Modeling Forest Cover Influences on Snow Accumulation, Sublimation, and Melt

A. N. GELFAN

Water Problems Institute, Russian Academy of Sciences, Moscow, Russia

J. W. POMEROY

Department of Geography, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

L. S. KUCHMENT

Water Problems Institute, Russian Academy of Sciences, Moscow, Russia

(Manuscript received 1 December 2003, in final form 23 March 2004)

ABSTRACT

A comprehensive, physically based model of snow accumulation, redistribution, sublimation, and melt for open and forested catchments was assembled, based on algorithms derived from hydrological process research in Russia and Canada. The model was used to evaluate the long-term snow dynamics of a forested and an agricultural catchment in northwestern Russia without calibration from snow observations. The model was run with standard meteorological variables for the two catchments, and its results were tested against regular surface observations of snow accumulation throughout the winter and spring period for 17 seasons. The results showed mean errors in comparison to observations of less than 3% in estimating snow water equivalent during the winter and melt seasons. Snow surface evaporation and blowing snow were found to be small components of the mass balance, but intercepted snow sublimation removed notable amounts of snow over the winter from the forested catchment. Average snow accumulation was 15% higher in the open catchment, largely due to a lack of intercepted snow sublimation. Melt rates were 23% higher in the open than in the forest, but the effect on melt duration was suppressed by the smaller premelt accumulation in the forest. Only a moderate sensitivity of snow accumulation to forest leaf area was found, while a substantial variation was observed from season to season with changing weather patterns. This suggests that the ensemble of snow processes is more sensitive to variations in atmospheric processes than in vegetation cover. The success in using algorithms from both Canada and Russia in modeling snow dynamics suggests that there may be a potential for large-scale transferability of the modeling techniques.

1. Introduction

Differences between snow characteristics in open and forested areas have been observed for many years in both North America and the former USSR (Kuz'min 1954, 1960, 1961; Komarov 1959; Popov 1963; Miller 1964; Subbotin 1966; Apollov et al. 1974; Adams and Findlay 1966; Golding and Swanson 1986; Jeffrey 1970; Meiman 1970; Barry 1991; Pomeroy and Goodison 1997). These differences depend both on factors such as relief, type of forest, season, and climate and on the size of areas under consideration. For example, according to Russian experiences (e.g., Kuz'min 1960) mean maximum snow water equivalent (SWE) in small forest meadows or in clearings is close to that in deciduous forests, but is 5%–35% larger than SWE in pine forests and 10%–60% larger than in spruce forests. At the same time, mean maximum SWE in the large fields and in the open catchments, as a rule, is 15%–20% smaller than in neighboring forested catchments. Mean spring snowmelt rate in dense coniferous forest may be 3–5 times smaller and for deciduous forest 1.5–2 times smaller than the rate in open areas.

In Canada, many similar observations have been recorded. For example, Swanson (1988) found that accumulation in forest clearings peaked in clearings 2–3 times the tree height in diameter and then declined to values smaller than adjacent forests for clearings larger than 10 times the tree height in diameter. Pomeroy et al. (2002) found snow accumulation in forests varied with winter effective leaf area index (LAI), and for low LAI stands, such as deciduous stands, they found snow accumulation to be similar to that in open areas. Pomeroy and Granger (1997) and Faria et al. (2000) found snowmelt rates in forest clearings to be roughly 3 times larger than that in adjacent forests. Melt rates decreased

Corresponding author address: Alexander Gelfan, Water Problems Institute, Russian Academy of Sciences, 117735, Gubkin 3, Moscow, Russia.
E-mail: hydrowpi@aqua.laser.ru

with increasing leaf area index. Faria et al. (2000) discussed the effect of substand-scale variability in melt energy, SWE, and covariability between melt energy and snow accumulation in accelerating the melt rates at forest-stand scales; the greatest acceleration of melt due to covariance was in mixed-wood stands where at microscales, locations with relatively low SWE had relatively high melt energy.

