APPLICATION OF A DISTRIBUTED BLOWING SNOW MODEL TO THE ARCTIC

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ABSTRACT

Transportation, sublimation and accumulation of snow dominate snow cover development in the Arctic and produce episodic high evaporative fluxes. Unfortunately, blowing snow processes are not presently incorporated in any hydrological or meteorological models. To demonstrate the application of simple algorithms that represent blowing snow processes, monthly snow accumulation, relocation and sublimation fluxes were calculated and applied in a spatially distributed manner to a 68-km^2 catchment in the low Arctic of north-western Canada. The model uses a Landsat-derived vegetation classification and a digital elevation model to segregate the basin into snow 'sources' and 'sinks'. The model then relocates snow from sources to sinks and calculates in-transit sublimation loss. The resulting annual snow accumulation in specific landscape types was compared with the result of intensive surveys of snow depth and density. On an annual basis, 28% of annual snowfall sublimated from tundra surfaces whilst 18% was transported to sink areas. Annual blowing snow transport to sink areas amounted to an additional 16% of annual snowfall to shrub–tundra and an additional 182% to drifts. For the catchment, 19·5% of annual snowfall sublimated from blowing snow, 5·8% was transported into the catchment and 86·5% accumulated on the ground. The model overestimated snow accumulation in the catchment by 6%. The application demonstrates that winter precipitation alone is insufficient to calculate snow accumulation and that blowing snow processes and landscape patterns govern the spatial distribution and total accumulation of snow water equivalent over the winter. These processes can be modelled by relatively simple algorithms, and, when distributed by landscape type over the catchment, produce reasonable estimates of snow accumulation and loss in wind-swept regions.

Hydrol. Process., Vol. 11, 1451–1464 (1997).

(No. of Figures: 7 No. of Tables: 1 No. of Refs: 37)

KEY WORDS blowing snow; Arctic; snow distribution; hydrology; sublimation; precipitation; snowfall; snow drifts; tundra; evaporation

INTRODUCTION

The Arctic is famous for its severe blizzards and long snow-covered season. As meteorological events, the significance of blowing snow storms to Arctic travel and outdoor life is well known. However, it is less well known that wind transport and sublimation of blowing snow can promote significant annual vertical and horizontal fluxes of water and energy from the Arctic regions. These fluxes are:

(1) water vapour — transferred upwards from the surface as a result of blowing snow sublimation,
(2) sensible heat — transferred downwards from the atmosphere towards the surface to drive sublimation,
(3) snow — blown from low aerodynamic roughness to high aerodynamic roughness surfaces, and
(4) latent heat — transferred with the relocated snow and realized as the latent heat of fusion at the time of spring melt.

In open, level, wind-swept locations on the Canadian Prairies, blowing snow has been found to remove up to 70% of annual snowfall with over 40% of annual snowfall sublimating in transit, depending upon the uniform upwind fetch (Pomeroy *et al.*, 1993a). Tabler *et al.* (1990b) estimated snow transport from wind

CCC 0885–6087/97/111451–14$17·50

Received 29 November 1995
Accepted 15 February 1996

speed records on the Arctic coast of Alaska, finding that over several years about 86 Mg/m width of snow was transported to snow fences near Prudhoe Bay. Benson (1982) estimated snow relocation in the same region using snow drifts formed by the Meade River, snow surveys on adjacent tundra and corrected precipitation records. When his data are re-examined using a snow mass balance about 58% of snowfall remained as accumulation on the tundra with 11% of annual snow relocated to the Meade River valley and 32% of snowfall sublimated as blowing snow. It is not known how representative these values are of the Arctic region, however, they demonstrate that the fluxes are significant even in the cold Arctic winter.

Relocation and sublimation of blowing snow can cause dramatic changes to the water balance of high latitude and/or altitude basins. There is increasing evidence for this in the Arctic as improved estimates of winter precipitation become available. These corrected Arctic snowfall estimates confirm that snowfall is greater than previously thought; some estimates have been compared to snow on the ground and show that snowfall significantly exceeds snow accumulation (Goodison, 1981; Benson, 1982; Woo *et al.*, 1983; Claggett, 1988; Tabler *et al.*, 1990a; Benson and Sturm, 1993). The redistribution of snow by wind forms snow covers of highly variable depth and density, whose variation governs the energetics at the surface during melt (Shook, 1995). In the irregular terrain typical of much of the Arctic, relocation is governed partially by topography. Highly dissected terrain with short fetches has less opportunity for sublimation than level plains with long fetches, and snow is relocated from areas of low to high aerodynamic roughness without regard for watershed boundaries or overland flow divides. The resulting spatial distribution of snow water equivalent is important for modelling the timing, amplitude and persistence of the snow melt freshet (Marsh and Pomeroy, 1995). For these reasons, physically based, spatially distributed process models of snow hydrology are required to calculate snow fluxes over a range of scales (Marsh *et al.*, 1995; Pomeroy and Gray, 1995). The processes of winter mass exchange between the snow cover and the atmosphere have not been fully investigated for the Arctic, nor have they been incorporated in any hydrological or atmospheric model. It is the purpose of this paper to outline a spatially distributed blowing snow model, using simple algorithms of snow transport and sublimation, and to apply the model to an experimental catchment in the low Arctic. The seasonal variation in blowing snow fluxes and the annual values for the catchment are calculated and compared with field measurements.

BLOWING SNOW PROCESSES

Blowing snow occurs in two substantive modes of transport, saltation and suspension. The physics of blowing snow has been recently reviewed by Pomeroy and Gray (1995) and is briefly summarized here. The source of blowing snow can be surface snow or falling snow. For simplicity, blowing snow without falling snow is described. Saltation is the motion of blowing snow crystals in a two-phase flow layer several centimetres thick, skipping just over the snow cover (Kobayashi, 1972; Schmidt, 1986; Pomeroy and Gray, 1990). The snow surface is eroded and entrained by saltation when the atmospheric shear stress exceeds the stress that is, (a) necessary to shatter the bonds of snow surface crystals, and (b) applied to exposed vegetation stalks or leaves.

The snow transport rate by saltation increases linearly with wind speed and is sensitive to snow surface hardness, being, in general, higher over a hard snow crust. Total snow transport increases with the fourth power of wind speed. Because of the difference in the rate of increase with wind speed of saltation transport and total transport, Pomeroy (1989) and Pomeroy and Gray (1990) found that saltation contributes generally less than 25% of annual transport, and an ever smaller portion to total transport as wind speed increases.

Before snow can become suspended by the wind it must be entrained by saltation (Schmidt, 1980). Suspension is the motion of snow crystals supported by atmospheric turbulence (Budd *et al.*, 1966; Schmidt, 1982; Pomeroy and Male, 1992). The suspended layer has a lower mass concentration than does the saltation layer and it extends from the top of the saltation layer up to several tens of metres above the snow surface. Because of its high horizontal velocity and vertical thickness, the suspended transport rate increases with the fourth power of wind speed. For high wind speeds (>15 m/s), the suspended transport rate is over an order of magnitude higher than that for saltation (Pomeroy, 1989).

