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Snow accumulation in forests from ground and remote-sensing data

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Abstract:

Winter-forest processes affect global and local climates. The interception-sublimation fraction (F) of snowfall in forests is a substantial part of the winter water budget (up to 40%). Climate, weather-forecast and hydrological modellers incorporate increasingly realistic surface schemes into their models, and algorithms describing snow accumulation and snow-interception sublimation are now finding their way into these schemes. Spatially variable data for calibration and verification of wintertime dynamics therefore are needed for such modelling schemes. The value of F was determined from snow courses in open and forested areas in Hokkaido, Japan. The value of F was related to species and canopy-structure measures such as closure, sky-view fraction (*SVF*) and leaf-area index (*LAI*). Forest structure was deduced from fish-eye photographs. The value of F showed a strong linear correlation to structure: $F = 0{\cdot}44 - 0{\cdot}6 \times SVF$ for $SVF < 0{\cdot}72$ and $F = 0$ for $SVF > 0{\cdot}72$, and $F = 0{\cdot}11\ LAI$. These relationships seemed valid for evergreen conifers, larch trees, alder, birch and mixed deciduous stands. Forest snow accumulation (S_F) could be estimated from snowfall in open fields (S_o) and to *LAI* according to $S_F = S_o\ (1 - 0{\cdot}11\ LAI)$ as well as from *SVF* according to $S_F = S_o\ (0{\cdot}56 + 0{\cdot}6\ SVF)$ for $SVF < 0{\cdot}72$. The value of S_F was equal to S_o for *SVF* values above 0·72. The value of sky-view fraction was correlated to the normalized difference snow index (*NDSI*) using a Landsat-TM image for observation plots exceeding 1 ha. Variables F and S_F were related to *NDSI* for these plots according to: $F = -0{\cdot}37 NDSI + 0{\cdot}29$ and $S_F = S_o\ (0{\cdot}81 + 0{\cdot}37 NDSI)$. These relationships are somewhat hypothetical because plot-size limitation only allowed one sparse-forest observation of *NDSI* to be used. There is, therefore, a need to confirm these relationships with further studies. Copyright © 2004 John Wiley & Sons, Ltd.

KEY WORDS evaporation; sublimation; forest; interception; Landsat; remote sensing; *NDSI*; snow; *SVF*; snow course

INTRODUCTION AND GOAL

The importance of snow and forest for climate has been highlighted by, for example, Thomas and Rowntree (1992), Bonan *et al.* (1992), Betts *et al.* (1996) and Harding *et al.* (2001). Boreal forests broadly encircle the globe between 48 °N and 65 °N occupying 12×10^6 km^2 (Harding *et al.*, 2001). These forests mainly consist of evergreen coniferous trees that intercept snow throughout the winter. Compared with snow on the ground, snow sublimates quicker in forest canopies because of greater absorption of short-wave radiation by the canopy and a higher exposure to turbulent-exchange forces. Several studies during the last decade show snow-accumulation reductions in forests compared with open areas of up to 40% caused by sublimation of intercepted snow (e.g. Lundberg and Halldin, 2001). Sublimation of snow precipitation in forests is thus a substantial part of the winter water budget.

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The terms evaporation and sublimation can be sometimes be confusing in snow research. Sublimation is ambiguous because it is used both for conversion of vapour into solid phase and *vice versa*. The term evaporation is used for the conversion of liquid and solid to gas phase. Sublimation is used by many snow interception researchers to designate the vaporization of snowfall, so this denomination has been adopted here even though some vaporization may include evaporation of meltwater. Interception sublimation (I) (millimetres per ground area) is used to designate the vaporization of snowfall caught by canopy branches. We use the interception-sublimation fraction (F) ($F = I/S_O$) in this study to simplify comparison between areas with different open-area snow accumulation (S_O).

Many operational models (hydrological, weather-forecast, and global- and regional-climate) parameterize snow processes, particularly snow-interception processes, in a very (sometimes overly) simplified way (e.g. Harding and Pomeroy, 1996; Yang *et al*., 1998; Nakai *et al*., 1999a; Eklund *et al*., 2000). During recent years, snow models that include snow-interception processes have been developed and tested by, for example, Hedstrom and Pomeroy (1998), Pomeroy *et al*. (1998a,b), Nakai *et al*. (1999a,b), Parviainen and Pomeroy (2000), Essery and Pomeroy, (2001), Gusev and Nasonova, (2001), Koivusalo and Kokkonen (2002), Pomeroy *et al*. (2002) and Gustafsson *et al*. (2003). Lettenmaier and Bowling (2001) present a comparison of high-latitude hydrological models within the PILP2 programme of GEWEX. Application of advanced snow parameterizations to many operational models could improve their wintertime performance. The algorithm to calculate surface reflectance over a snow-covered area, for example, significantly increased the accuracy in prediction of surface temperature in the High Resolution Local Area Model (HIRLAM), operated by many European weather services (Järvenoja and Lönnberg, 1995). The importance of forest albedo on temperature forecasts is illustrated by Betts and Ball (1997). A relationship between snow accumulation or F and a remotely sensed canopy-structure index would facilitate the inclusion of physically sound winter algorithms in hydrological, weather-forecast and climate models.

The boreal forest landscape is a mosaic of different forest types, densities and species. Advanced parameterizations, therefore require distributed information on canopy properties. Canopy structure is important in controlling the interception-sublimation process. The value of F seems to increase with canopy closure (C_C) according to Kuzmin (1963), Pomeroy *et al*. (2002), and Lundberg and Koivusalu (2002). Pomeroy and Goodison (1997) report that snow accumulation decreases with increasing leaf-area index (*LAI*). Hedstrom and Pomeroy (1998), Pomeroy *et al*. (1998a) and Pomeroy *et al*. (2002) report similar results. The latter also suggest a unique relationship between C_C and *LAI* such that they would be exchangeable measures of canopy structure. We are not aware of any study, except the one by Kuzmin (1963), that investigates the influence of species on F.

Different remote sensing (RS) techniques have been applied to determine canopy structure, given as *LAI*, C_C or sky-view fraction (*SVF*). Several studies present forest *LAI* classified from different RS image types (e.g. Price and Bausch, 1995; Chen *et al*., 1997; White *et al*., 1997; Bicheron and Leroy, 1999; Zhang *et al*., 2000; Eklundh *et al*., 2001). Remote sensing techniques also have been used to separate snow-covered and snow-free ground from different land covers. Kangas *et al*. (2001) separate winter 'forests' (coniferous and mixed) from both 'transitional land' (woodland scrub, broad-leaved forest and discontinuous urban fabric) and 'open land' (rivers, lakes, arable land, peat bogs and bare rock) with satellite broadband reflectance. Kurvonen *et al*. (1998) determine the forest-coverage fraction from microwave data. The normalized difference snow index (NDSI) has been used increasingly to separate snow-covered from snow-free ground (Hall *et al*., 1995; Gomez-Landesa *et al*., 2001). This index also has been used for vegetated surfaces, i.e. as a way to distinguish snow-covered from snow-free ground below a snow-free canopy. Klein *et al*. (1998) report decreasing NDSI with increasing canopy cover for forests with snow-covered ground and defoliated deciduous trees.

This paper aims at using the extensive snow-survey data set of Nakai (1996), published in Japanese, to elaborate on the influence of both species and canopy structure on snow accumulation and F. Assuming a sufficiently good relationship, our second objective is to study the possible use of NDSI as a way to map wintertime canopy structure and, thus, indirectly snow accumulation and F.

DATA AND METHODS

Nakai, 1996) made snow courses between 1991 and 1995 in forests with different species and densities at five locations in Hokkaido, Japan. The snow courses were used to estimate I. Along with the snow surveys, measurements of SVF were made with fish-eye photographs. As Pomeroy *et al.* (2002) show that winter LAI and C_C are exchangeable and SVF is just the complement to C_C, we have selected to use SVF as our main measure of canopy structure.

