THE APPLICATION OF REMOTE SENSING AND A BLOWING SNOW MODEL TO DETERMINE SNOW WATER EQUIVALENT OVER NORTHERN BASINS

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ABSTRACT
In basins along the sub-arctic/arctic transition, land cover varies dramatically from shrubs and open forest in the sub-arctic to sparsely vegetated tundra in the arctic. This change in vegetation has a strong effect on blowing snow processes and hence snow accumulation, with much less redistribution and much greater snow retention in basins covered with sub-arctic vegetation. To determine basin snow water equivalent and sublimation after redistribution of snow, a series of blowing snow algorithms was applied in a spatially-distributed model using standard meteorological data. The model uses a Landsat TM derived vegetation classification coupled with a digital elevation model to segregate the basin into sources and sinks. Thematic Mapper channels 3, 4, 5 and 7 from a mid-summer image were classified to standard ecological categories using a supervised maximum likelihood procedure on PCI software. The classification was combined with digitised elevations to provide a landscape classification of 40 m resolution. About 22% of annual snowfall sublimated during blowing snow in the arctic basin examined but only one-fifth of this value sublimated from the subarctic basin. Snow accumulation in drift areas was over 5 times that on level tundra surfaces. Snow accumulation predicted by the model compares well with the results of landscape-stratified snow surveys. The resulting snow water equivalent distribution maps show the extreme variation in snow water available for melt in small sub-catchments and the large-scale transition in snow accumulation regimes as the arctic "treeline" is crossed.

INTRODUCTION
Remote sensing of snow water equivalent has been a goal of hydrologists and atmospheric scientists for many years. The need is especially great in the remote Arctic and sub-arctic regions where snow is extremely important to the annual runoff, surface snow surveys are sparsely distributed and many basins have little or no data on snow accumulation. The cost of operating high-resolution airplane-based sensors is prohibitive in this region. However, passive microwave radiometry from satellites has been used extensively to estimate snow water equivalent of dry snow covers in both open and forested northern terrain for several
years (Goodison et al., 1990; Foster et al., 1991; Gan, 1993; Hall et al., 1991) and more recently with the BOREAS project in the northern boreal forest. Unfortunately the resolution of satellite based passive microwave is roughly 25 km or larger and snow covers deeper than 1 m are difficult to accurately measure. The high resolution of LANDSAT TM images is close to that required for many snowmelt and accumulation studies, but the low frequency of passes with clear sky conditions and inability to accurately estimate snow water equivalent in most cases has limited the use of this imagery in direct observations of snow.

The interest in the distribution of snow accumulation in the North is partially due to the inhomogeneity of the snow depth and density cause by snow redistribution processes. In the Arctic, blowing snow is the key redistribution process; further south in the coniferous forest, snow interception becomes important. Benson (1982) estimated snow relocation along the arctic coast of Alaska and found that 58% of annual snowfall remained on the tundra, 11% was transported to form drifts in a river valley and 32% was unaccounted for and presumed to have sublimated in transit. Tabler et al. (1990) in the same region found about 86 Mg of snow per m width of fence was transported by blowing snow to snow fences near Prudhoe Bay, Alaska. In the southern boreal forest, Pomeroy and Schmidt (1993) estimated that about 1/3 of annual snowfall sublimated from intercepted snow. Sublimation losses have not been quantified for northern sub-arctic forests, but snow survey records suggest that losses are small (Pomeroy et al., 1995). The result is a dramatic change in snow accumulation regime as the arctic treeline is crossed (Pomeroy et al., 1995).

Hydrological models of snow accumulation processes have recently advanced to include blowing snow transport and sublimation processes (Pomeroy et al., 1993; Pomeroy and Gray, 1994). These models suggest that snow erosion caused by blowing snow can ablate significant amounts (up to 70%) of the seasonal snow cover from open sites in cold, windswept continental regions. Not all this snow is relocated to drifts, much sublimates in transit, in general the longer the fetch the greater is the sublimation loss. On the southern Canadian Prairies sublimation of blowing snow can be from 12% to 65% of annual snowfall in open environments. Operation of models such as the Prairie Blowing Snow Model (Pomeroy et al., 1993) on relatively uniform landscapes of the prairie environment often does not require detailed information on landcover type and topography. However, in the more heterogeneous landscape of the Arctic such information may well be required for the models to produce reasonable results.