The transport of blowing snow is sensitive to the availability of surface snow for erosion. This is controlled by the depth and cohesive bonding of snow, which are indexed by the magnitude of the 'threshold' wind speed to initiate saltation. When the snow temperature exceeds $-5\,^{\circ}$C, liquid layers on snow crystals increase notably in thickness (Conklin and Bales, 1993) and increase cohesion between particles; as a result the threshold wind speed for saltation increases (Ôura *et al.*, 1967). Surface crusts having high thresholds for transport can form even on normally cold snowpacks, and are most commonly caused by surface metamorphism driven by advection of warm air masses and high winds (Colbeck, 1989), 'radiation melt' at below-freezing air temperatures (Brandt and Warren, 1993) and freezing rain. These processes can affect a layer several centimetres deep into the pack via sharp temperature gradients, wind-pumping, radiation penetration and infiltration of rain into snow.

Transport is also sensitive to the atmospheric shear stress applied to the surface. For a constant, upper-level wind speed, the shear stress applied varies with the degree to which vegetation is exposed above the snow and with the topography of the surface. For dense, i.e. closely spaced, plant growth, saltation cannot normally become established until vegetation is filled in with accumulated snow to within one to five centimetres of the stem tops (Tabler and Schmidt, 1986; Pomeroy and Gray, 1995). The exact degree of plant exposure required to inhibit transport depends upon vegetation spacing and wind speed. Flow separation in the lee of hills, ridges or uplands may cause the cessation of saltation and suspension and subsequent deposition of snow. Acceleration near hill-crests may promote increased frequency and intensity of blowing snow and enhance erosion and transport. Tabler (1975b) modelled the profiles of snowdrifts in topographic catchments. His model suggests that lee slopes of about 9° promote the formation of drifts with about 5–7° grade, drift grade increasing with slope length. Tabler (1994) suggests that lee slopes of from 10° to 12° are required to cause separation of air flow. Over natural Arctic landscapes redistribution will occur from sparsely vegetated tundra plateaux, ridges and hilltops to hillsides, incised valleys, gulleys, shrub tundras and pockets of subarctic woodland.

Sublimation is the transfer of snow crystal mass to water vapour and occurs rapidly during blowing snow because of the high ratio of particle surface area to mass, excellent ventilation of blowing snow particles and atmospheric water vapour deficit. Ventilation and water vapour deficits are usually large during blowing snow events because of the strong atmospheric mixing that occurs (Dyunin, 1959; Schmidt, 1982; Pomeroy *et al.*, 1993a). Sublimation is an extremely dynamic phenomenon, occurring from a column of blowing snow particles, rather than from a static 'surface'. To describe this degree of complexity, physically based calculations use involved numerical energy and mass balance solutions as part of two-phase flow descriptions (Schmidt, 1982; Pomeroy *et al.*, 1993a). Certain patterns of sublimation are apparent over longer time-scales which are described by semi-empirical formulations for monthly or annual sublimation fluxes on various landscape types (Tabler, 1975a; Pomeroy and Gray, 1994).

Several aspects of the blowing snow phenomenon are important for calculating fluxes on irregular terrain such as an arctic landscape. Fetch (upwind distance of uniform terrain) is an important variable. Takeuchi (1980) showed that at least 300 m of snow-covered fetch is required for snow transport to reach a steady state (no further increase with fetch) in the lowest metre of the atmosphere. For lengths less than 300 m, snow transport increased asymptotically with increasing fetch distance. Pomeroy and Gray (1995) suggest that about 500 m of snow-covered fetch is required for blowing snow development to a 5 m height (fully developed flow). They also found that when the fetch is not completely snow-covered (for example, where wind-scoured zones have developed in the lee of obstacles to transport), annual transport was fully developed for fetches greater than 500–1000 m. The fetch distance strongly controls the proportion of blowing snow transported to an accumulation zone compared with the amount that sublimates in transit. The sublimation flux is vertical (mass loss accumulates with fetch) and the transport flux is horizontal (mass loss does not accumulate with fetch), hence in the summation of these fluxes over a fetch, sublimation begins to dominate for longer fetches (Pomeroy and Gray, 1995). This result is corroborated for the Arctic by

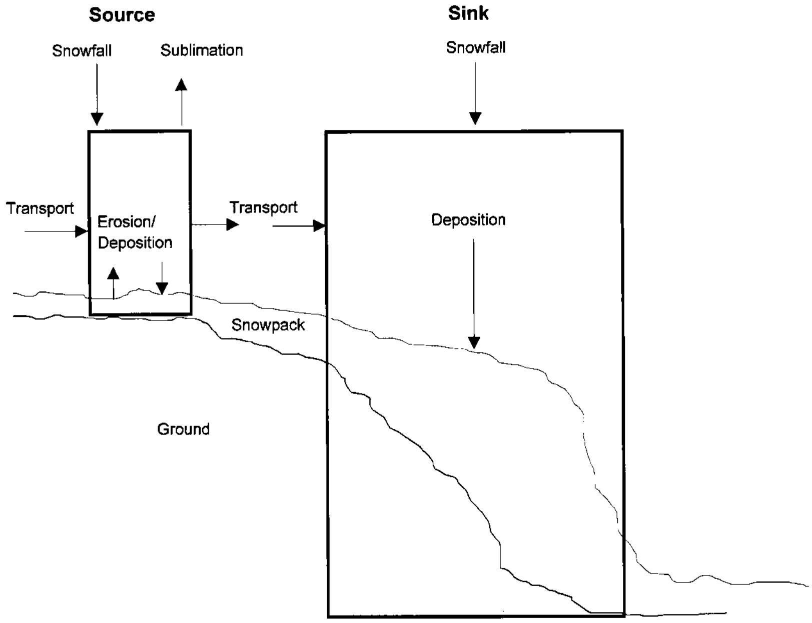

Figure 1. Distributed blowing snow model. Side-view schematic showing snow fluxes and mass balance using the concept of a control volume for 'source' and 'sink' landscape types

Benson (1982), who calculated that snow particles may blow an average transport distance of 2·5–3·5 km before completely sublimating on the Alaskan Arctic coast.

