68</td>
<td>47</td>
</tr>
<tr>
<td>Shrub tundra</td>
<td>0.2152</td>
<td>100</td>
<td>230</td>
<td>252</td>
<td>54</td>
</tr>
<tr>
<td>Drift</td>
<td>0.0821</td>
<td>177</td>
<td>250</td>
<td>617</td>
<td>51</td>
</tr>
<tr>
<td>Sparse forest</td>
<td>0.0050</td>
<td>95</td>
<td>260</td>
<td>244</td>
<td>1</td>
</tr>
<tr>
<td>Basin</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>154</td>
</tr>
</tbody>
</table>

(1) The wetting front is idealized in two components: the background front, above which all snow is wet and isothermal at 0°C, and a finger front, which describes the deepest penetration of meltwater into the snow.

(2) The size and flow volume in the finger front is determined from measurements in the Canadian Arctic (Marsh and Woo, 1984) and in the Sierra Nevada Mountains of California (McGurk and Marsh, 1995). These studies have demonstrated that flow fingers vary in size only over a very small range of sizes, and that, on average, they cover 22% of a horizontal surface, while carrying 48% of the total flow. These values were confirmed by field measurements at Trail Valley Creek.

(3) Ice layers are allowed to grow at premelt strata boundaries, with the rate of growth controlled by the temperature gradient above and below the strata boundaries.

(4) The snow and soil temperature are linked by a finite difference scheme, which accounts for the soil heat flux.

RESULTS

The snow distribution within the TVC study area is highly variable, with shallow (25–50 cm) snowcover depths on the wind-blown tundra, 80–130 cm snowcover in the shrub tundra and 100–400 cm in drift areas. The mean snow depths, densities and water equivalent for each landscape type are given in Table III. Given these values and the distribution of landscape types within the basin, the basin mean snow water equivalent is 154 mm (Pomeroy et al., 1996). The distribution of snow water equivalent (SWE) for each landscape class is shown in Figure 3.

Snow patch

Surface fluxes and melt. The 1993 snowmelt period can be divided into two separate intervals. The first is driven by net radiation and characterized by air temperatures that fluctuated around the freezing point, and the second by turbulent transfer and air temperatures that were generally above freezing. This first period extended from 7 May (JD 127), when the air temperature briefly rose above 0°C (Figure 4) to 23 May (JD 143). The second portion of the melt period extended from 24 May (JD 144) to the complete melt of the snowcover on 22 June (JD 173).

Over the period between 7 May (JD 127) and 23 May (JD 143), the air temperature varied from −11 to 9°C, with considerable refreezing of the snow surface for most nights, and at times for a few days. Measurements of snow surface temperature are not available, but the estimated values shown in Figure 4 suggest that radiative cooling lowers the surface temperature by a maximum of 10°C below the air temperature. This is similar to the pattern of observed snow surface temperature shown in Jordan (1991). The surface energy...
Figure 3. Distribution of snow water equivalent for each landscape class, as determined from premelt surveys of snow depth and density at 59 points.

Figure 4. Air temperature, calculated snow surface temperature, net radiation over a snow patch and measured basin snow-covered area. Also shown are the calculated components of daily energy balance for a snow patch, and the resulting calculated daily melt.
Figure 5. Net radiation over bare ground and the resulting sensible, latent and ground heat fluxes for the period after the appearance of bare patches. The sensible heat flux from the bare ground is assumed to be advected to the snow patches, where it is used to increase the patch melt rate fluxes during this period are small in magnitude (Figure 4), with sensible heat and net radiation being the largest positive fluxes to the snow surface, and with the latent heat flux being negative on all days. The latent heat flux will be described in more detail in the following section. Over the 17-day period 7–23 May (JD 127–143) the daily heat balance was positive on 12 days, with melt rates typically below 10 mm/day (Figure 4), and with a total of 83.9 mm.