It is proposed here that the combination of detailed information gleaned from remote sensing of landscape, coupled with hydrological process models of snow accumulation and redistribution can provide the sought after spatial distribution of snow water equivalent in heterogeneous basins that lie near the arctic treeline. The use of process-based models permits an assessment of blowing snow sublimation loss on the basins, and hence an indication of the effect of landcover and topography on boundary-layer phenomena in the atmosphere. The purpose of this paper is to demonstrate the use of data from satellite imagery in a process-based snow hydrology model to estimate the snow water equivalent over a basin. In doing so, LANDSAT-TM information is used to spatially-distribute a blowing snow model to calculate the snow accumulation, sublimation, and relocation fluxes in two nearby northern basins, one arctic in character and the other sub-arctic.
DISTRIBUTED BLOWING SNOW MODEL

The Distributed Blowing Snow Model (DBSM) is a routine to calculate the transport and sublimation fluxes of blowing snow along with the resulting snow accumulation over heterogeneous basins. It contains no melt routine so verification can only be accomplished is where mid-winter melts are infrequent. A meso-scale domain (10's of km) larger than the basin of interest is modelled as an assembly of internally homogeneous landscape elements having specific vegetation, slope, exposure and fetch attributes. Blowing snow processes are calculated over the season for combinations of attributes using standard meteorological information. The DBSM distributes the landscape by dividing landscape elements into sources and sinks of snow. Mass balances are calculated for each landscape element using control volumes and the continuity equation while retaining a mass balance over the meso-scale domain (Fig. 1). A mass balance is not required for the basin of interest, permitting a net snow import or export due to relocation.

![Control Volumes for Distributed Blowing Snow Model](image)

**Figure 1.** Schematic describing fluxes and control volume concept used in the Distributed Blowing Snow Model.

Transport and sublimation fluxes are calculated on a monthly basis in this application of DBSM because incomplete datasets precluded use of a fine time resolution blowing snow model. The monthly blowing snow model uses regression relationships of mean monthly
values of wind speed, air temperature, relative humidity, snow depth on the ground and the cumulative monthly snowfall with monthly sublimation and transport fluxes (Pomeroy and Gray, 1994; 1995). A full description of the current version of DBSM along with calculation procedures is provided by Pomeroy et al. (1996).

USE OF REMOTE SENSING TO APPLY THE BLOWING SNOW MODEL.

Location
Blowing snow fluxes were calculated for two NHRI-GEWEX research basins located in north-western Canada near Inuvik, Northwest Territories. The location of the sites within the Mackenzie river basin and the local vegetation patterns are shown in Fig. 2. Trail

Figure 2. Map of Northwestern Canada, showing vegetation, Mackenzie River basin and the study sites.
Valley Creek, centred at 68°44’N 133°29’W, is a 68 km² Arctic basin located about 50 km north of Inuvik and generally north of the treeline. Havikpak Creek, centred at 68°19’N 133°30’W is a 16.5 km² "subarctic" basin located about 15 km south-east of Inuvik and generally south of the treeline. Basin elevations range from 60 to 190 m above sea level for the Trail Valley catchment (TVC) and from 60 to 260 m above sea level for the Havikpak catchment (HC). Topography differs between basins, TVC has a main valley incised east-west through a tundra plateau. The steep valley sides form very effective traps for blowing snow from the NW-SE directions. A broad lowland with scattered lakes and wetlands dominates lower HC, rising to a hilly tundra upland to the NE. The hills of upper HC limit fetch distances and provide a less effective snow trap than the deep valley of TVC. Vegetation in the region is forest-tundra transition and is restricted by wind exposure, shallow active layers and poor drainage (Bliss and Matveyeva, 1992). In TVC the high plateau is poorly-vegetated tundra while moister hillslopes and valley bottoms are shrub tundra or sparse black spruce forest. In HC the lowlands display open shrub tundra or sparse black spruce forest in the wetlands and open black spruce forest on the better-drained hills; the upland plains and hilltops are sparse tundra with shrub tundra and sparse forest on hillsides and in narrow valley bottoms.