DISTRIBUTED BLOWING SNOW MODEL

The distributed blowing snow model (DBSM) is a series of algorithms intended to calculate the mass and energy fluxes of blowing snow and the resulting annual snow accumulation over heterogeneous catchments. The catchment, as part of a meso-scale 'domain' (scale: 10s of km) is modelled as an assortment of internally homogeneous landscape elements having specific vegetation, terrain, exposure and fetch characteristics. Blowing snow processes for various elements are calculated using meteorological information. The DBSM is distributed in that it divides landscape elements into 'sources' and 'sinks' of blowing snow, and calculates the snow fluxes as a mass balance to control volumes over each element type whilst retaining a mass balance over the meso-scale domain (Figure 1). Note that a one-dimensional mass balance is not necessarily specified for the catchment, i.e. blowing snow can enter or leave the catchment. Net accumulation for the catchment is based on transport and sublimation fluxes from landscape elements and the proportion of sinks or sources contained in the catchment compared with the meso-scale domain. For a homogeneous landscape element, the instantaneous snow accumulation flux, Q_{surface}, is calculated using a mass balance of snow fluxes to a control volume

$$Q_{\mathrm{surface}} = Q_{\mathrm{snowfall}} - \frac{\mathrm{d}Q_{\mathrm{T}}}{\mathrm{d}x}(x) - Q_{\mathrm{E}} \tag{1}$$

in which Q_{T} is the transport flux (kg/ms), x is the distance along the fetch (m), Q_{E} is the sublimation flux (kg/m^{2}s) and Q_{snowfall} is the snowfall flux (kg/m^{2}s). The change in transport flux, Q_{T}, over horizontal distance x is an important term when fetches are short or terrain is uneven. For source areas, i.e. surfaces with $Q_{\mathrm{surface}} < Q_{\mathrm{snowfall}}$, the transport change term, $\mathrm{d}Q_{\mathrm{T}}/\mathrm{d}x(x) > -Q_{\mathrm{E}}$ in that the blowing snow processes combine to produce a net loss from the surface. For sink areas, i.e. surfaces where $Q_{surface} \geqslant Q_{snowfall}$, $dQ_T/dx(x) \leqslant -Q_E$, producing a net balance or gain from blowing snow processes to the surface. Cases where $Q_{surface} = Q_{snowfall}$ in wind-swept terrain are exceedingly rare. The term $dQ_T/dx(x)$ becomes insignificant over the long fetches typical of large homogeneous tundra plateaux; there, $Q_{surface}$ is a balance between sublimation and snowfall. Note that $Q_{surface}$, as a residual term of the precipitation and blowing snow processes, can be positive or negative. The task in applying the DBSM is specifying appropriate fluxes for various characteristic landscape elements and aggregating these to the catchment scale over a season.

The core of the DBSM requires a procedure to calculate fluxes of snow transport and sublimation. One such procedure is the prairie blowing snow model or PBSM (Pomeroy *et al.*, 1993a), a physically based fine-scale model of blowing snow erosion, saltation, suspension sublimation and deposition over level landscapes of varied roughness and fetch. Individual algorithms of the PBSM have been tested against field measurements of blowing snow. Tests to verify the complete PBSM operation have compared modelled accumulation with field measurements of snow water equivalent (SWE), with differences of the order of 10% or less in the prairie environment (Pomeroy *et al.*, 1993a). The PBSM relies on hourly observations of the occurrence of blowing and drifting snow (as well as a suite of standard meteorological parameters) to calculate a threshold wind speed and then calculate the appropriate hourly mass and energy fluxes over a set landscape Unfortunately, the data requirements and complexity of the PBSM restrict its application to major climatological stations, and therefore it is not yet a suitable 'core' for the DBSM in arctic applications. To estimate transport and sublimation where data is limited, a simplified climatological version (CBSM) was derived from the output of the PBSM for 15 stations on the prairies over six years of simulation (Pomeroy and Gray, 1994, 1995). The CBSM calculates the monthly flux of snow removed and sublimated in transit (sublimation flux) and that removed and transported to the end of the fetch (transport flux) for a standard fetch distance of 1 km over unvegetated (fallow) and vegetated (25 cm grain stubble) land cover. The mean differences between monthly sublimation and transport fluxes calculated by the CBSM, and monthly summations of these fluxes from the PBSM are of the order of 10^{-7} mm SWE, with correlation coefficients of 0·83 for simulations with a vegetated land cover.

The CBSM simulations for vegetated, level terrain provide a suitable and practical initial core for the DBSM. For source areas, fully developed flow at the lee end of the fetch and negligible transport into the fetch are assumed. Monthly transport change, dQ_T/dx, where $x = 1000$ m downwind from a barrier to blowing snow, is (Pomeroy and Gray, 1994),

$$\frac{dQ_T}{dx}(x) = -8{\cdot}259 + 0{\cdot}889u_{10} + 5{\cdot}698e^{[-0{\cdot}101(T_{max}+20)]} + 0{\cdot}041RH_{max} + 3{\cdot}318e^{\left(\frac{-8{\cdot}716}{d}\right)} \quad (2)$$

For source areas, the sublimation flux Q_E for a fetch $x = 1000$ m is then (Pomeroy and Gray, 1994),

$$Q_E(x) = -6{\cdot}927 + 1{\cdot}846u_{10} - 0{\cdot}171T_{min} - 0{\cdot}074RH_{min} + 0{\cdot}010S_m + 5{\cdot}218e^{\left(\frac{-6{\cdot}12}{d}\right)} \quad (3)$$

The input parameters are:

$$\begin{aligned}
u_{10} &= \text{mean monthly wind speed at 10 m height (m/s)},\\
T_{max} &= \text{monthly mean of daily maximum air temperature (°C)},\\
T_{min} &= \text{monthly mean of daily minimum air temperature (°C)},\\
RH_{max} &= \text{monthly mean of daily maximum relative humidity (\%)},\\
RH_{min} &= \text{monthly mean of daily minimum relative humidity (\%)},\\
S_m &= \text{mean monthly snowfall (mm SWE), and}\\
d &= \text{mean monthly depth of snowcover (mm SWE).}
\end{aligned}$$

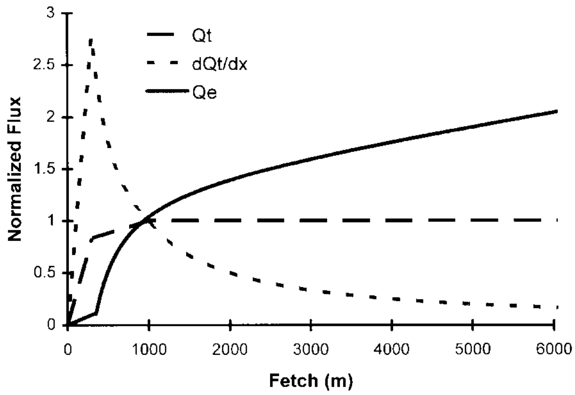

Figure 2. Variation of annual blowing snow fluxes with fetch distance. Increase in annual transport rate, change in transport rate and sublimation rate with fetch distance over level surfaces. All rates are normalized to rates found at fetch = 1000 m, i.e. $Q_x(1000\ \mathrm{m}) = 1$