During the second portion of the melt period [after 23 May (JD 143)], the air temperatures increased, remaining generally above 0°C, and the net radiation continued to increase (Figure 4) in response to increasingly higher sun angles. The surface energy balance gradually became more positive, owing to increases in both sensible heat and net radiation fluxes. As a result, the melt rate increased and the snow-covered area decreased from 78% on 23 May (JD 143) to 12% on 30 May (JD 150), when, essentially, only the drift areas retained snow. As a result, total energy available for melt from advection (Figure 4) began to increase in response to sensible heat flux from the increasing snow free areas. The source of this energy is net radiation to the bare ground areas, as shown in Figure 5. Net radiation of the snow free areas was consistently positive, with values of 100–200 W/m², and was transferred to ground, latent and sensible heat. Marsh et al. (1995) validated the calculated latent heat (evaporative) flux by comparing it to evaporation lysimeters and components of the water balance for a small tributary of Trail Valley Creek. After accounting for the ground heat flux, the remainder of the net radiation over the bare areas, approximately 50–100 W/m², was used to warm the overlying air. Figure 5 shows that $Q_{\text{He}}$ is consistently negative (i.e. flux from the ground surface to the atmosphere) over the entire study period. Given the large $Q_{\text{se}}$ over bare ground, and the relatively low air temperatures, it is reasonable to expect that the surface
temperature of the bare areas is greater than the air temperature, with a resulting negative (surface to air) sensible heat flux on a daily basis over the melt period.

The sensible energy from the bare areas was available to be transferred horizontally to nearby snow patches, where it was able to increase the melt rate around the edges of the snow patches. The potential impact of this horizontally advected heat, is shown on Figure 4, where the advected energy per unit snow patch area increases rapidly from less than 50 W/m² on 23 May (JD 143) to over 500 W/m² on 30 May (JD 150) in response to increasing snow free area. Remember, however, that: (i) this heat flux represents the maximum possible advection and, as a result, is an upper limit, and (ii) the advection values shown in Figure 4 are an average for a snow patch that contains a spatial variability, with more rapid melt at edges of patches and lower melt at the downwind edge of large patches (Shook, 1995).

The potential impact of advection on the melt rate (Figure 4) is an increase in melt of between 5 mm/day on 22 May (JD 142) and 125 mm/day by 30 May (JD 150). This increase in advection over the melt period is similar to the finding of Shook (1995), who suggested that the small-scale advection was most important later in the melt period. After 30 May (JD 150) the melt rate for the isolated snow banks, as calculated from the bulk aerodynamic approach without a separate advection term, were typically between 10 and 60 mm/day (Figure 4).

Validation of these average snow patch melt rates is very difficult. During the early and middle stages of melt, settling of the snowpack makes it hard to obtain direct measurements of surface melt rates. As a result, only general indications of the accuracy of melt estimates during the early period can be provided. Observations showed that basin snow-covered area decreased from 98% on 17 May (JD 137) to 88% on 22 May (JD 142), and snow ablation lines showed that a snow pack of 32 cm depth and 54 mm snow water equivalent (SWE) was completely melted by 19 May (JD 139). Clearly a significant amount of melt occurred during this early period. If a constant snow surface temperature of 0°C is assumed, only 2.8 mm of melt is calculated over the period 7-23 May (JD 127-143). This clearly underestimates actual melt. A more realistic estimate of 83.9 mm of melt is simulated with the introduction of a variable snow surface temperature (as calculated from the lagged air temperature and radiative cooling). The introduction of this variable snow surface temperature increases the melt rate during the early melt period, since it decreases the very large negative heat fluxes during periods when the air temperatures are below 0°C. Later in the melt period, however, calculated melt rates can be compared with measured water flux from the meltwater lysimeters. Figure 6 compares daily flow from the lysimeters with calculated daily surface melt. For six of the eight days of data, calculated melt is lower than lysimeter flow. Although this may be due to spatial variations in flow resulting in over-catch by the two lysimeters, Marsh and Woo (1984) showed that for cold Arctic snowcovers, a 1 m² lysimeter was sufficiently large to provide an accurate estimate of average flow. Since over-catch is doubtful, it seems more likely that actual melt is larger than computed (without

Figure 6. Comparison of meltwater collected by the snow lysimeters to energy balance (no advection) estimates of surface melt. The observation periods vary from 12 to 24 hours.
advection). The difference, approximately 5–25 mm/day is presumed to be a result of the influence of advection. This additional melt represents approximately 4 and 21% of the available sensible heat flux from the bare areas (shown as advected energy on Figure 4) on 30 May (JD 150) or \(F_s\) of 0.05 and 0.21, respectively.