Determination of F from snow courses

Interception, I, can be determined by measuring snow accumulation in open areas (S_O) and forests (S_F) before the melt period begins (e.g. Johnson, 1990; Nakai, 1996; Lundberg and Koivusalo, 2002). The technique is based upon a number of assumptions (see Lundberg and Halldin, 2001). Hereafter, the subscript O is used to designate open and F for forested areas. Accumulation in open areas can be calculated by

$$S_O = P_O - E_{SO} - M_{BO} - M_{SO} - R_S + R_C \quad (1)$$

where P is snowfall, E_S is sublimation from the snow-cover surface, M_B and M_S stand for meltwater that leaves the snow cover because of base melt and surface melt, respectively, and R_S and R_C represent snow relocation between forest and open areas (R_S is snow blown from open areas into forests, R_C is snow blown from canopy branches into open areas). The terms in Equations (1) to (4) are expressed in millimetres per ground area.

Accumulation in a forested area can be calculated by

$$S_F = P_F - \Delta C - E_{SF} - I - M_{BF} - M_{SF} - M_C + R_S - R_C \quad (2)$$

where ΔC is the increase in canopy snow storage during the period investigated and M_C is melted snow from canopy branches that enters the soil. The difference between S_O and S_F from Equations (1) and (2) becomes

$$S_O - S_F = P_O - P_F + \Delta C + E_{SF} - E_{SO} + I + M_{BF} - M_{BO} + M_{SF} - M_{SO} + M_C - 2R_S + 2R_C \quad (3)$$

Equation (3) was simplified using the following assumptions:.

1. $P_F = P_O$. Possible differences in precipitation between open areas and forests cannot be dismissed directly. Gary (1974) shows increased deposition at the windward side of small clearings and decreased deposition at the forest edge of the leeward side. Differences in deposition are attributed to local variations in wind speed. Dry, light snow is very sensitive to changes in wind speed whereas wet, heavy snow is rather insensitive. The maritime climate of Hokkaido with few periods of dry, cold snow in combination with large open field plots justifies the assumption in this study.
2. $\Delta C = 0$. Canopy branches were bare at the beginning and end of the measurement period; accordingly no change in ΔC had to be considered.
3. $M_{BF} = M_{BO}$. In Hokkaido, melt at the snow-cover base because of ground heat is reasonably constant with values around 0·3 to 1·0 mm day^{-1} for snow covers larger than 0·5 m. No indications of differences in base melt between forest and fields were found.
4. $M_{SF} = M_{SO} = 0$. Surface melt can be expected to differ between forest and fields, primarily because of differences in net radiation. Surface melt cannot be excluded if air temperatures exceed $-3\,^{\circ}$C, and warmer temperatures are quite frequent in this area. We therefore used data from winters only when the cumulative melt before refreezing was small. This screening was done by inspection of cumulative degree-days and of snowpack cross-sections on the dates of the snow-surveys. No traces of meltwater leaving the snow pack were observed.
5. $M_C = 0$. Melted snow, M_C, can be channelled as stemflow or drip down onto the snow pack. The value of M_C entering the soil was assumed negligible because winters with large melts were excluded from the

calculations and most stemflow studies report low stemflow rates. Delfs (1967), who measured in mature spruce stands, registered only 0·8% of precipitation as stemflow and Kominami (personal communication, 1998) did not observe any winter stemflow in Hokkaido. Any M_C channelled as drip is accounted for by the snow course measurements.

6. $R_C = R_S = 0$. Because of the maritime climate, dry, light snow is uncommon and observations of snow blown from canopy branches (R_C) at the sites were very rare. No observations of R_S were made at the sites.

With these assumptions Equation (3) becomes

$$I = (S_O - S_F) - (E_{SF} - E_{SO}) \tag{4}$$

If the difference between E_{SF} and E_{SO} also could be neglected, then I could be estimated as the difference between S_O and S_F. The order of magnitude of the error introduced by assuming $E_{SF} = E_{SO}$ can be estimated from measurements of E_{SO} and E_{SF}. From Hokkaido, only measurements of E_{SO} (<25 mm) are available. Doty and Johnson (1969) estimated the ratio E_{SF}/E_{SO} to 0·45, whereas Bernier and Swanson (1993) estimated it to 0·25. If E_{SO} is assumed to be 20 mm, and $E_{SF}/E_{SO}0{\cdot}35$, E_{SF} becomes 7 mm and the error in I becomes *c.* $7 - 20 = -13$ mm.

With the above simplifications, F can be estimated

$$F = \frac{I}{S_O} = \frac{S_O - S_F}{S_O} = 1 - \frac{S_F}{S_O} \text{ or } S_F = S_O(1 - F) \tag{5}$$

Kuzmin (1963) presented a linear relationship between S_F/S_O and C_C (0 to 1) based on data from Valdai in Russia

$$\frac{S_F}{S_O} = 1 - C_1 C_C \tag{6}$$

Equation (6) can be transformed into a relationship between F and SVF by substituting C_C with $1 - SVF$ and combining Equations (5) and (6)

$$F = C_1(1 - SVF) \tag{7}$$

Kuzmin (1963) report different values of C_1 for fir (0·37) and pine (0·22) forests. Pomeroy *et al.* (2002) report $C_1 = 0{\cdot}43$ for a variety of forest stands (mainly coniferous, but including one aspen-dominated stand) in western Canada.

Sites

Measurements at five locations in south-west Hokkaido (Figure 1) were used for the analysis. The average altitude ranged from 65 m a. s. l. for site N to 370 m a. s. l. for site M (Table I). The area has a maritime climate with high precipitation rates and rather mild winters. The average December–March temperature was $-3{\cdot}8\,^{\circ}$C and the incoming short-wave radiation was approximately 250 (MJ m^{-2} $month^{-1}$) during the study period. S_O in the region increases from west (around 250 mm at sites N, H and S) to east (around 500 mm at site M) (Table I, average S_O of years 1993 and 1995). The seasonal variation in S_O is much smaller than the difference between sites. S_O in J varied, for example, between 360 mm and 411 mm (Table I) during the three observation years.

Plot sizes ranged from small (<1·0 ha) at H and J to large (8 to 27 ha) at S (Figure 2 and Table II). All forest stands (except the deciduous in S) were uniform plantations (Table I). More than half the stands were young, around 20 years, and the oldest one was 66 years (Table I). Stand density varied from 500 to 3300 stems ha^{-1} and average tree height ranged from 7 to 25 m. Most stands were coniferous (seven spruce, four fir, and two pine), but a few deciduous and larch forests were included. Site H had the largest number (10) of plots, with SVF ranging from 5% to 63% (Table I).