These parameters are commonly available from meteorological station records, although S_m should be corrected for wind-induced undercatch and systematic wetting loss as recommended by Goodison (1978) and Sevruk (1992). For locations of varying fetch, Q_T and Q_E will vary from the 1000 m values. Pomeroy and Gray (1995) suggest the following adjustments, based on the variation with fetch of annual summations of transport and sublimation fluxes predicted by PBSM. For fetches of length $300 < x < 1000$ m with $Q_\mathrm{T} > 20\,000$ kg/m, the transport flux may be found following Pomeroy and Gray's (1995) approximation of fluxes greater than 20 000 kg/m as a function of the 1000 m fetch flux, $Q_\mathrm{T}(1000)$ as

$$Q_\mathrm{T}(x) = 4{\cdot}6(1000 - x) + Q_\mathrm{T}(1000\ \mathrm{m})\left(0{\cdot}58 + \frac{x}{2381}\right) \tag{4}$$

For $x > 1000$, and for $Q_\mathrm{T} < 20\,000$ kg/m then, Equation (4) does not apply and the relationship is $Q_\mathrm{T}(x) = Q_\mathrm{T}(1000\ \mathrm{m})$. From Pomeroy and Gray (1995) the sublimation flux may be found for $x > 300$ m as

$$Q_\mathrm{E}(x) = Q_\mathrm{E}(1000\ \mathrm{m})\left(1{\cdot}37 - \frac{457}{x} + \frac{x}{8000}\right) \tag{5}$$

The variation of Q_T, $\mathrm{d}Q_\mathrm{T}/\mathrm{d}x$ and Q_E with fetch, x, is shown in Figure 2 with reference to their normalized values for $x = 1000$ m and the assumption that fluxes are 0 at $x = 0$. It is seen that both $\mathrm{d}Q_\mathrm{T}/\mathrm{d}x$ and Q_E increase rapidly for fetches less than 300 m and that the normalized sublimation flux, $Q_\mathrm{E}(x)/Q_\mathrm{E}(1000\ \mathrm{m})$, increases steadily to about 2·0 for a fetch distance of 5700 m. The normalized vertical transport flux rapidly declines from $>2{\cdot}5$ at small fetches to less than 0·2 for fetch distances greater than 5000 m.

For sink areas, Equation (1) still applies, but sublimation is assumed to be negligible. Transport into the fetch ($x = 0$) is equal to transport out of the upwind source fetch, Q_T (x upwind). Transport out of the fetch is zero, which presumes that a sink landscape element is a completely efficient snow trap. Accumulation in a sink is therefore a summation of snowfall and blowing snow inputs.

For a simple 'first approximation' application of the distributed model, fluxes are calculated for source areas and distributed to various sink landscape elements based on the proportion of sink landscape that is oriented perpendicular to the major winter wind directions. Mean fetch lengths for various landscape types are measured for the catchment and assigned to appropriate landscape elements.

APPLICATION OF THE DBSM

Location, measurements and characteristics

The DBSM was applied to the Trail Valley Creek basin, a 68 km^2 NHRI-GEWEX research basin located north of the tree line and 50 km north of Inuvik, Northwest Territories. Basin elevations range from 60 to

190 m above sea level. The valley adjacent to the creek is incised in a tundra plateau and runs east–west, forming an effective trap for blowing snow from the NW–SE directions. Vegetation in the region is forest–tundra transition (Bliss and Matveyeva, 1992), hence upland plains are poorly vegetated tundra (lichens, mosses, willows) whilst moister hillslopes and valley bottoms are either shrub tundra (alder, willows) or sparse black spruce forest. The surfaces evident on a recent Landsat TM image (mid-summer) were classified using a supervised classification on PCI software. The classification used a maximum likelihood procedure on channels 3, 4, 5 and 7 (red, near- and mid-infrared) and was resampled from 30 m to 20 m to provide higher resolution and to aid in the classification of mixed pixels. The classification was extensively field tested (species, leaf area index, vegetation height, understory, etc.) to produce a vegetation map of the catchment with a 20 m resolution and classifications corresponding to vegetation types proposed by Strong *et al.* (1990).

Elevations from 1:50 000 topographic map sheets were digitized to produce a digital elevation model of 40 m spatial resolution. Drift areas were determined from the vegetation and slope ($>9^\circ$ or stream channel and tundra vegetation), and verified using aerial photographs of late-lying drifts in three years. Pomeroy *et al.* (1993b) found that large drifts, which often exceed 2 m in depth with densities in excess of 350 kg/m^3, form on these areas; the 9° slope definition also fits well with the findings of Tabler (1975b). The resulting landscape classification (water, exposed soil, tundra, sparse shrub tundra, open shrub tundra, closed shrub tundra, sparse forest and drift) as shown in Figure 3, was simplified into landscapes characteristic of snow accumulation in the region (Pomeroy *et al.*, 1993b; Pomeroy *et al.*, 1995). Water, tundra, sparse shrub tundra and exposed soil were lumped into the 'tundra' category and the open and closed shrub tundras were lumped into the 'shrub tundra' category so that the classification could be used to apply the DBSM. Based upon this classification, the Trail Valley Creek basin has an arctic–subarctic transitional vegetation pattern of 70% tundra, 21·5% shrub tundra and 0·5% sparse forest (taiga). Approximately 8·2% of the basin is a 'potentially drift-forming' surface, either a hillslope of greater than 9° grade or a stream channel (presumed steep grade), and is non-forested. About 70% of these slopes occur on hills and valley sides and 30% occur alongside stream channels.

Average fetches were determined using 'ruler' measurements of uniform landscape class in two directions (NW–SE, SW–NE). The NW–SE ordinate is the dominant wind direction during snow transport, however, other directions are not negligible and the model does not include a directional component; it was therefore determined best not to bias the fetch sample without such a feature in the model. Fetches determined in this manner over the catchment, range from 1450 m on tundra to 500 m for shrub tundra and 100 m for drifts.

Trail Valley Creek has a 'low arctic' climate with snow accumulation commencing in October and melt occurring in May. Consistent low winter temperatures and high winds result in frequent blowing snow storms; the occurrence of mid-winter melts is uncommon. The simulation used data from October 1992 to May 1993. A meteorological station at Trail Valley and precipitation time-series from Inuvik were used to develop the monthly climatology for this period. The unmanned (in winter) NHRI-GEWEX meteorological station within the basin provides half-hourly measurements of air temperature, humidity, wind speed, depth of snow and snowfall occurrence and is described by Marsh and Pomeroy (1995). All instruments have been calibrated but are not regularly maintained in mid-winter. Snow-filled cups and rime-ice on the anemometers reduced wind-speed measurements over some unattended periods. To correct for wind undermeasurement, periods when the anemometers were noticeably 'seized' or obviously not responding to blowing snow events as indicated by snow particle detectors (Brown and Pomeroy, 1989) were excluded from the monthly average. Even with this correction, wind speeds are probably underestimated and a solution to this problem for remote Arctic sites will remain a priority of future field work.