**Latent heat flux from snow-covered areas.** The energy balance calculations suggest that early in the melt period (7–25 May, JD 127–145), the latent heat flux \((Q_{eh})\) was negative, with a total sublimation of 120 mm. Such a large flux contradicts previous work which suggested that sublimation rates are small. Turbulent diffusion of water vapour from a stationary snow surface is limited by only partially turbulent flow at the snow surface, low surface roughness and the relatively small surface area of ice exposed to the wind. Wind pumping may increase these values above previous estimates, but sublimation rates are limited by the available energy. For Arctic snows, the small air temperature gradients and low rate of turbulent transfer result in the latent heat flux tending to correlate to the net radiation flux. This is analogous to cold, continuous prairie snowcovers, where Male and Granger (1979) reported daily sublimation ranging from 0.02–0.3 mm SWE/day with a mean of 0.1 mm SWE/day in premelt conditions. Most sublimation during high sun periods was counter-balanced by frost accumulation in the evening. The values reported by Male and Granger (1979) are for a higher energy environment than is normally the case for Arctic snow and should be considered an upper limit to sublimation from 'in situ' snowcovers.

After 25 May (JD 145), the surface energy balance suggests that, for 21 out of the remaining 28 days during the melt period, there was a positive latent heat flux from the snow surface (Figure 4), with a total condensation of 56 mm. Condensation probably did occur during the evening periods, but it is unlikely that condensation dominated over the entire melt period. Estimated condensation was of the order of 4% of total melt over this period.

It seems unlikely that sublimation and condensation from the snow-covered areas were as large as estimated. The most likely source of error in estimating the latent heat flux is the measurement of relative humidity. For example, an overestimation of only 5% in measuring relative humidity (only slightly outside the reported error limits of the sensor) would be sufficient to change the calculated latent flux from slightly positive to slightly negative.

**Meltwater percolation and runoff.** The availability of meltwater at the base of the snowpack is controlled by wetting front advance. During the time period before the finger front reaches the base of the snowcover, the snowpack does not contribute (NC) meltwater for runoff. Once the finger front reaches the base of the snow, then the snowpack partially contributes (PC) meltwater to runoff, with the volume of runoff dependent on the surface melt rate and the flow within the flow fingers. Finally, once the background front reaches the base of the snowcover, the snowpack fully contributes (FC) meltwater, and on a daily basis all of the surface melt is available for runoff. Note that the term runoff, as used above, only implies release of meltwater from the snowpack to the soil or as overland flow, or as storage at the base of the pack. Ongoing work is addressing the issue of lateral flow to the stream channel (Quinton and Marsh, 1995).

Comparison with field measurements at a drift site shows that the measured location of the background wetting front was similar to that estimated on 12 and 15 May (JD 132 and 135). This suggests that the estimation of surface melt during the early melt phase is reasonable. On 18 May (JD 138), the measured wetting front was similar in depth to the finger front estimated using either melt rate. It seems most likely that this is because of a field measurement error, rather than a model error. When measuring wetting fronts in the field, an attempt is made to estimate the depth at which all snow above it is wet (i.e. background wetting front). However, given the large number of flow fingers, it is possible to overestimate the depth of the background wetting front. On 23 and 24 May (JD 143 and 144), the measured wetting front was only slightly deeper than the predicted one. The slightly improved prediction of the background wetting front when compared with the measured wetting front on 25 May (JD 145), when using melt with advection (Figure 7) compared to melt without advection (Figure 7), suggests that the addition of the advected energy term simulates more realistically the snowmelt rate and therefore the movement of water through the snow pack during the period when the snow free area is still less than 40%. Up to this time it appears that \(F_s = 1\) is appropriate.