Meteorological Data

The region has a "low" arctic climate, with snow accumulation usually commencing in October and melt occurring in late April to early May at HC and mid May to early June at TVC. Mid-winter melts are uncommon as the region is extremely cold in winter with average winter (Oct.-May) temperatures of -19°C. There is a strong gradient in winter wind speed across the region, with stronger winds near the Arctic Ocean coast to the north and much weaker winds in the relatively-sheltered Mackenzie river valley. Average winter wind speed (Oct.-May) at Tuktoyaktuk (80 km north of TVC) is 6.0 km/hour, while at the Inuvik weather station (near HC) it is only 2.5 km/hour. A strong snowfall gradient runs counter to the wind speed gradient, with mean winter snowfall of 86 mm SWE at Tuktoyaktuk and 177 mm at Inuvik. These are uncorrected values and underestimate at higher wind speeds, but even a correction for gauge undercatch for wind and trace snowfall does not raise the Tuktoyaktuk mean close to that of Inuvik (B. Goodison, pers. communication).

Data from NHRI meteorological stations in TVC and HC and from the Environment Canada Inuvik weather station were used to operate the blowing snow model over the winter of 1992-93. These data include air temperature, humidity, wind speed, depth of snow on the ground and snow accumulation. Because no wind speed measurements were available for the upland tundra portion of upper HC, and the lowland measurements at Inuvik were not representative of the wind-swept tundra, wind speeds from TVC were used to run the model for upper HC. Snowfall at TVC was reconstructed from the Inuvik monthly time series and a good estimate of annual accumulation in a small glade within a sheltered forest. Snowfall at HC was also reconstructed from the Inuvik time series and a series of monthly SWE surveys in a sparsely forested sheltered valley.

Remote Sensing Data

Vegetation classification from a LANDSAT Thematic Mapper (TM) image, coupled with a digital elevation model was used to segregate the basin areas into sources and sinks of
blowing snow. A mid-summer clear-sky LANDSAT TM image was obtained from 30 June 1992, georeferenced to NAD83 and analysed with PCI software in the NHRI GIS/Image Analysis Laboratory. A domain of several kilometres around each basin was selected for image analysis. LANDSAT channels 3, 4, 5 and 7 (red, near and infrared) were sampled, classified using a hybrid unsupervised-supervised maximum likelihood procedure (Richards, 1986) and resampled from 30 to 20 m resolution to aid in the classification of mixed pixels. The choice of a mid-summer image and red to infrared imagery was made to maximise contrast between varying vegetation categories as influenced by structure and surface temperature. The unsupervised classification was used to extract the appropriate spectral classes, resulting in approximately 25 classes for each domain. These spectral classes were then regrouped based on information from aerial photographs, maps, vegetation reports and ground truth data. Ground truth data was collected throughout the area via helicopter in the spring and early summer of 1994 and included leaf area index, species mix, canopy height and photographs of representative landscapes in the region, referenced to specific pixels on the LANDSAT image using a Global Positioning System and a laptop computer displaying the georeferenced image and the current location. The 25 classes were reduced by merging into 8 categories based on ecological criteria proposed by Strong et al. (1990). The regrouped classification was used to select training areas for a supervised classification. Large homogeneous areas with ground truth data were selected and analysed for spectral statistics over the area. The maximum likelihood procedure was used to conduct the supervised classification into the final ecological categories. To aid in the classification of mixed pixels and to smooth the classification, a 5x5 mode filter was applied. This provided a verified vegetation map of the domain surrounding each basin with less than 8% of the image unclassified.