'True' winter snowfall was estimated from measurements of snow water equivalent made within a small, circular glade, well within an open forest. The forest fetch was sufficiently long in all directions to cause deposition of blowing snow to surfaces upwind of the glade. The aerodynamic characteristics of this glade ensured that no blowing snow is eroded or trapped and no snow is intercepted, hence it is neither a sink nor a source of blowing snow. Cold winter weather ensures that there was little or no melting and that *in-situ* sublimation is negligible over the winter. Snowfall measured 'in-glade' was 45% higher than the uncorrected

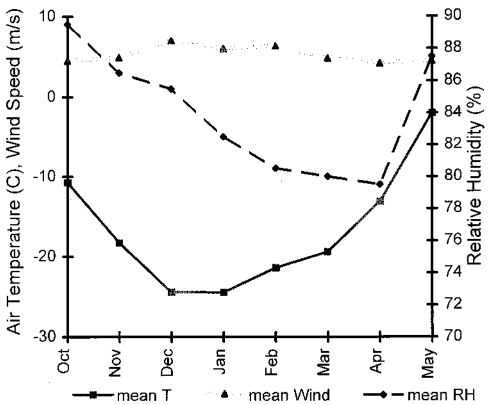

Figure 4. Mean monthly climate of Trail Valley Creek, winter 1992–1993. Mean monthly air temperature, relative humidity and wind speed (corrected) from half-hourly measurements by the NHRI-GEWEX climatological station located on the tundra plateau

winter snowfall recorded at the Inuvik weather station. Such a difference is not unexpected for the Arctic (Claggett, 1988; Benson and Sturm, 1993). Using the 'true' winter snowfall from the glade, monthly measured snowfall for Inuvik and the output of an optoelectronic snow particle detector (Brown and Pomeroy, 1989), the monthly snowfall record for Trail Valley Creek was reconstructed. Monthly snowfall relative to the annual total at Inuvik was weighted by the 'true' winter total for Trail Valley Creek to provide the best estimate of monthly snowfall at Trail Valley Creek.

The mean monthly climate at the catchment over the winter of model application (1992–1993) is shown in Figure 4. Mean monthly wind speed ranged from 4·2 to 7·0 m/s, with peak wind speeds from December to February. Mean monthly daily maximum temperature ranged from $-20{\cdot}3^{\circ}$ to $+1{\cdot}9\,^{\circ}$C and mean monthly daily minimum temperature from $-28{\cdot}7^{\circ}$ to $-5{\cdot}9\,^{\circ}$C, with the coldest period from December to January. Mean monthly daily maximum relative humidity declined from 94 to 85% as winter progressed, recovering to early winter values in May. Monthly snowfall varied from 8 to 39 mm SWE with peak snowfalls in October and January and smallest snowfalls from March to May. Depth of snow on the ground (expressed as equivalent water and measured on tundra) increased from 35 to 50 mm SWE over the winter. Total winter snowfall was estimated from early May snow survey measurements in the small glade trees as 190 mm SWE.

Validation data

Extensive snow surveys were used to validate the annual accumulation of snow predicted by the DBSM. Surveys were conducted in early May and stratified by landscape type in the basin. Transects were devised to cross the 'characteristic' landscape types of the landscape classification corresponding to 'open tundra', 'shrub tundra', 'drifts' and 'sparse forest'. In each landscape, 125 depth measurements at 5 m spacings were taken along a transect. Density was measured, using the ESC-30 metric snow sampler, at five to six points along every transect. Snow sampling along a drift was varied as necessary to accommodate the dimensions of the drift.

Application technique

For an Arctic application, tundra plains are considered sources; transport and sublimation fluxes for standard 1 km fetches of tundra are calculated initially using Equations (2) and (3). These fluxes are adjusted for the measured fetch using Equations (4) and (5), and used with Equation (1) to calculate:

(1) sublimation and transport loss and accumulation in mm snow water equivalent (SWE) over the tundra surface, and

(2) the mass flux (kg/month) to sink areas of the basin.

Sink areas are defined as:

(1) drifts — slopes greater than 9° and stream channels adjacent to tundra fetches, and

(2) shrub tundra and sparse forest — areas of tall shrubs and open transitional forest that buffer tundra areas.

The mass fluxes to sinks are distributed according to the percentage cover of a landscape type downwind that is not oriented to the prevailing snow transport directions from the tundra. To be counted in calculating the percentage cover, downwind sink areas must have fetches greater than or equal to the average fetch for that landscape type. This filters out short buffer strips of shrub–tundra bordering a drift area that would fill early in the winter and have little effect on drift accumulation. In Trail Valley Creek, the distribution of blowing snow mass flux from the tundra is 30% to shrub–tundra and 70% to drifts. The incoming mass of blowing snow is distributed over the average fetch length of the sink landscape to arrive at a snow transport value in mm SWE. Equation (1) is used to calculate the mass balance for sink landscapes with the presumption that sublimation from the sink is negligible. The model is run for the whole area shown in Figure 3, with no limitation to fluxes at the catchment boundary. Hence, snow can be imported or exported from Trail Valley Creek by blowing snow, although the mass balances shown are only for terrain within the catchment boundary.

SIMULATION RESULTS

Monthly fluxes of precipitation, transport, sublimation and accumulation are shown for the tundra, shrub–tundra and drift landscapes, respectively, in Figures 5a, b and c. They show peaks in blowing snow transport and sublimation fluxes in December when cold, dry snow covers and high wind speeds promote relocation. The spring and autumn months have wet or icy snow covers which are less frequently eroded, and then only by the highest wind speeds. The simulation for tundra landscape (Figure 5a) shows the temporal accumulation pattern roughly following that of precipitation in the autumn and spring, but it is largely decoupled from this pattern for December to March. Mid-winter losses to transport and sublimation often cause monthly accumulation to be less than half of snowfall. Sublimation from tundra blowing snow caused the greatest losses in all months except May. Monthly sublimation loss exceeded 10 mm for December to February. Negative accumulation in March reflects low snowfall and erosion of deposited snow.

The simulation for shrub–tundra landscape (Figure 5b) shows an accumulation regime that is well coupled to the precipitation regime with a persistent additional input from incoming blowing snow that is generally of the order of 5 mm SWE per month or less. The drift landscape simulation (Figure 5c) shows that accumulation is highly uncoupled from precipitation, and that accumulation fluxes are mostly comprised of the incoming transport flux of blowing snow. Blowing snow transport exceeds the snowfall input to the drifts except in October. Peak accumulation is in January and December. For December, March and April, the incoming blowing snow transport flux is over four times greater than snowfall, and in most months it is at least double that of snowfall.