Table I. Characteristics of forest plots at sites Hitujigaoka (H), Jyozankei (J), Nopporo (N), Shikotsukohan (S) and Makkari (M): plot size, sky-view fraction (*SVF*), normalized difference snow index (*NDSI*) and leaf-area index (*LAI*; calculated by Equation 8); interception sublimation (I) and interception-sublimation fraction (F) of snowfall in open areas (S_O); canopy characteristics (species, average stem diameter, number of stems per ha, age and average height) (modified from Nakai, 1996)

Plot	Size (ha)	*SVF*	*NDSI*	*LAI*		Interception sublimation I (mm winter^{-1}) 1991	I 1992	I 1993	I 1995	F (mm winter^{-1}) 1991	F 1992	F 1993	F 1995	Species[a]	Diameter (cm)	Stems (no. ha^{-1})	Age years	Height (m)
					S_O =	341	247	245	231									
H	0·20	0·07	0·13	3·71		123	114	122	122	0·36	0·46	0·50	0·53	Fir, As(1)	10	3200	20	8
	0·18	0·06	0·58	3·84			113	92	99		0·46	0·38	0·43	Fir, As(2)	12	3300	20	8
	0·30	0·06	0·59	3·84		128	127	111	111	0·38	0·51	0·45	0·48	Pine, Pb	13	1900	19	8
	0·21	0·05	0·15	3·97			96	107	100		0·39	0·44	0·43	Pine, Pp	12	1500	20	7
	0·17	0·09	0·71	3·46			92	96	85		0·37	0·39	0·37	Spruce, Pg	12	3000	20	7
	0·18	0·10	0·12	3·34			81	83	72		0·33	0·34	0·31	Spruce, Pj	12	2900	20	7
	1·00	0·21	0·48	2·29			95	83	73		0·38	0·34	0·32	Larch, Lg	13	2100	20	10
	0·97	0·21	0·44	2·29		52	72	83	59	0·15	0·29	0·34	0·26	Larch, Ll	14	1600	20	12
	0·50	0·31	0·37	1·62		51	66	63	60	0·15	0·27	0·26	0·26	Larch, Ll	16	900	20	12
	0·12	0·63	0·45	0·54			11	3	3		0·04	0·01	0·01	Birch, Bm	14	1900	20	15
					S_O =		360	370	411									
J	0·63	0·10	−0·03	3·34			131	122	155		0·36	0·33	0·38	Fir, As(1)	14	1300	21	12
					S_O =			274	259									
N	4·18	0·09	−0·11	3·46				78	81			0·28	0·31	Fir, As(2)	14	2100	24	12
					S_O =			261	241									
S	14·27	0·11	0·08	3·23				78	95			0·30	0·39	Spruce, Pj(1)	11	2600	35	12
	14·74	0·11	−0·36	3·23				71	93			0·27	0·39	Spruce, Pj (2)	13	2100	35	12
	24·26	0·13	−0·20	3·01				72	77			0·28	0·32	Spruce, Pj (3)	29	620	62	14
	27·44	0·12	−0·10	3·12				101	108			0·39	0·45	Spruce, Pj (4)	32	720	66	18
	8·17	0·60	0·03	0·60				18	28			0·07	0·12	Decidu., Df	25	500	Natural[c]	23
					S_O =			488	534									
M	1·44	0·12	0·53	3·46				143	168			0·29	0·31	Spruce, As	36	620	62	25
	0·58	0·62	0·8	3·12				19	23			0·04	0·04	Alder, Ah	21	670	33	22

[a] As, *Abies sachaliensis* Masters; Pj, *Picea jezoensis* Carr.; Pg, *Picea glehnii* Masters; Ll, *Laris leptolepsis* Gordon; Lg, *Laris gmelini* var *japonica* Pilg.; Pp, *Pinus parviflora* var *pehtaphylla* Henry; Pb, *Pinus baksiana* Lamb.; Bm, *Betula maximowicziana* Regel; Ah, *Alnus hirsuta* Turcz.; Df, Deciduous broad leaf species.
[b] Natural regeneration.

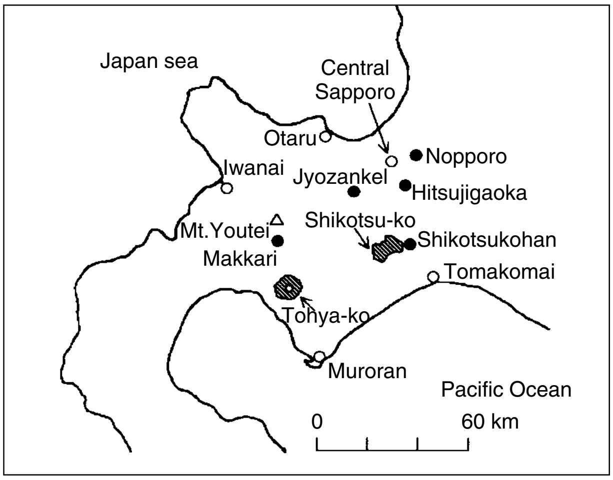

Figure 1. Location of sites (black dots): Hitujigaoka (H), Jyozankei (J), Nopporo (N), Shikotsukohan (S) and Makkari (M) (from Nakai, 1996)

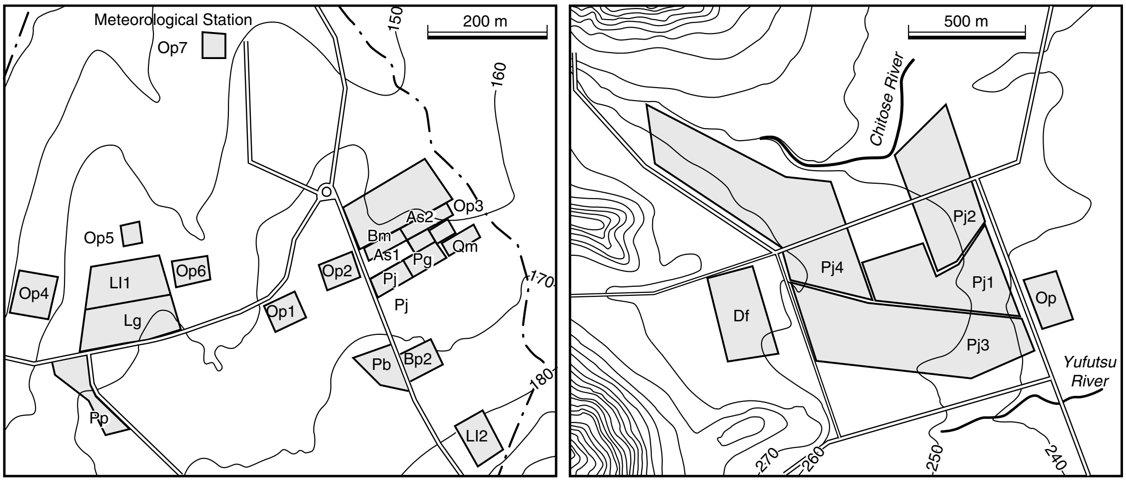

Figure 2. Plot sizes at Hitujigaoka (left) and Shikotsukohan (right). Note the different scales. Op denotes open areas and other notations are given in Table I

Table II. The normalized difference snow index (*NDSI*) values for the open sites, together with plot area, altitude and location (modified from Nakai, 1996)

Site	*NDSI*	Plot area (ha)		Altitude (m a.s.l.)	Latitude (N)	Longitude (E)
		Open sites	Forests sites			
H (Hitujigaoka)	0·32–0·83	0·13–0·53	0·12–1·00	160	42°58′	141°23′
J (Jyouzankei)	0·77	0·61	0·63	300	42°58′	141°10′
N (Nopporo)	0·81	0·96	4·14	65	43°02′	141°30′
S (Shikotsuko)	0·72	3·09	8.17–27.44	250	42°45′	141°26′
M (Makkari)	0·71	1·42	0.58–1.44	370	42°48′	140°47

Snow courses

Open-area snow accumulation, S_O, was measured in one or several observation plots at each site and S_F was measured in stands with documented *SVF*. The snow courses were made at the end of the February or beginning of March every year from 1991 to 1995. Forest snow accumulation, S_F, was determined using a grid of measurement points in the stands, where snow depth varied greatly (little snow close to the stems and much snow in the gaps between trees). Each stand was divided into quadrants limited by four stems. Snow depth and density were measured along the diagonals in three to five quadrants per stand. The distance between each measurement point was adjusted to stem spacing. Snow depth was measured every 0·1 to 0·2 m along both diagonals and density was measured every 0·2 to 0·3 m along one diagonal, meaning that three to five S_F values were obtained for each quadrant.