Elevation contours and hydrologic features from 1:50,000 topographic map sheets (Energy, Mines and Resources Canada) were digitised to produce a digital elevation model (DEM) of 40-m spatial resolution. The DEM was referenced to the NAD83 vegetation classification and slopes were combined with the vegetation classification to provide a complete landscape database. Drift areas were classified from vegetation (adjacent tundra), slope (>9°), aspect (SE or NW) and lakeshore or tundra streambeds and verified from aerial photograph observations of the location of late-lying drifts in late May. The DEM elevations for both basins along with the outline of lakes and the course of streams are shown in Fig. 3, illustrating the contrasting topography discussed above.

The drift classification was added to the vegetation classification to produce a landscape classification of the domains around each of TVC and HC. This classification is mapped in Fig. 4 for each basin domain. For modelling snow accumulation, water, tundra, sparse shrub tundra, disturbed, and exposed soil categories were lumped in the tundra category, open and closed shrub tundras were lumped in the shrub tundra category, "drift" includes steep tundra slopes, the margins of lakes and tundra streambeds, sparse forests are classified as "taiga" and open forests as "forest". Because of the elevation/wind exposure difference between uplands and lowlands in HC is great, the tundra category there was divided into "upland tundra" for elevations greater than 260m and "lowland tundra" for elevations below this point; all tundra in TVC is well-exposed and can be considered "upland".

The Landsat-derived landscape classification was inspected to determine characteristic fetch distances for each landscape type in each basin. These distances were determined by measuring the distance of uniform landscape type in the primary winter wind direction, NW-SE. In redistributing transport fluxes from sources to sinks, two types of sink areas
Figure 3. Elevations and hydrological features of a) Trail Valley Creek, and b) Havikpak Creek basins and area.
are defined, dense vegetation (shrub tundra, taiga) and drifts. Based on the proportion of landscape type downwind of tundra source areas, transport fluxes were apportioned 70% to drifts and 30% to dense vegetation in TVC and 30% to drifts and 70% to dense vegetation in HC. The resulting mean fetch lengths and proportion of basin area occupied by various landscape types are shown in Table 1. It is seen that TVC is characterised primarily by long fetches of tundra (greater than 1 km) and HC is characterised primarily by long fetches of taiga. In this sense each basin is quite characteristic of conditions on either side of the treeline.
Validation Data

Extensive snow surveys were used to provide accumulation of snow values with which to validate the estimates of the DBSM. Surveys were conducted before melt commenced at the end of April, 1993 in HC and in mid May, 1993 in TVC. Transects were devised to cross characteristic areas of the landscape classification. In each landscape type, 125 depth measurements were made at 5-m spacings. Density was measured using an Environment Canada ESC-30 metric sampler at five to six points along each transect.
MODEL RESULTS AND VALIDATION

Landscape and Basin Snow Redistribution and Accumulation

The model calculates snow erosion due to transport from the fetch and sublimation during blowing snow over the fetch for standard 1-km fetches of tundra and then adjusts them for the actual tundra fetch. It then calculates snow transport as a horizontal flux to the edge of the tundra fetch. Snow accumulation on the tundra is therefore a mass balance of transport and sublimation loss with snowfall inputs. Snow transported from the tundra is apportioned to drifts and dense vegetation and distributed over the fetch as a transport deposition. Snow accumulation in these source areas is a mass balance of transport and snowfall inputs. The model is not optimised in any way, the apportionment of relocated snow driven by measured fetches and calculated fluxes.

Table 1. LANDSAT-derived landscape classification for Trail Valley Creek and Havikpak Creek, Northwest Territories, Canada.

<table>
<thead>
<tr>
<th>Landscape Type</th>
<th>Trail Valley Creek</th>
<th>Havikpak Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>% Area</td>
<td>Fetch (m)</td>
</tr>
<tr>
<td>Upland Tundra</td>
<td>69.8</td>
<td>1450</td>
</tr>
<tr>
<td>Lowland Tundra</td>
<td>0.0</td>
<td>N/A</td>
</tr>
<tr>
<td>Shrub Tundra</td>
<td>21.5</td>
<td>200</td>
</tr>
<tr>
<td>Taiga</td>
<td>0.5</td>
<td>200</td>
</tr>
<tr>
<td>Forest</td>
<td>0.0</td>
<td>N/A</td>
</tr>
<tr>
<td>Drift</td>
<td>8.2</td>
<td>80</td>
</tr>
</tbody>
</table>