Over the winter 49 Mg/m (per metre width perpendicular to the tundra edge) of snow was transported off the tundra into sinks in shrub–tundra and drift areas. The snow was removed in primarily south-east and north-west directions. When distributed over the spatial extent of the sink areas, this results in an annual water equivalent input from blowing snow events of 346 mm to drifts and 30 mm to shrub–tundra landscapes in Trail Valley Creek.

The DBSM shows that on tundra landscape in Trail Valley Creek, 28% of annual snowfall was sublimated, 18% transported to drifts or shrubs and 54% remained to accumulate as the annual snow cover. Blowing snow transport to shrub–tundra and drifts amounted to 16% and 182% of winter snowfall, respectively. The fluxes result in snow accumulations in shrub–tundra and drifts that are 113 and 419% greater than that on level tundra, respectively.

The landscape classification was used to calculate the basin-weighted averages for snow fluxes, shown in Table I. Trail Valley Creek received a net import of blowing snow transport from outside its drainage

Table I. Annual snow fluxes in Trail Valley Creek as predicted by the DBSM, 1992–1993

Snow flux	% of Snowfall	Snow water equivalent (mm)
Sublimation	19·5	−37
Transport	5·8	11
Accumulation	86·5	164

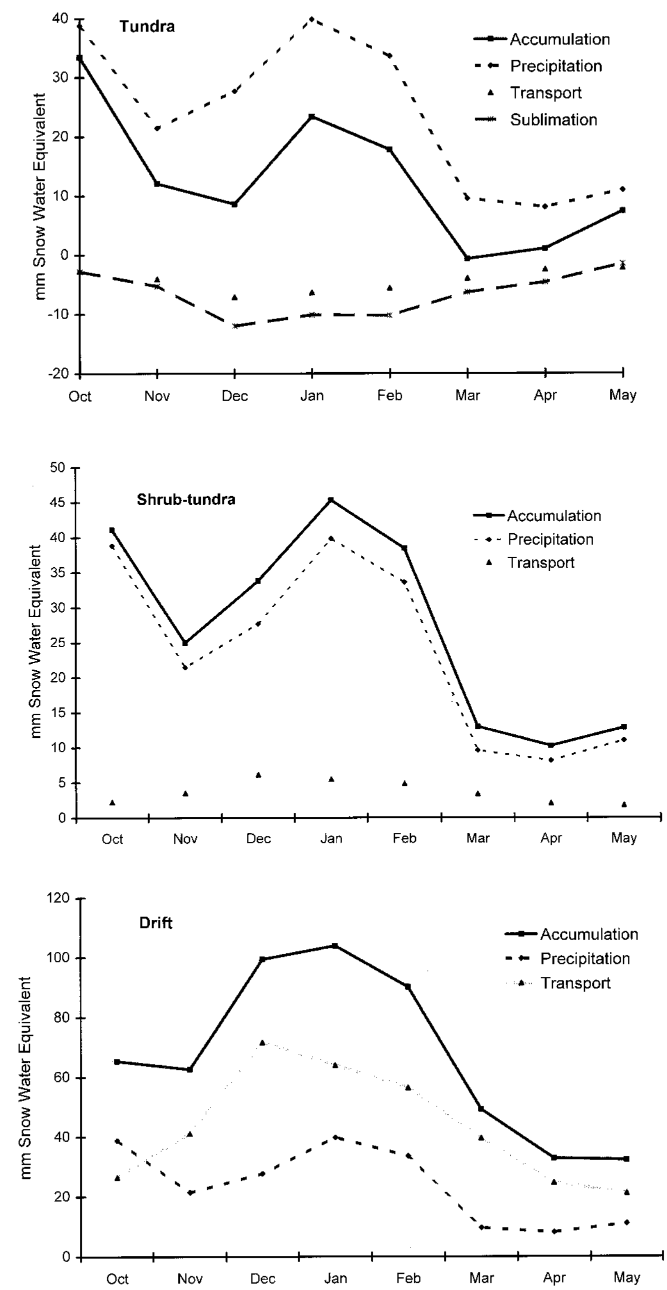

Figure 5. Monthly fluxes of snow accumulation, precipitation, transport and blowing snow sublimation, Trail Valley Creek, 1992–1993. (a) Tundra simulation. (b) Shrub–tundra simulation. (c) Drift simulation

boundaries equivalent to 11 mm SWE. Other catchments in the area or subcatchments would receive more or less, depending upon their topography, vegetation and surroundings. Basin snow accumulation is over five-sixths of winter snowfall; the 19·5% loss to sublimation is partially compensated by a net transport flux into the basin equivalent to an import of 5·8% of winter snowfall.

The spatial distribution of snow water equivalent predicted by the DBSM is shown in Figure 6. The pattern is complex, but because of the links between topography and vegetation it largely follows the stream valley system through the basin, with low accumulations on the tundra plains that form the basin edges and high accumulations on the shrub–tundra and drifts that dominate the valley walls and bottomlands. The spatial variation in sublimation flux follows roughly the spatial distribution of the tundra landscape, the lowest snow water equivalent class. Percentages of total basin snow contained in each landscape type were, tundra (43%), shrub–tundra (29%), drifts (27%) and sparse forest (0·6%). Hence, for this year, 43% of basin snow is contained in shallow snow covers and 57% in deep snow covers.

VERIFICATION OF RESULTS

Without direct measurements of sublimation, frequent landscape-based snow surveys over the winter and a greater spatial density of snow survey measurements, the DBSM simulation cannot be fully verified. Extensive snow surveys conducted in the spring were used to test the model results for accumulation, transport to drifts and sublimation loss in the specified landscape classes. The comparison of class averages at the end of winter is useful, but, as the class definitions have not been proven to correspond to classes of snow accumulation and no variation within class is specified by the DBSM, a statistical comparison is not yet warranted.

The classification of basin landscapes was used to distribute average landscape SWE from the May measurements over the basin. The basin average SWE is modelled as 164 mm by the DBSM and estimated as 154 mm by the snow surveys. The difference represents an overestimation by the model of 6%, which is well within the expected measurement errors for this type of comparison. Simulations for individual landscape types are not as well matched to measurements as the basin total. Figure 7 shows the modelled and measured SWE for each landscape type and indicates that the pattern of accumulation is overestimated for source areas and is underestimated for sinks. The DBSM overestimated snow accumulation on the tundra by 35 mm SWE (51%) and underestimated snow accumulation on the shrub–tundra by 32 mm (13%), in the sparse forest by 24 mm (10%) and for drifts by 81 mm (13%). Nevertheless, the relative pattern of snow accumulation is correct and errors are compensatory, largely averaging out over the basin area.