Sky-view fraction and leaf-area index

Ground-based measurements of *SVF* were made in each stand during the autumns of 1991 and 1992 when the deciduous trees had defoliated. The measurements were made during days with no direct solar radiation. The photographs were taken with a Nikon Nikkormat camera with a fisheye lens with focal length (f) of 8 mm and a maximum aperture of $f/2{\cdot}8$. Grey-level thresholding was used to segment the images into two-value images. In this study, we have expressed canopy structure also as *LAI* through the relationship of Pomeroy *et al*. (2002). This relationship is based on data from several different coniferous stands (jack pine, black spruce) and a mixed-wood stand (75% aspen; 25% white spruce), and is assumed valid for $LAI > 0{\cdot}15$

$$C_C = 0{\cdot}29\ln(LAI) + 0{\cdot}55 \tag{8}$$

This is transformed into a relationship between *LAI* and *SVF* by using $SVF = 1 - C_C$

$$LAI = e^{(0{\cdot}45 - SVF)/0{\cdot}29} \tag{9}$$

Normalized difference snow index

A Landsat-TM image was used for the satellite-based estimates of *SVF*. Landsat TM was chosen because its spectral channel number 5 has a good absorption band for snow. The pixel size of the sensor, 30 × 30 m, gives a reasonably good spatial resolution and data are readily available.

We were interested in winter *SVF,* so only periods with defoliated deciduous trees were considered. Images were considered only from 1993 to 1995 in order to assure that *SVF* had not changed much from the time of the ground-based measurements. We wanted snow-covered ground and bare canopies to simplify separation of canopy and ground. Solar angles had to exceed a certain limit to avoid interpretation of images with insufficient amounts of reflected radiation. A time span from the end of February to the beginning of March was considered suitable to meet these criteria. Air temperature and precipitation information on canopy storage from snow courses a few days earlier as well as canopy photographs for sites S and H were used to quantitatively estimate the amount of canopy snow at the date of the analysed image. We settled for 9 March 1993, a day with a very light cloud cover over the Sea of Japan and almost no clouds over Hokkaido. On this day, canopies at H and N were probably snow-free and there was very little snow on the canopies at the sites S and M. The canopy at the site J may have stored up to 10 mm per ground area.

The Sea of Japan and the Pacific Ocean as well as the largest lakes at Hokkaido were scanned for the darkest pixels. All subareas of the scene showed roughly the same values for the darkest pixels, indicating homogeneous atmospheric conditions throughout the area. Between different parts of the scene, the variations in the darkest pixel values for channels 2 and 5 were within one or two DN (a dimensionless unit), and were of about the same magnitude as the stripes caused by the detector differences. A representatively dark pixel value for the entire scene was estimated to 17 DN for channel 2 and 4 DN for channel 5. The image was corrected for atmospheric influences by subtracting these values from all pixels, assuming zero reflectance at the darkest pixels. Image data from the surroundings of the snow-course locations were cut from the scene. These smaller images were geometrically rectified to topographic maps with the help of ground control points (GCPs). The normalized difference snow index (*NDSI*) was calculated for the subimages using (Hall *et al.*, 1995)

$$NDSI = (\text{channel } 2 - \text{channel } 5)/(\text{channel } 2 + \text{channel } 5) \tag{10}$$

As snow absorbs short-wave infrared radiation at channel 5 and has high reflectance at visible wavelengths such as channel 2, high *NDSI* values are expected. Slightly higher reflectance is expected for channel 5 compared with channel 2 for spruce or pine forest, thus giving slightly negative *NDSI* values. The *NDSI* values were extracted from the image data for the different plots at the snow-course sites. Average values were calculated without any obvious outliers.

RESULT

Sky-view fraction varied with forest type from 0·05 to 0·13 in coniferous stands, between 0·21 and 0·31 in larch stands and up to approximately 0·60 in alder, birch and mixed deciduous stands (Table I and Figure 4). Light conditions were rather variable during the *SVF* measurements and the margin of error was estimated to ±0·05 *SVF* units.

The maximum sublimation was 168 mm for site M in 1995 and the maximum sublimation fraction was 53% at site H also in 1995 (Table I). The relationship between *F* and *SVF* was similar for all years (Figure 3a and b) except for 1994, which showed a deviating pattern. This year was omitted from the analysis because a substantial mid-winter surface melt exceeded the water-holding content of the snowpack.

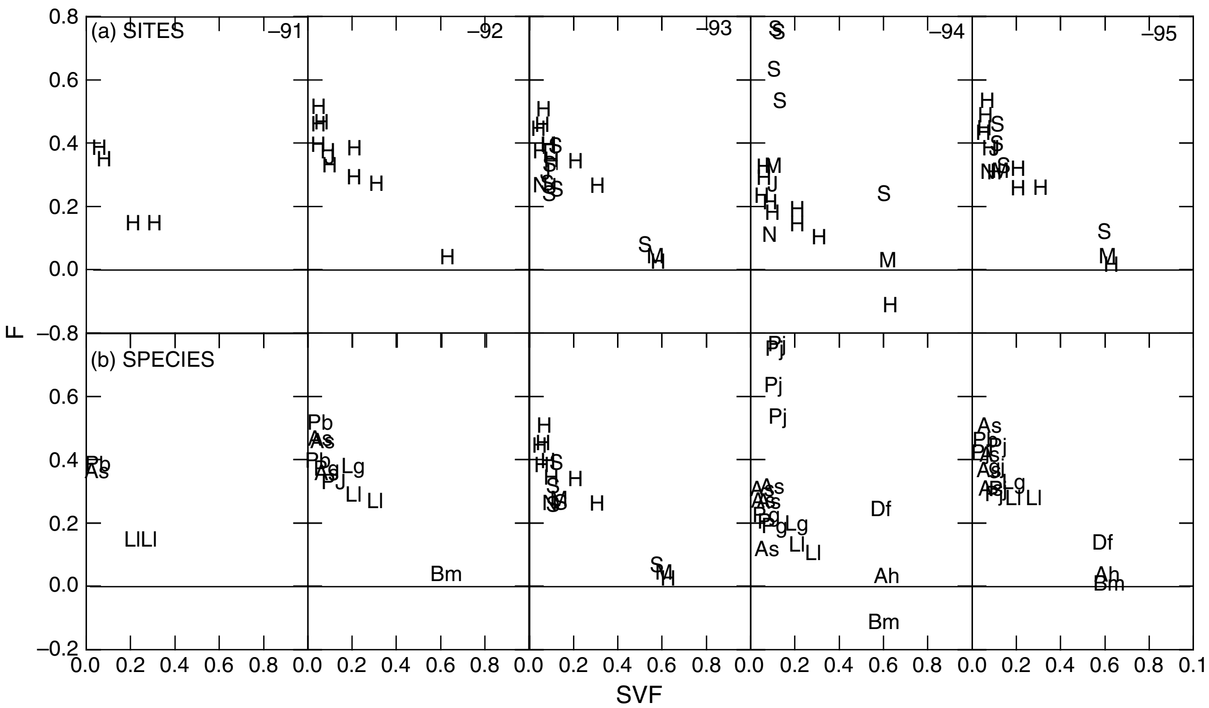

Figure 3. Sublimation fraction, *F*, versus sky-view fraction, *SVF*. The data are winter-season averages from years 1991 to 1995 based on data from Nakai (1996): (a) separated according to sites (H, Hitujigaoka; J, Jyozankei; N, Nopporo; S, Shikotsukohan; M, Makkari); (b) according to species (notations given in Table I)

The F value for fully closed stands was 0·44 (Figure 4) and ≈0·05 for open stands with SVF ≈0·60. No detectable sublimation seemed to take place for stands with SVF >0·72. A unique relationship between F and SVF seemed to hold both for coniferous forests, larch trees and deciduous forests

$$F = 0{\cdot}44 - 0{\cdot}6 \times SVF \quad \text{for } SVF < 0{\cdot}72; F = 0 \text{ for } SVF > 0{\cdot}72 \tag{11}$$