Figure 8 demonstrates that meltwater percolation controls the timing of meltwater release. Meltwater was available for infiltration to frozen soil or for runoff at the tundra on 16 May (JD 136), approximately six days after the start of melt. The background front, however, did not reach the base of the pack until...
three days later, on 18 May (JD 138). For the tundra sites, the wetting front advance and the timing of water first reaching the base of the snowpack are nearly identical for melt modelled both with and without advection. This similarity occurs because the snow free area at this time was very small, and therefore the advection term was negligible (Figure 4). Over this period, runoff from tundra sites was small (0–13 mm/day) owing to partial contribution early in the melt period, and to low melt rates later on (Figure 8). The drift areas demonstrate another extreme; the finger front did not reach the base of the snowpack until 20 May (JD 140), 10 days after the start of melt. The background front reached the base of the drift snowpack six days later. For drifts, when advection is included in the simulation, wetting fronts reach the base of the snowpack one day earlier than if advection is not included. Runoff was limited to only a portion of
MELTWATER FLUXES

Figure 8. Release of meltwater from the base of the snowpack for each landscape type and for melt without and with advection during the period up to Julian Day 151.

Figure 9. Calculated and measured basin snow-covered area. Note that the calculated snow-covered area is using both the melt rate calculated with and without advection.

the surface melt (5–45 mm/day) (Figure 8) during the period 19–25 May (JD 139–145) until the background front reached the snow base. After that time, runoff is equal to the surface melt rate. Owing to the extreme thickness of the drift snowcover, these areas continue to contribute water until late June (Figure 8). The shrub tundra meltwater generation is intermediate between the extremes of tundra and drift (Figure 7).

Basin snowcover and melt

Changes in basin snow-covered area. Changes in snow-covered area are calculated by reducing the distribution of SWE shown in Figure 3, as meltwater is released from the snowcover (Figure 8). The resulting calculated and measured basin snow-covered areas are shown in Figure 9. The two estimates of snow-covered area are similar and in most cases close to the measured values, with the inclusion of advection removing the snowcover earlier. On 22 May (JD 142) the measured snow-covered area lies...
between the two estimates, while on 30 May (JD 150) the measured value is closer to the estimate without advection. On 3 June (JD 154) all estimates are very similar, with measured closest to the estimate without advection. The largest error occurs on 25 May (JD 145), when both estimates are much lower than the measured snow-covered area. The reason for this is unclear, but its seems possible that the measured SWE underestimates the area covered by snowpacks with higher SWE. As a result, the calculated snow-covered area drops very quickly between 23 and 25 May (JD 143 and 145). When the advected energy term is used, the estimated snow-covered area reaches zero on 28 June (JD 179), six days after observations showed that snow remained in only a few, very small, isolated patches. Without advection, the model estimates that 1% of the basin was snow-covered on 30 June (JD 181).

Although not conclusive, the comparison above suggests that melt with advection overestimates actual melt slightly during the early portions of the melt period, and significantly during later portions of the melt. This is what might be expected if advection comprised a large portion of the bare ground sensible heat flux early in the melt period, and a decreasing portion of it during the remainder of melt. In other words, it suggests that $F_s$ is close to 1 early in the melt, with $F_s$ decreasing over the remainder of the melt period. Further work is required to substantiate this.