The simulation of blowing snow transport fluxes for TVC and HC is shown in Fig. 3 and expressed as mm SWE per unit area of the landscape type, eroded and transported off the fetch (negative values) or transported to the fetch and deposited (positive values). The highest transport loss is for the upland tundra of HC (up to 9 mm/month) because the short fetches there permit greater removal of blowing snow from the tundra than on the long fetches of TVC. The greatest transport gain is for the drifts of TVC (up to 90 mm/month).

Figure 5. Modelled blowing snow monthly transport fluxes a) Trail Valley Creek and b) Havikpak Creek for various landscape types, 1992-93.
where the large blowing snow transport fluxes and extensive perimeter of drifts adjacent to tundra zones, permit large accumulations despite the long fetch relative to those for drifts in HC. The shrub-tundra and taiga of TVC receive significant snow from transport and deposition (up to 15 mm/month) while the transport input to these zones in HC is minimal (up to 2 mm/month).

Blowing snow sublimation fluxes are shown in Fig. 5 for all landscape types that undergo blowing snow, and are expressed as mm SWE per unit area of the landscape type, eroded and sublimated whilst blowing across the fetch. Monthly sublimation losses can exceed transport losses, with the highest values found for TVC tundra (12 mm/month and the lowest for the HC lowland tundra (< 2 mm/month). The lower sublimation losses for HC are due to its shorter tundra fetch (500-1000 m) compared to TVC (1450 m), the shorter fetch permitting proportionately more blowing snow to be transported off the fetch. There is a marked seasonality to the sublimation losses, with December, January and February being the months of greatest loss and losses in October and May being quite small.

![Figure 6. Modelled blowing snow monthly sublimation fluxes at Trail Valley and Havikpak Creeks for various landscape types, 1992-93.](image)

Monthly modelled snow accumulation is shown in Fig. 6 for the various landscape types in TVC and HC. Accumulation is expressed as mm SWE over the landscape type. Monthly accumulation amounts vary dramatically by landscape type with taiga and shrub-tundra accumulating more than tundra and drifts accumulating much more than any other landscape. Interestingly the seasonal accumulation sequence does not completely follow the blowing snow transport or sublimation sequence, with October and May being important accumulation months for all landscapes except the drifts. The upland tundras of both HC and TVC lost snow in March, when accumulation was negative due to blowing snow erosion exceeding snowfall. For both upland and lowland tundra at TVC and HC, accumulation practically ceased after February.

To evaluate the impact of the transport, sublimation and accumulation fluxes on basin hydrology, annual fluxes are presented as flux from the landscape type and mean flux from the basin in Table 2. The fluxes should be evaluated in the context of the annual snowfall (October-May) of 171 mm SWE in TVC and 154 mm SWE in HC for the winter 1992-93. The importance of sublimation and transport fluxes to basin hydrology depends on the spatial distribution of the landscape type within the basin as well as the fluxes. For instance sublimation represents a loss of 41 mm and 53 mm SWE to upland tundras in HC and TVC respectively but to the basins it represents a much smaller loss in HC (4 mm SWE), while still a respectable 37 mm SWE over TVC. This explains the nature of the change in snow accumulation regimes as the treeline is crossed. Blowing snow processes are quite active north of the treeline, with large losses to sublimation and gains to transport in TVC. The transport gain means that TVC receives a net import of blowing snow transport, though sublimation losses more than eliminate this gain. The net import of blowing snow
transport is due to the dissected topography of TVC, which forms a natural snow trap. Drift areas contribute 50 mm SWE/basin to TVC but only 5 mm to HC. By contrast, HC does not possess a large tundra valley and hence transport losses and gains are balanced. The influence of forested vegetation is strongly apparent in HC where taiga and forest contribute 74 mm SWE/basin as contrasted with 1 mm in TVC.