A further check of sublimation and transport calculations was made by a basin mass balance using Equation (1) and balancing the sum of modelled basin sublimation and transport losses or gains to the difference between winter snowfall and the basin-averaged snow survey results. Winter snowfall exceeds the

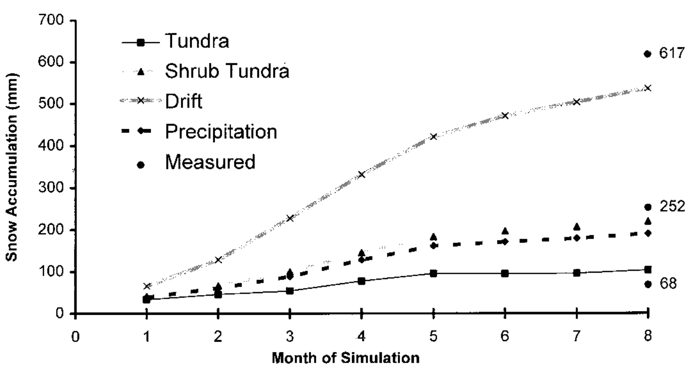

Figure 7. Modelled cumulative snow accumulation, measured maximum snow accumulation (●) and measured cumulative precipitation for landscape types in Trail Valley Creek. Simulation from October 1992 to May 1993

basin-averaged snow survey by 37 mm SWE. Basin-averaged sublimation loss (37 mm) less the net transport gain (11 mm) equals 26 mm SWE, a 30% difference indicating an underestimation of sublimation. The 11 mm difference is relatively small, being less than 6% of winter snowfall, and suggests that sublimation as a portion of the basin water budget would not increase substantially upon an improvement of model performance or of data quality.

It is possible that underestimation of wind speed contributed to the underestimation of redistribution. The model indicates that a 1 m/s increase in wind speed would cause a 24% decrease in tundra snow accumulation and a 10% increase in drift snow accumulation. The long-term average winter wind speed at Tuktoyaktuk, 80 km north of Trail Valley, and more representative of a tundra wind regime than Inuvik, is 6 m/s, whilst the estimated average for Trail Valley in the winter 1992–1993 is 5.3 m/s. The potential difference in measured wind speeds could account for much of the underestimation of transport in this application.

DISCUSSION

These results are specific to one winter at Trail Valley Creek, given its size, vegetation cover and topography. A less dissected basin, or a subcatchment dominated by tundra plateaux will have a much higher sublimation loss, whilst a small, incised subcatchment in the same area would have a lower sublimation loss and a much higher transport gain from snow imported to the catchment. Rather than providing sublimation and transport values that can be extrapolated without modification elsewhere, the model results show the need for and success of relatively simple, distributed blowing snow algorithms in predicting the spatial distribution and basin averages of snow water equivalents in the Arctic. The strong variation in snow water equivalent with landscape class has important implications for a variety of hydrological and climatological processes. The spatial distribution of snow depth will exert a control on the energetics of melt by causing a more rapid areal albedo decline during melt than if the snow were more evenly distributed (Shook, 1995). The melt pattern will also be influenced, as areas of shallow snow melt several days ahead of areas with deep snow (Marsh and Pomeroy, 1995). The total meltwater runoff and its timing from individual subcatchments in the basin will depend largely upon the configuration of snow sources and sinks in these subcatchments. This configuration will therefore affect subsequent routing of runoff at larger scales.

CONCLUSIONS

The application of a distributed blowing snow model to a dissected low arctic basin and the results of landscape-stratified snow surveys and corrected snowfall measurements have demonstrated the following.

(1) Blowing snow fluxes are large in the Arctic and can exceed snowfall fluxes on many surfaces in mid-winter.
(2) Incised catchments can 'import' wind-transported snow from adjacent tundra plains. The magnitude of such gains will depend on the scale of comparison, basin topography and vegetation, and are very sensitive to wind speed.
(3) Sublimation losses are notable on a site scale and catchment scale (up to 28% and 19·5% of snowfall, respectively, in this case), providing an important latent heat flux, loss of surface water supply and source of atmospheric water vapour that have not been explicitly considered in winter hydrological or meteorological models.
(4) Snow relocation causes snow covers in the Arctic to be highly variable at the end of winter (varying from 54 to 419% of snowfall in this case). The interaction of blowing snow processes with landscape aerodynamic features controls the spatial variation and average value of the net basin snow accumulation; hence, both meteorological and landscape parameters must be considered to calculate snow retention in wind-swept regions.

(5) Blowing snow processes and landscape patterns govern the spatial distribution and total amount of snow water equivalent at the time snowmelt begins (varying from 103 to 536 mm water equivalent in this case), influencing the subsequent surface energetics, timing and magnitude of snowmelt, infiltration and stream discharge.
(6) Relatively simple algorithms of snow transport and sublimation can, when distributed by landscape type over a basin, produce reasonable estimates of basin snow redistribution, accumulation and sublimation loss in wind-swept regions.

Because of the influence of basin landscape characteristics on blowing snow fluxes, these results should not be extrapolated to other catchments without reference to local climate and terrain.

ACKNOWLEDGEMENTS

The authors acknowledge the support of the Canadian GEWEX Programme; the Canadian Climate Research Network, Land Surface Node; the Polar Continental Shelf Project, Dept of Natural Resources Canada; the Science Institute of the Northwest Territories, Inuvik Research Laboratory, Government of the Northwest Territories; Water Resources Division, Dept of Indian and Northern Affairs Canada (Yellowknife) and their respective institutions. The field assistance of C. Onclin, R. Reid, K. Dion and W. Quinton and analytical, GIS and mapping assistance of N. Neumann is greatly appreciated.