The relationship between F and LAI (Figure 5) seemed equally good

$$F = 0{\cdot}11LAI \tag{12}$$

The value of S_F could, thus, be estimated from S_O and SVF or LAI (Equations 5 and 11/12)

$$S_F = S_O(0{\cdot}56 + 0{\cdot}6 \times SVF) \text{ for } SVF < 0{\cdot}72;\ S_F = S_O \text{ for } SVF > 0{\cdot}72 \tag{13}$$

Estimated values of S_F agreed rather well with measured values (Figure 6a; $R^2 = 0{\cdot}86$)

$$S_F = S_o(1 - 0{\cdot}11LAI) \tag{14}$$

The correlation between estimated and measured S_F values became marginally better with this approach (Figure 6b; $R^2 = 0{\cdot}91$). The average deviation between measured and calculated S_F was 14 mm for both methods and the maximum deviation was 53 mm for estimates based on SVF and 40 mm for estimates based on LAI (Figure 7a). The errors in the S_F estimates seemed to be random (Figure 7a and c). The average relative error was 7–8 mm for both methods and the maximum relative error was 24 mm for both methods.

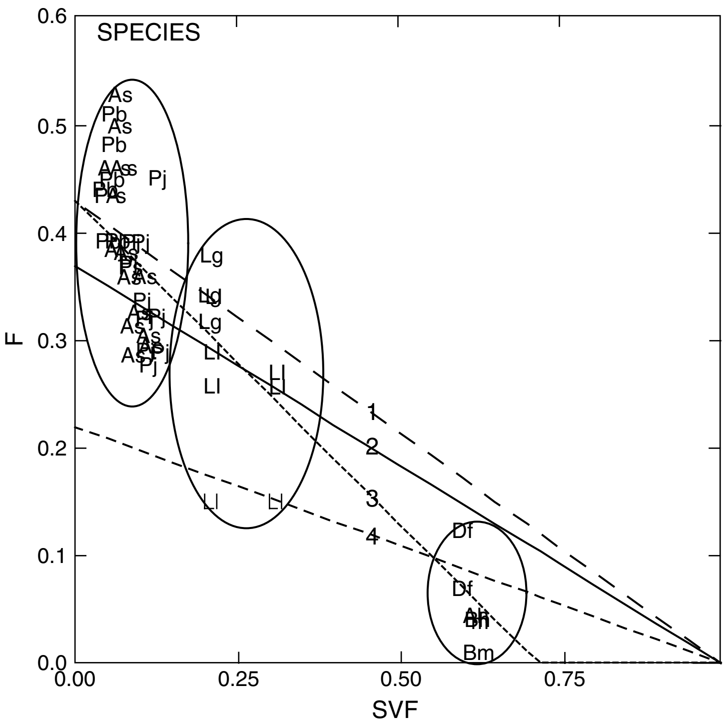

Figure 4. a) Sublimation fraction, F, versus sky-view fraction, SVF, for the years 1991–1993 and 1995 separated according to species (notations given in Table I). Lines 1, 2 and 4 are Equation (7) with $C_1 = 0{\cdot}43$ (Pomeroy *et al.* (2002), mainly coniferous forests), 0·37 (Kuzmin, (1963), fir) and 0·22 (Kuzmin (1963), pine), respectively. Line 3 is Equation (11) ($R^2 = 0{\cdot}76$) from this study. Ellipses encircle groups of species: evergreen conifers to the left, larch in the middle and deciduous trees to the right

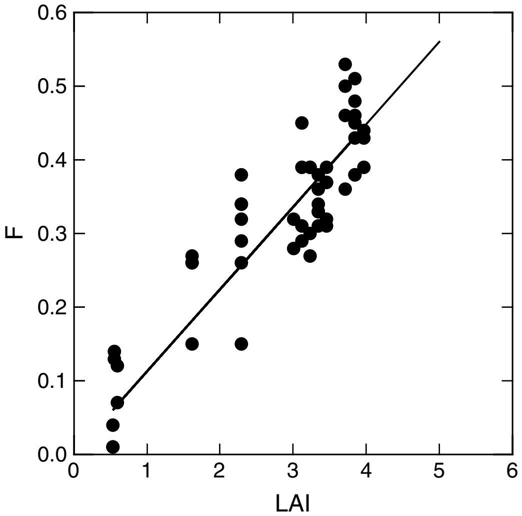

Figure 5. Sublimation fraction, F, versus *LAI* for the years 1991–1993 and 1995. $F = 0{\cdot}11$ *LAI*. $R^2 = 0{\cdot}75$ with the regression line forced through origin

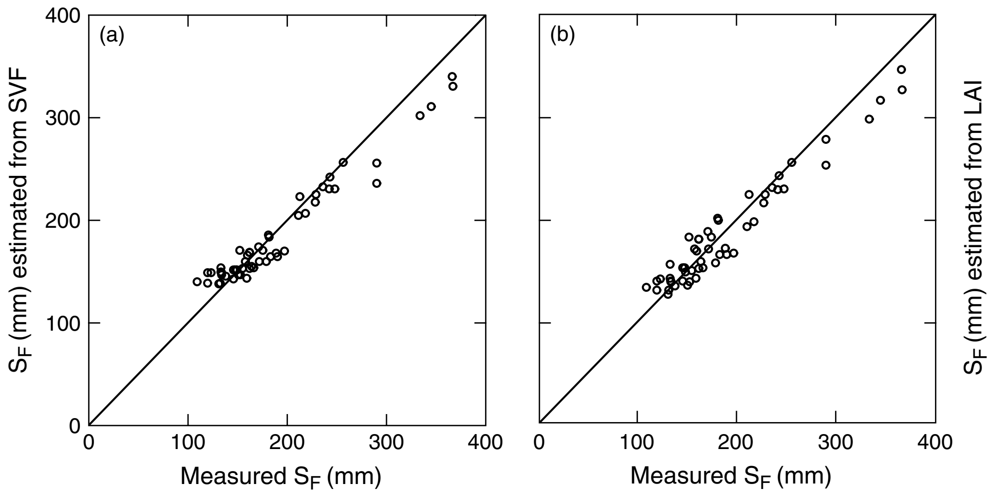

Figure 6. Calculated versus measured S_F. The slope 1 : 1 is shown as comparison. (a) Calculated from *SVF* (Equation 11). $R^2 = 0{\cdot}86$ with regression forced through origin. (b) Calculated from *LAI* (Equation 14). $R^2 = 0{\cdot}91$ with regression forced through origin

A plot of *SVF* versus *NDSI* for site H (plot sizes <1 ha) showed no clear relationship between *NDSI* and *SVF* (Figure 8a). The dimensions (length and width) of these plots were of the same order of magnitude as the accuracy of the determination of the *GCP*s, so reliable *NDSI* data could not be acquired for them. Observations from the forested plots at M (plot sizes between 0·6 and 1·4 ha) did not show any clear relationship between *NDSI* and *SVF* (Figure 8a). The snow-course plots at site S were clearly larger than at the other sites, whereas the plots at N and J were roughly the same size as M (Table II). For the S, J and N sites, with larger plots, the *NDSI* increased with *SVF* from approximately −0·2 for a closed forest to approximately 0·8 for an open field (Figure 8b). The *NDSI* values for the open plots were rather consistent (varying between 0·72 and 0·81), whereas the values for the dense forests with *SVF* ranging between 0·09 and 0·13 varied more (−0·36 to