Table 2. Annual modelled fluxes of blowing snow sublimation and transport and snow accumulation for a) Havikpak Creek and b) Trail Valley Creek catchments, presented as point (landscape) fluxes and basin-averaged fluxes in mm snow water equivalent, 1992-93. Fluxes are presented for landscape types and as a basin total.

**a) Havikpak Creek**

<table>
<thead>
<tr>
<th>Landscape Type</th>
<th>Annual Sublimation</th>
<th>Annual Transport</th>
<th>Annual Accumulation</th>
<th>Basin Sublimation</th>
<th>Basin Transport</th>
<th>Basin Accumulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upland Tundra</td>
<td>-41</td>
<td>-39</td>
<td>73</td>
<td>-3</td>
<td>-3</td>
<td>5</td>
</tr>
<tr>
<td>Lowland Tundra</td>
<td>-6</td>
<td>-23</td>
<td>124</td>
<td>-1</td>
<td>-3</td>
<td>19</td>
</tr>
<tr>
<td>Shrub Tundra</td>
<td>0</td>
<td>9</td>
<td>156</td>
<td>0</td>
<td>3</td>
<td>45</td>
</tr>
<tr>
<td>Taiga</td>
<td>0</td>
<td>3</td>
<td>154</td>
<td>0</td>
<td>1</td>
<td>60</td>
</tr>
<tr>
<td>Drift</td>
<td>0</td>
<td>293</td>
<td>446</td>
<td>0</td>
<td>3</td>
<td>5</td>
</tr>
<tr>
<td>Forest</td>
<td>0</td>
<td>0</td>
<td>154</td>
<td>0</td>
<td>0</td>
<td>14</td>
</tr>
<tr>
<td><strong>Basin Total</strong></td>
<td><strong>-6</strong></td>
<td><strong>0</strong></td>
<td><strong>147</strong></td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>

**b) Trail Valley Creek**

<table>
<thead>
<tr>
<th>Landscape Type</th>
<th>Annual Sublimation</th>
<th>Annual Transport</th>
<th>Annual Accumulation</th>
<th>Basin Sublimation</th>
<th>Basin Transport</th>
<th>Basin Accumulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upland Tundra</td>
<td>-53</td>
<td>-34</td>
<td>84</td>
<td>-37</td>
<td>-24</td>
<td>59</td>
</tr>
<tr>
<td>Shrub Tundra</td>
<td>0</td>
<td>74</td>
<td>245</td>
<td>0</td>
<td>16</td>
<td>53</td>
</tr>
<tr>
<td>Taiga</td>
<td>0</td>
<td>74</td>
<td>245</td>
<td>0</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Drift</td>
<td>0</td>
<td>432</td>
<td>603</td>
<td>0</td>
<td>35</td>
<td>50</td>
</tr>
<tr>
<td><strong>Basin Total</strong></td>
<td><strong>-37</strong></td>
<td><strong>28</strong></td>
<td><strong>162</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The spatial variation of annual snow accumulation in HC and TVC is shown in Fig. 7, where modelled snow accumulation is mapped by landscape type in the domains containing the two basins. It is evident that the average snow accumulation in any basin or sub-basin is dependent upon the arrangement of vegetation and topography within the basin rather than mean snowfall and that substantial spatial variability in snow accumulation is evident for both arctic and subarctic landscapes. The degree of variation suggests that techniques which resolve snow water equivalent at large scales, say greater than 1 km, will be unable to detect the large accumulations of snow in narrow drifts. Particular sub-basins in the region might be dominated by either blowing snow sources or sinks with resulting dramatically different snow accumulation patterns. As runoff is largely derived from snow accumulation in the north, much of the variation in basin runoff is probably due to this snow accumulation variability at a small scale.

![Graphs showing snow accumulation for Trail Valley Creek and Havikpak Creek](image)

**Figure 7.** Modelled monthly snow accumulation of a) Trail Valley Creek and b) Havikpak Creek for various landscape types, 1992-93.