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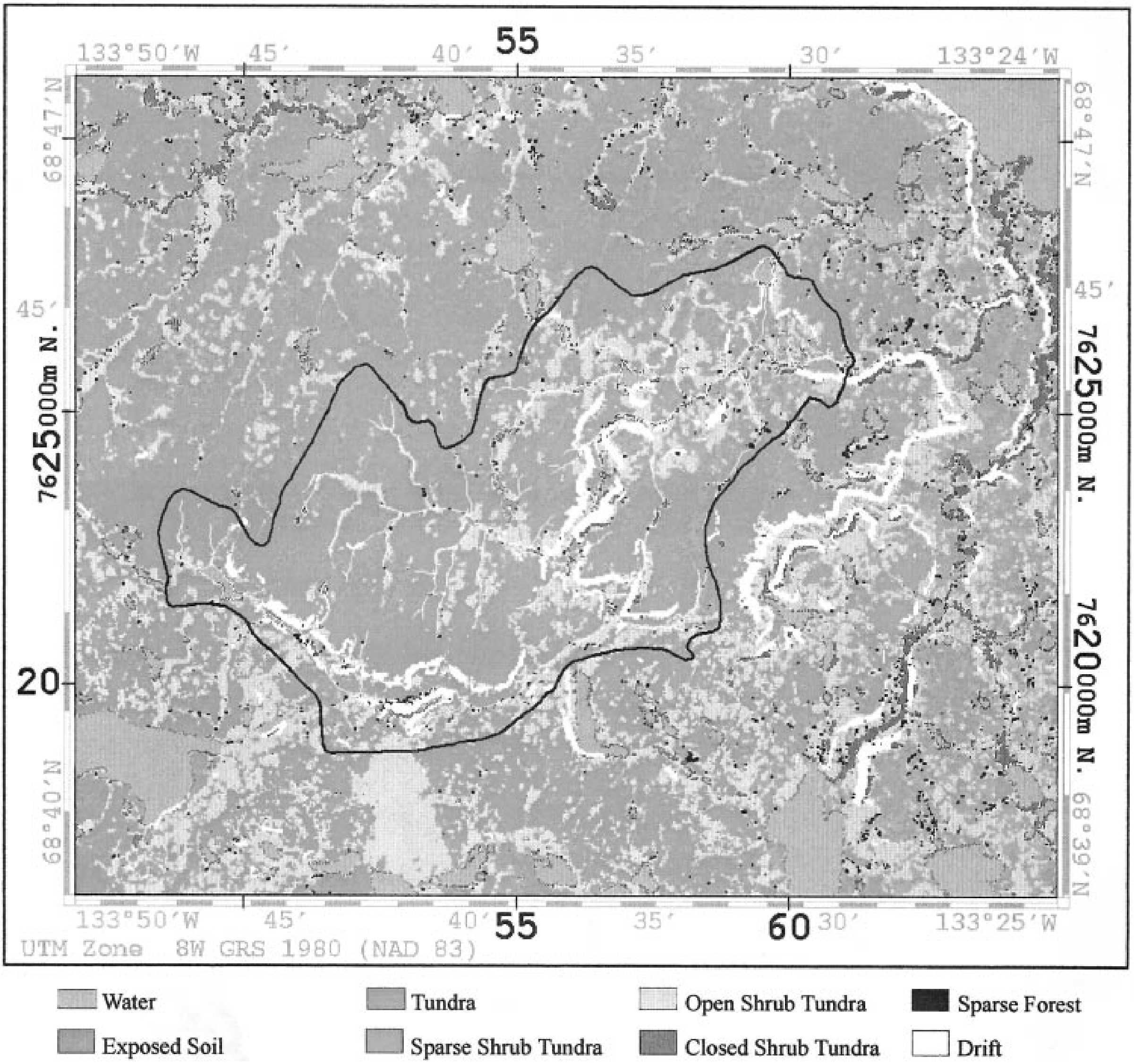

Figure 3. Landscape classification of Trail Valley Creek, Northwest Territories, Canada. The landscape was classified using a Landsat TM image and digital elevation model. Catchment outline derived from topographic maps and aerial photographs. For blowing snow modelling *tundra* includes exposed soil, tundra, water and sparse shrub tundra, *shrub–tundra* includes open and closed shrub–tundra. *Drifts* were classified by location in stream channels or by slope angle and adjacent vegetation

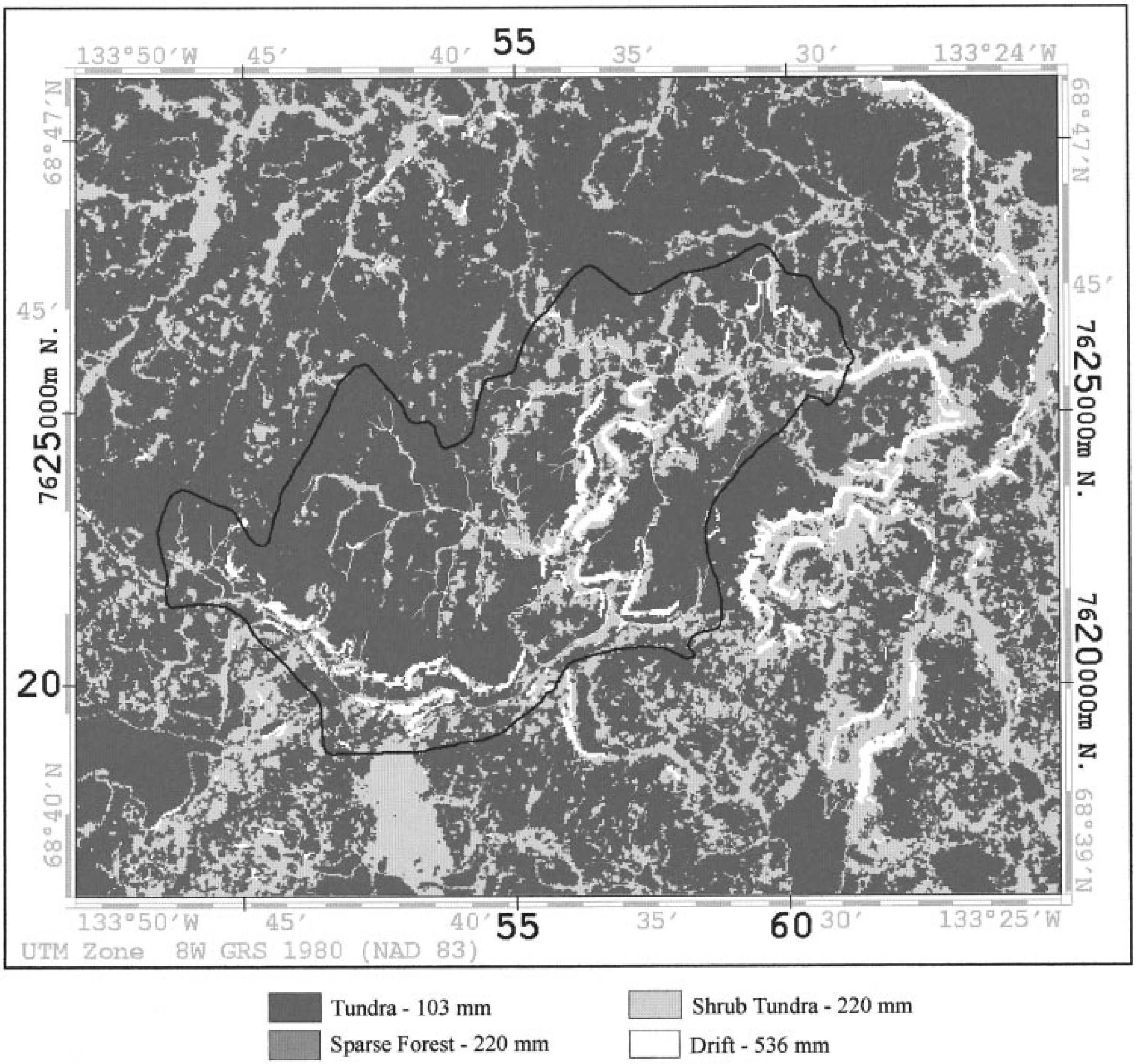

Figure 6. Pre-melt spatial distribution of snow water equivalent predicted by the DBSM for Trail Valley Creek, 1993