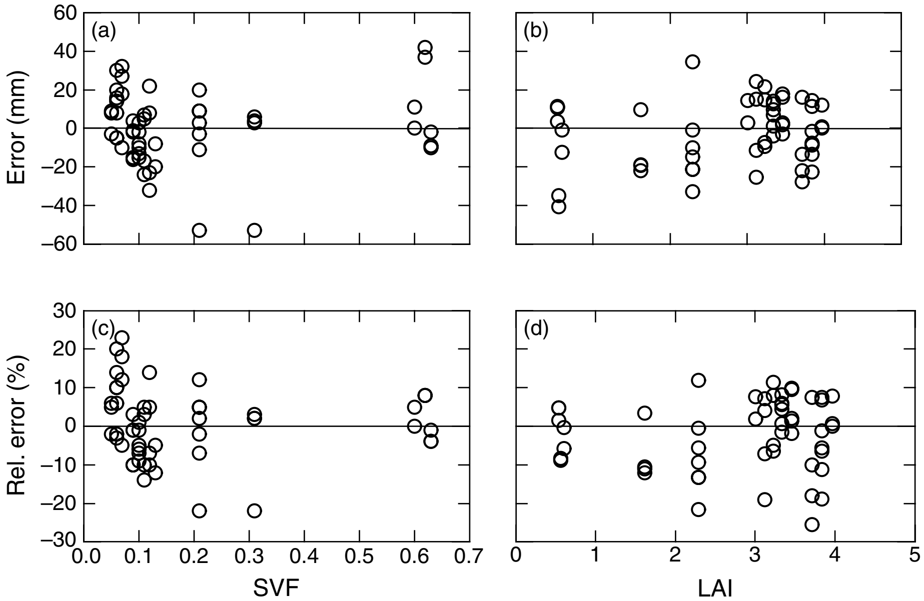

Figure 7. Error and relative error ($S_{\text{FCalculated}} - S_{\text{FMeasured}}/S_{\text{Fmeasured}}$) in estimate of S_F. (a) Error in S_F, estimated from *SVF*, plotted against *SVF*. (b) Relative error in S_F, estimated from *SVF*, plotted against *SVF*. (c) Error in S_F, estimated from *LAI*, plotted against *LAI*. (d) Relative error in S_F, estimated from *LAI*, plotted against *LAI*

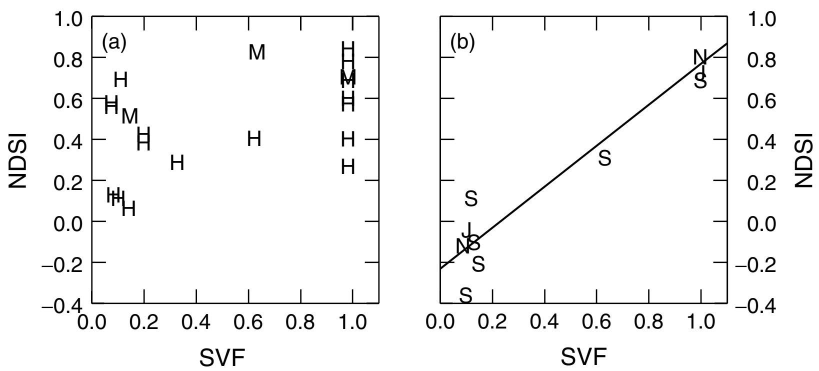

Figure 8. Sky-view fraction, *SVF*, as a function of *NDSI*. (a) Sites with small plots (H and M). (b) Sites with larger plots (S, N and J). Regression line: $SVF = 0{\cdot}23 + NDSI$. Correlation: $R^2 = 0{\cdot}92$

0·08). The only sparse-forest observation ($SVF = 0{\cdot}60$) had an *NDSI* between the dense stands and the open fields (0·30).

With *NDSI* for the snow-course sites with sufficiently large plots (S, J and N), *F* could be estimated from (Figure 9)

$$F = -0{\cdot}37NDSI + 0{\cdot}29, R^2 = 0{\cdot}85 \quad (15)$$

The agreement between S_F values measured and estimated from *NDSI* values was considerably less ($R^2 = 0{\cdot}54$; Figure 10) than for the ground-based estimates.

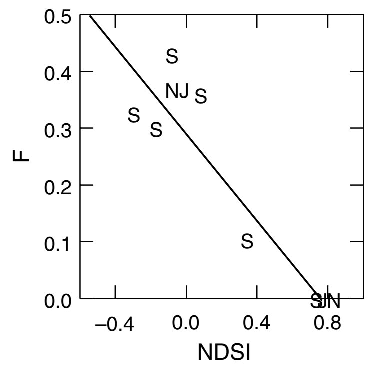

Figure 9. Relationship between F and $NDSI$ for snow-course plots larger than 1 ha (S, N and J). Regression line: $F = -0.37\ NDSI + 0.29$. Correlation: $R^2 = 0.85$

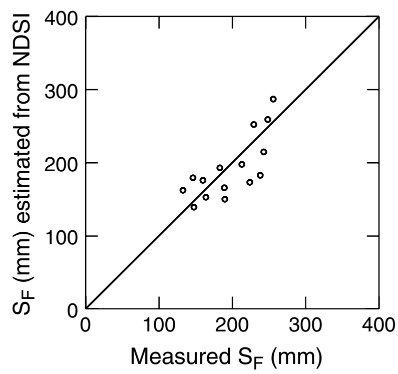

Figure 10. S_F calculated from $NDSI$ (Equation 11) versus measured for plots larger than 1 ha. $R^2 = 0.54$ with regression line forced through origin. The slope 1 : 1 is shown as comparison

DISCUSSION

Our analysis of Nakai's (1996) data showed that forest structure plays a more important role than forest species to predict snow accumulation and interception sublimation. The correlation between measured and estimated S_F was high irrespective of which measure was used for canopy structure (SVF with $R^2 = 0.86$ or LAI with $R^2 = 0.91$). Pomeroy *et al.* (2002) found lower correlation ($R^2 = 0.77$) when they estimated S_F based on Equation (7) with $C_1 = 0.43$. The larger correlation found here is probably caused by a larger S_O range (231 to 534 mm) compared with that of Pomeroy *et al.* (2002) (44 to 124 mm). The F value (0·43) for closed stands was higher than those of Kuzmin (1963) for both fir (0·37) and pine (0·22), but similar to the value reported by Pomeroy *et al.* (2002). The F values for open forests (SVF ≈0·60) were lower (≈0·05) than those of both Kuzmin (1963) and Pomeroy *et al.* (2002) (Figure 4). Two counteracting processes may lead to the non-detectable sublimation from stands with high SVF/low LAI: a small snow-interception sublimation from the deciduous forests in combination with a slightly enhanced snow accumulation in the small openings between trees, acting as miniature glades. Most of the forests in this study were forest plantations (*c.* 20 years old). Sublimation rates are positively correlated to the aerodynamic roughness of the canopy (e.g. Lundberg and Halldin, 2001). For the same SVF, a naturally regenerated forest canopy can be expected to have a higher roughness than a canopy with uniform tree height. This could indicate an even higher sublimation rate in natural forests than reported here.

An F value of 0·43 means that as much as 260 mm water equivalent per winter evaporates directly from the canopy branches in a closed forest such as M, where P_G is around 600 mm, and that 105 mm per winter evaporates at sites N, H and J, where P_G is around 240 mm. Our I values should be approximately 13 mm larger per winter than estimated because the difference between E_{SF} and E_{SO} was neglected. This means that the F values are between two to five percentage units too low ($13/250 = 5\%$; $13/600 = 2\%$; Equation 5). During a winter season (around 120 days at M and 100 days at H, J and N) the average interception–sublimation rate becomes $\approx$2·2 mm day^{-1} at M and $\approx$1·1 mm day^{-1} for the area around Sapporo, where the sites N, H, and J are located. No independent estimates for F or I are available from the high-precipitation site M to confirm these findings. A mid-winter eddy-correlation study near Sapporo indicates lower I rates (0·6 mm day^{-1}; Nakai *et al.* 1999a) than found in this study. Calder (1990) reported I equal to 0·5 mm h^{-1} and Harding and Pomeroy (1996) measured 4 mm sublimation for a 36-h period. An interception–sublimation rate of 2 mm day^{-1} for an area with very frequent snowfall therefore may be possible.