**Validation of the Model**

To test the results of the model, comparisons were made to measured snow accumulation in the basins at small and large scales. Snow surveys were conducted in April and May for HC and TVC respectively and compared to the modelled accumulations for these months in Fig. 8. It is seen that the model represents the measured pattern of snow accumulation in the two basins reasonably well. For TVC the percent difference between measurement and model is greatest at 19% for upland tundra (model overestimate of 14 mm SWE or 8% of snowfall), while for other landscape types it is less than 3%. For HC the percent difference between measurements and model is at 16% (model overestimate of 9 mm SWE or 6% of snowfall) for upland tundra, countered by -12% for lowland tundra (model underestimate of 15.5 mm SWE or 10% of snowfall) and 8% for shrub tundra (overestimate of 11 mm SWE or 7% of snowfall). Given coefficients of variation of measured snow water equivalent of 0.3, these differences in mean accumulation are not considered severe and seem to indicate a general underestimation of erosion by blowing snow. As this underestimation is not matched by an underestimation of deposition, it is more likely due to an underestimation of sublimation than that of transport if the input meteorological data have been correctly characterised for the basins.
To assess the model performance at the basin scale, modelled and measured SWE were weighted by percentage cover of the respective landscape types and summed to provide basin average SWE. The basin average and contribution of SWE by landscape type are shown in Fig. 9 for both basins. It is seen that the pattern of relative contribution of snow water by landscape has been modelled correctly and that basin averages correspond reasonably well. There is a 6% difference in basin snow accumulation between model and measurements for TVC (overestimate by 9 mm SWE or 5% of snowfall) and 1% difference for HC (1.5 mm SWE overestimate or 1% of snowfall). The degree of model overestimation of SWE seems to relate the percentage of tundra in the basin and the potential underestimation of sublimation discussed above. The differences in basin snow accumulation are surprisingly small given the potential errors in estimates of precipitation, wind speed over the basin and measured SWE over landscape types and are encouraging, though a longer time series for assessment would be desirable. In both basins, modelled and measured accumulation are only slightly less than seasonal snowfall, though the reasons are quite different. TVC acts as a snowtrap and "imports" blowing snow (28 mm SWE/basin) from outside the catchment boundary; this is not considered a "typical" situation for an arctic basin, as surrounding arctic basins will "export" snow to TVC. HC
Figure 8b. Spatial variation of annual snow accumulation predicted by DBSM in Havikpak Creek and area for spring of 1993.

does not import significant blowing snow but has a relatively small tundra area, much of which is sheltered lowland tundra, and short fetches of this tundra that limits sublimation; this is probably a very typical situation for a sub-arctic basin. For these reasons it is possible to extrapolate values from HC to other sub-arctic basins but values in the arctic should not be extrapolated.

CONCLUSIONS
Remote sensing of landcover characteristics, can, with a process-based model, help to accurately predict snow water equivalent over a basin and at small-scales. Snow accumulation in wind-swept environments varies significantly with landscape type, because wind relocation and sublimation during blowing snow reduces accumulation significantly. For the arctic and subarctic basins examined, sublimation consumed 22% of annual snowfall from the arctic basin but only 2.5% from the sub-arctic basin, though losses from upland tundra surfaces in each basin were large, at 31% and 27% of annual snowfall respectively. Import of transported blowing snow had an important impact on
Figure 9. Modelled and measured spring snow accumulation in various landscape types of a) Trail Valley Creek and b) Havikpak Creek basins.

Snow accumulation in the arctic basin, equivalent to 16% of annual snowfall, though this is due to the distinctive topography of the basin. The large blowing snow fluxes and strong variation in accumulation at small scales suggests that accurate determination of snow accumulation in the arctic and subarctic will require characterisation of basin vegetation from satellite imagery and basin topography from digitised elevations in order to apply models of blowing snow transport and sublimation. Such models can provide detailed information on snow distribution that is needed to calculate the timing and magnitude of snowmelt and the variation in meltwater production from various sub-catchments.

Figure 10. Contribution of landscape types to basin annual snow accumulation, snow water equivalent per basin area, for a) Trail Valley Creek and b) Havikpak Creek basins.
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