**Meltwater release from the snowpack.** Figure 10 shows daily runoff (expressed per unit basin area) from each landscape type. The calculations were determined from combining the estimates of surface melt, wetting front advance and snow-covered area within each landscape type. Although melt began on 10 May (JD 130), meltwater runoff was delayed, with the tundra areas contributing first. Runoff increased up to nearly 10 mm/day, but then ended by 24 May (JD 144) when the snowpack was completely melted. The shrub tundra contributes next with runoff also nearly up to 10 mm/day, but derived from a smaller portion of the basin (Table II). The similar maximum runoff from both tundra and shrub tundra is a result of the considerably higher melt rates during the period 24–28 May (JD 144–148) than during the period of runoff from tundra sites. The drift areas begin to contribute by 22 May (JD 142), but runoff is generally less than 5 mm/day owing to its small source area (Table II). In all but the tundra sites, runoff rates are lower if advection is not included.

Figure 11 shows the data from Figure 10 as cumulative runoff for melt both with and without advection.
Regardless of whether or not advection is included, there is a great difference in the timing and magnitude of meltwater release from each landscape type. For each landscape type, the runoff is similar in magnitude with and without advection, and in all cases is similar in size, but consistently lower than, the SWE stored in each landscape at the end of winter (Table III). The principal difference between the cases with and without advection, therefore, is not the total runoff, but the timing of that runoff, with an earlier and more rapid rise in cumulative runoff with time (Figure 11) in simulations that include advection.

Snowpack latent heat flux. At the basin scale, sublimation dominated, with a total of approximately 91 mm over the entire melt period. This value is probably high because of the overestimated sublimation early in the melt period. As snow-covered area decreased, the condensation later in the melt period was relatively small at the basin scale.

DISCUSSION

Combining model results with the map of landscape types for Trail Valley Creek (Figure 2) allows the contributing areas of meltwater runoff to be mapped on a daily basis. Figure 12 shows an example for 19 May (JD 139) when the tundra areas (70% of the basin) were fully contributing meltwater to runoff, the shrub tundra areas (22% of the basin) were partially contributing meltwater and the drift areas (8% of the basin) were not contributing any meltwater. Such daily calculations of the contributing areas throughout the melt period could be used to drive a distributed hydrological model. The importance of this large spatial and temporal variability of meltwater release has significant implications for predicting snowmelt runoff in these environments.

Table III shows that, even though each of the main landscape types varies greatly in area, each contains approximately one-third of the total SWE stored within Trail Valley Creek. This clearly demonstrates the importance of considering the spatial variations in meltwater runoff. For example, only 8% of the basin area (drifts) contains 51 mm SWE. Without accounting for this, accurate modelling of snowmelt runoff seems unlikely.

CONCLUSIONS

Surface energy balance and meltwater percolation models have been utilized to enhance our understanding of the processes controlling snowmelt runoff in the forest–tundra transition zone, and to illustrate possible
methods to deal with the large temporal and spatial variations in energy and meltwater fluxes. Two modifications were made to the standard bulk aerodynamic method to estimate the surface sensible and latent fluxes from the snow-covered areas of the basin. Firstly, a variable surface temperature procedure was used to improve melt predictions during periods when the air temperature was fluctuating around the freezing point. Secondly, a simple parameterization procedure was tested to illustrate the maximum energy that was available for advection from bare areas to snow patches. Although actual advection is difficult to verify, comparisons were made with measured water flux at the base of a deep snowpack, and with changes in snow-covered area. These comparisons suggest that advection is an important contribution to melt, with its magnitude increasing with increasing snow free area, and the portion of sensible heat flux from the bare areas that is advected to the snowpatches decreasing with time over the melt period.

Application of a melt water percolation model demonstrates the delay between the start of melt and the availability of meltwater for runoff. In addition, owing to the variability of snow depths between different
terrain types, there is also a variability in the timing of runoff from various landscape types, with runoff starting first in the tundra areas, and the contribution from drift areas being delayed. Because of the occurrence of preferential flow fingers, runoff from each landscape type may be divided into three periods: a period of no contribution to flow, a period when only a portion of the meltwater is contributing to runoff and a period where all of the surface melt is contributed.

By combining the above snowmelt and melt water percolation models with a landscape map, it was possible to distribute the meltwater release over the study basin. This is an important step in implementing distributed hydrological models in this environment.

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