The NDSI technique used here to separate snow on forest floors from snow-free canopies is based on differences in reflectance between snow and vegetation. In the solar-radiation range, a typical reflectance (albedo) is 0·6–0·9 for new snow in open areas, whereas forest albedo is normally much lower (around 0·2). The literature is somewhat ambiguous about the influence of intercepted snow on forest winter albedo. Pomeroy and Dion (1996) showed it related to solar elevation and reported values between 0·1 and 0·3. Their analysis was restricted to small amounts of canopy storage (less than 1·5 mm per unit ground area) and they found a negligible influence from intercepted snow. Leonard and Eschner (1968), who measured albedo above a red-pine plantation after a 12-inch snow storm, report a maximum albedo below 0·2. Betts and Ball (1997) reported very low winter albedo values for a boreal forest, typically in the range 0·1–0·2 and rarely as high as 0·3. These findings are challenged by Kangas *et al.* (2001) who used broadband satellite reflectance to suggest that occurrence of snow on trees could explain locally high forests reflectance. Nakai *et al.* (2000) report that albedo above a Todo-fir stand increased with canopy snow up to a maximum value of 0·4. Albedo values as high 0·4 caused by canopy snow will make it difficult to use *NDSI* to estimate canopy forest density, but the small and moderate amounts of canopy snow in this study were assumed to have a small influence on canopy reflectance. Analysis using the NDSI requires information about canopy snow storage at the time of the scene, because large amounts of canopy snow may alter the canopy reflectance. This creates an obstacle in making this technique operational because information of canopy snow is not normally available. It should, however, be possible to identify periods with bare canopies from temperature and precipitation data.

The NDSI was used in this study to estimate *SVF*. Gomez *et al.* (2001) tested NDSI alone and in combination with NDVI (normalized difference vegetation index), the so-called SNOWMAP algorithm, to separate bare and snow-covered ground. They found the combination to be more accurate in forested areas. It is possible that use of SNOWMAP could have further improved our estimates of *SVF*.

The *NDSI* values reported by Klein *et al.* (1998) for jack pine and black spruce were around −0·4 for a totally closed stand and around +0·75 for a sparse stand, whereas the *NDSI* value for aspen was around 0 for a closed stand and around +0·75 for a sparse stand. These values are in accordance with the *NDSI* values observed here (smallest plots excluded) for closed forests (−0·4 to +0·1) but higher than our value (+0·3) for the sparse deciduous stand. The range of *NDSI* values for open snow fields reported here were 0·71–0·82 (smallest plots excluded), which are much smaller than the range (0·45–0·95) of Klein *et al.* (1998). They attribute their large range to differences in grain size. The small range found here indicates that a similar grain size was approximately the same for all observations. This is not surprising because observations were made at the same time of the year. The influence of differences in grain size decreases as canopy coverage increases according to Klein *et al.* (1998).

The *NDSI* values for forested and open plots at site M deviated from those found at other sites even where plot sizes were similar (sites N and J). No plausible explanation for this was found. Because the snow-course plots at site H (many observations in forests with varying *SVF*) were too small for the NDSI analysis, very few sparse forests observations were available. This makes the relationships between *NDSI* values and *SVF* or *F* somewhat hypothetical. There is therefore a need to acquire more data to approve the validity of these relationships.

CONCLUSIONS

Techniques are needed to incorporate realistic parameterizations of the forest-canopy snow-sublimation processes in land-surface models. A simple measure of forest properties is needed to relate sublimation fraction (F) of snowfall. This study shows that general measures of forest structure can be used for a variety of species. Our study indicates that *SVF*, C_C or *LAI* may be used interchangeably for this purpose and that they may be retrieved from satellite data using the NDSI algorithm.

The value of F was shown to decrease linearly with sky-view fraction (*SVF*) and increase linearly with LAI for five sites in southwest Hokkaido. The relationship was approximately the same, independent of snowfall, which ranged from 250 to 600 mm per winter. Snow accumulation (S_F) in forests can thus be estimated with a high correlation from snowfall on open fields (S_O) and *SVF*, C_C or *LAI* in the forest. The sky-view fraction was shown to be related to the normalized difference snow index (*NDSI*) retrieved from a Landsat-TM image. There is a need to test this relationship with more *SVF* observations for sparse forests.

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REFERENCES

Bernier PY, Swanson RH. 1993. The influence of opening size on snow evaporation in the forests of Alberta Foothills. *Canadian Journal of Forest Research* **23**: 239–244.

Betts AK, Ball JH. 1997. Albedo over the boreal forest. *Journal of Geophysical Research* **102**(D24): 28 901–28 909.

Betts AK, Ball JH, Beljaars ACM, Miller MJ, Viterbo PA. 1996. The land surface atmosphere interaction: A review based on observational and global modeling perspectives. *Journal of Geophysic Research* **101**: 7209–7225.

Bicheron P, Leroy M. 1999. A method of biophysical parameter retrieval at global scale by inversion of a vegetation reflectance model. *Remote Sensing of Environment* **67**(3): 251–266.

Bonan GB, Pollard D, Thompson SL. 1992. Effects of boreal forest vegetation on global climate. *Nature* **359**: 716–718.

Calder IR. 1990. *Evaporation in the Uplands*. Wiley: Chichester; 148 pp.

Chen JM, Rich PM, Gower ST, Norman JM, Plummer S. 1997. Leaf index of boreal forests: theory, techniques, and measurements. *Journal of Geophysical Research* **102**: 29 429–29 444.

Delfs J. 1967. Interception and stemflow in stands of Norway spruce and beech in West Germany. Proceedings from International Symposium held at University Park, Pennsylvania, Aug–Sept 1965. *Forest Hydrology*, Sopper WE, Lull HW (eds). Pergamon Press: Oxford; 179–185.

Doty RD, Johnson RS. 1969. Comparison of gravimetric measurements and mass transfer computations of snow evaporation beneath selected vegetation canopies. In *Proceedings of the 37th Western Snow Conference*, Salt Lake City, Utah; 57–62.

Eklund A, Gardelin M, Lindroth A. 2000. *Vinteravdunstning med HBV-modellen—Jämförelse med Mätdata*. Swedish Meterological and Hydrological Institute: Norrköping, Sweden; Nr. 83, 31 pp.

Eklundh L, Harrie L, Kuusk A. 2001. Investigating relationships between Landsat ETM + sensor data and leaf area index in a boreal conifer forest. *Remote Sensing of Environment* **78**(3): 239–251.

Essery R, Pomeroy J. 2001. Sublimation of snow intercepted by coniferous forest canopies in a climate model. In *Proceedings Symposium Soil-Vegetation-Atmosphere Schemes and Large Scale, Hydrological Models*, Sixth IAHS Assembly at Maastricht, The Netherlands, July 2002. IAHS-AISH Publications **270**: 343–348.

Gary HL. 1974. Snow accumulation and snowmelt as influenced by a small clearing in a Lodgepole pine forest. *Water Resources Research* **10**: 348–353.

Gomez-Landesa E, Rango A, Hall DK. 2001. Improved snow cover remote sensing for snowmelt runoff forecasting. *Proceedings of Remote Sensing and Hydrology Symposium, St Fee, New Mexico, USA, April 2000*. IAHS Publication 267, International Association of Hydrological Sciences: Wallingford; 61–65.

Gusev EM Nasonova ON. 2001. Parametrization of heat and moisture transfer processes in ecosystems of boreal forests. *Izvestiya. Atmospheric and Ocean Physics* **37**: 167–185.

Gustafsson D, Lewan E, van den Hurk BJ, Viterbo P, Grelle A, Lindroth A, Cienciala E, Mölder M, Halldin S, Lundin L-C. 2003. Boreal forest surface parametrization in the ECMWF model—1D test with NOPEX long term data. *Journal of Applied Meteorology* **42**(1): 95–112.

Hall DK, Riggs GA, Salomonson VV. 1995. Development of methods for mapping global snow cover using moderate resolution imaging spectroradiometer data. *Remote Sensing of Environment* **54**(2): 127–140.

Harding RJ, Pomeroy JW. 1996. The energy balance of a winter boreal landscape. *Journal of Climate* **9**: 2778–2787.

Harding R, Gryning S-E, Halldin S, Lloyd CR. 2001. Progress in understanding of land surface/atmosphere exchanges at high latitudes. In *Land-surface/Atmosphere Exchange in High-latitude Landscapes*, Grassl H, Halldin S, Gryning S-E, Lloyd CR (eds). *Theoretical and Applied Climatology* **70**(1–4): 5–18.

Hedstrom N, Pomeroy J. 1998. Accumulation of intercepted snow: measurements and modelling of snow interception in the boreal forest. *Hydrological Processes* **12**: 1611–1625.

Järvenoja S, Lönnberg P. 1995. Impact of a new physiographic data and parametrized albedo—recent change in the operational HIRLAM system at FMI. *HIRLAM Newsletter* **19**: 16–18.

Johnson R. 1990. The interception, throughfall and stemflow in a forest in highland Scotland and the comparison with other upland forests in the UK. *Journal of Hydrology* **118**: 281–287.

Kangas M, Heikinheimo M, Laine V. 2001. Accuracy of NOAA AVHRR-based surface reflectance over a winter-time boreal surface—comparison with aircraft measurements and land-cover information. In *Processes and Measurement Techniques. Land-surface/Atmosphere Exchange in High-latitudes Landscapes*, Grassl H, Halldin S, Gryning S-E, Lloyd CR (eds). *Theoretical and Applied Climatology* **70**(1–4): 231–244.

Klein AG, Hall D, Riggs GA. 1998. Improving snow cover mapping in forests through the use of a canopy reflectance model. *Hydrological Processes* **12**: 1723–1744.

Koivusalo H, Kokkonen T. 2002. Snow processes in a forest clearing and in a coniferous forest. *Journal of Hydrology* **262**: 145–164.

Kurvonen L, Pulliainen J, Hallikainen M. 1998. Monitoring of boreal forests with multitemporal special sensor microwave imager data. *Radio science* **33**(3): 731–744.

Kuzmin PP. 1963. *Formirovanie Snezhnogo Pokrova i Metody Opredeleniya Snegozapasov* (Snow cover and snow reserves). (English translation by Israel Program for Scientific Translation, Jerusalem). 139 pp.

Leonard RE, Eschner AR. 1968. Albedo of intercepted snow. *Water Resources Research* **4**(5): 931–935.

Lettenmaier DP, Bowling LC (eds). 2001. *Report of the ACSYS/GEWEX GLASS-PHILS2E stage 1 Arctic Hydrological Model Intercomparison Study Workshop*, Seattle, WA, March 2001. World Climate Programme (WCRP) Informal Report no. 15/2001.

Lundberg A, Halldin S. 2001. Snow interception evaporation—rates, processes and measurement techniques. In *Land-surface/Atmosphere Exchange in High-latitudes Landscapes*, Grassl H, Halldin S, Gryning SE, Lloyd CR (eds). *Theoretical and Applied Climatology* **70**(1–4): 117–133.

Lundberg A, Koivusalu H. 2002. Estimating winter evaporation in boreal forests with operational snow course data. *Hydrological Processes* **17**: 1479–1493.

Nakai Y. 1996. *An observational study on evaporation from intercepted snow on forest canopies*. Doctoral thesis, Department of Agriculture, Kyoto University, Kyoto, Japan, 107 pp. (In Japanese, Tables and Figures in English).

Nakai Y, Sakamoto T, Terajima T, Kitamura K, Shirai T. 1999a. Energy balance above a boreal coniferous forest: a difference in turbulent fluxes between snow-covered and snow-free canopies. *Hydrological Processes* **13**: 515–529.

Nakai Y, Sakamoto T, Terajima T, Kitamura K, Shirai T. 1999b. The effect of canopy-snow on the energy balance above a coniferous forest. *Hydrological Processes* **13**: 2371–2382.

Nakai Y, Kitamura K, Sakamoto T, Terajima T, Shirai T. 2000. The albedo above forests: seasonal variation, comparison of forest species, and the influence of thinning. *Biometeorology and Urban Climatology at the Turn of the Millennium: Selected Papers from the conference ICB-ICUC 99*, de Dear RJ, Kalma JD, Oke TR, Auliciems A (eds). WMO/TD No.1026, World Meteorological Organization; Geneva; 223–227.

Parviainen J, Pomeroy JW. 2000. Multiple-scale modelling of forest snow sublimation: initial findings. *Hydrological Processes* **14**(15): 2669–2681.

Pomeroy JW, Dion K. 1996. Winter radiation extinction and reflection in a boreal pine canopy: measurement and modelling. *Hydrological Processes* **10**: 1591–1608.

Pomeroy JW, Goodison BE. 1997. Winter and snow. In *The Surface Climates of Canada*, Bailey WG, Oke TR, Rouse WE (eds). McGill-Queen University Press: Montreal; 68–100.

Pomeroy JW, Parviainen J, Hedstrom N, Gray DM. 1998a. Coupled modelling of forest snow interception and sublimation. *Selected Papers from the 55th Eastern Snow Conference held in Jackson, New Hampshire, USA*, Taylor S, Hardy JP (eds). *Hydrological Processes* **12**: 2317–2337.

Pomeroy JW, Gray DM, Shook KR, Toth B, Essery RLH, Pietroniro A, Hedstrom N. 1998b. An evaluation of snow accumulation and ablation processes for land surface modelling. *Selected Papers from the 55th Eastern Snow Conference held in Jackson, New Hampshire, USA*, Taylor S, Hardy JP (eds). *Hydrological Processes* **12**: 2339–2367.

Pomeroy JW, Gray DM, Hedstrom NR, Janowicz JR. 2002. Prediction of seasonal snow accumulation in cold climate forests. *Hydrological Processes* **16**: 3543–3558.

Price JC, Bausch WC. 1995. Leaf area index estimation from visible and near-infrared reflectance data. *Remote Sensing of Environment* **52**(1): 55–65.

Thomas G, Rowntree P. 1992. The boreal forest and climate. *Quarterly Journal of the Royal Meteorological Society* **118**: 469–497.

White JD, Running SW, Nemani R, Keane RE, Ryan KC. 1997. Measurement and remote sensing of LAI in rocky mountain montane ecosystems. *Canadian Journal of Forest Research* **27**(11): 1714–1727.

Yang Z-L, Dickinson RE, Schuttleworth WJ, Shaikh M. 1998. Treatment of soil, vegetation and snow in land surface models: a test of the biosphere–atmosphere transfer scheme with the HAPEX-MOBILHY, ABRACOS and Russian data. *Journal of Hydrology* **212–213**: 109–127.

Zhang Y, Tian Y, Knyanzikhin Y, Martonchik, JV, Leroy M, Myneni RB. 2000. Prototyping of MISR LAI and FPAR algorithm with POLDER data over Africa. *Remote Sensing of Environment* **38**(5): 2402–2418.