Kuz'min (1961) suggested that the maximum SWE in ungauged forested catchments could be estimated by multiplying measured field SWE by a ''coefficient of snow accumulation,'' which is a constant for the specific regions (the value of this constant is obtained from measurements in the regions with similar physiographic and canopy conditions). However, the coefficient of snow accumulation varies over a wide range for such regions from year to year. According to data presented by Komarov (1959), Kuz'min (1960), Subbotin (1966), Vershinina (1972), and Shutov (1994) for the forest zone of European Russia, this coefficient changes from 0.61 to 1.40 (for the coniferous forests) and from 1.04 to 1.23 (for the deciduous forests). Furthermore, in some regions the coefficient of snow accumulation can change by 200%–300% during a winter (Vershinina 1972). During the last decade, there has been a dramatic reduction in the number of snow course observations of accumulation in both Canada and Russia. For confident estimates of SWE and snowmelt, especially for ungauged forested catchments, it is important to have physically based models of snow accumulation and melt processes that describe the influence of forest and can directly predict differences in snow characteristics for various meteorological and canopy conditions.

Hedstrom and Pomeroy (1998), Lundberg et al. (1998), Pomeroy et al. (1998), Nakai et al. (1999), Ohta et al. (1999), Suzuki et al. (1999), Parviainen and Pomeroy (2000), Storck et al. (2002), and Essery et al. (2003) described how snow interception and sublimation processes influence mass and energy exchange in the canopy and reduce net precipitation under forest canopies. Snow redistribution by wind in open areas has been described by Pomeroy et al. (1993), Pomeroy and Gray (1995), Liston and Sturm (1998), Essery et al. (1999), and Pomeroy and Li (2000). Energy balance calculations that take into account the influence of the forest on snowmelt, neglecting interception effects, have been developed by Price and Dunne (1976), Hardy et al. (1997, 1998), Davis et al. (1997), Link and Marks (1999), and Woo and Giesbrecht (2000) and show the importance of the forest canopy in reducing net radiation and turbulent transfer. Pomeroy and Granger (1997) investigated the combined influence of the canopy on interception, sublimation, and melt in a mixed Canadian boreal forest. Koivusalo and Kokkonen (2002) constructed a model of forest snow accumulation and melt that described the differences between snow processes in a coniferous forest and in an adjacent clear-cut area on the basis of four seasons of observations in southern Finland. This model helped show that midwinter melting in open areas and canopy interception in forested areas reduced snow accumulation such that maximum snow accumulation could be similar between forested and open sites because of these competing processes.

The objective of this paper is to develop a physically based, uncalibrated model that comprehensively describes the processes of snow accumulation and melting for open and forested catchments, and to use the model to compare the effect of forest cover on snow accumulation and melt. For the confident evaluation of such a comprehensive snow model for a range of meteorological conditions, it is important to have a long series of observations that include a range of both warm and cold and wet and dry winters. Given this aim, we use long-term data series available from the Valdai water balance station in northwestern Russia.

2. Site description and data

The Valdai water balance station (57°58′N, 33°14′E) is situated in the northwestern part of European Russia in the southern taiga forest zone (Fig. 1a). The climate of the region is relatively temperate with wet cold winters. Mean annual air temperature is 3.1°C (mean temperature for January is −9.7°C). Mean annual precipitation is 700 mm, of which 30% is snowfall. Stable snow cover typically forms during the third week in November and persists until the end of April.

Analyses of snow processes and physically based model development were carried out for two small catchments. The main part (73%) of the catchment of Taezgny Creek (area = 0.45 km^2) is covered by a mature spruce forest aged 90–110 yr, with an average tree height of 27–29 m. The canopy coverage of these stands ranges from 0.6 to 0.7. The remaining part of the catchment is a marshy area covered by pine stands, aged 70–90 yr, with an average tree height of 18–20 m. The canopy coverage of these stands is 0.6. The main part (81%) of the catchment of Usadievsky Creek (area = 0.36 km^2) is used for agriculture (ploughed fields and grasslands); the remaining part is covered by marshes. Mean maximum SWE for the forested catchment is 126 mm; for the open catchment it is 142 mm. In average, snow cover persists until 30 April and 12 April, respectively.

At a meteorological station nearby both sites, 3-hourly meteorological data (air temperature, humidity, precipitation, wind speed, and cloudiness) were collected (Fig. 1b). Measurements of snow depth and density (six transects with total length of 3800 m for the open catchment and five transects with total length of 3100 m for the forested one), surface snow evaporation, and subcanopy precipitation were collected over 17 snow accumulation and melt seasons (1966–82) in each catchment. Subcanopy precipitation measurements at Taezgny Creek were in a stand of typical characteristics for the catchment.

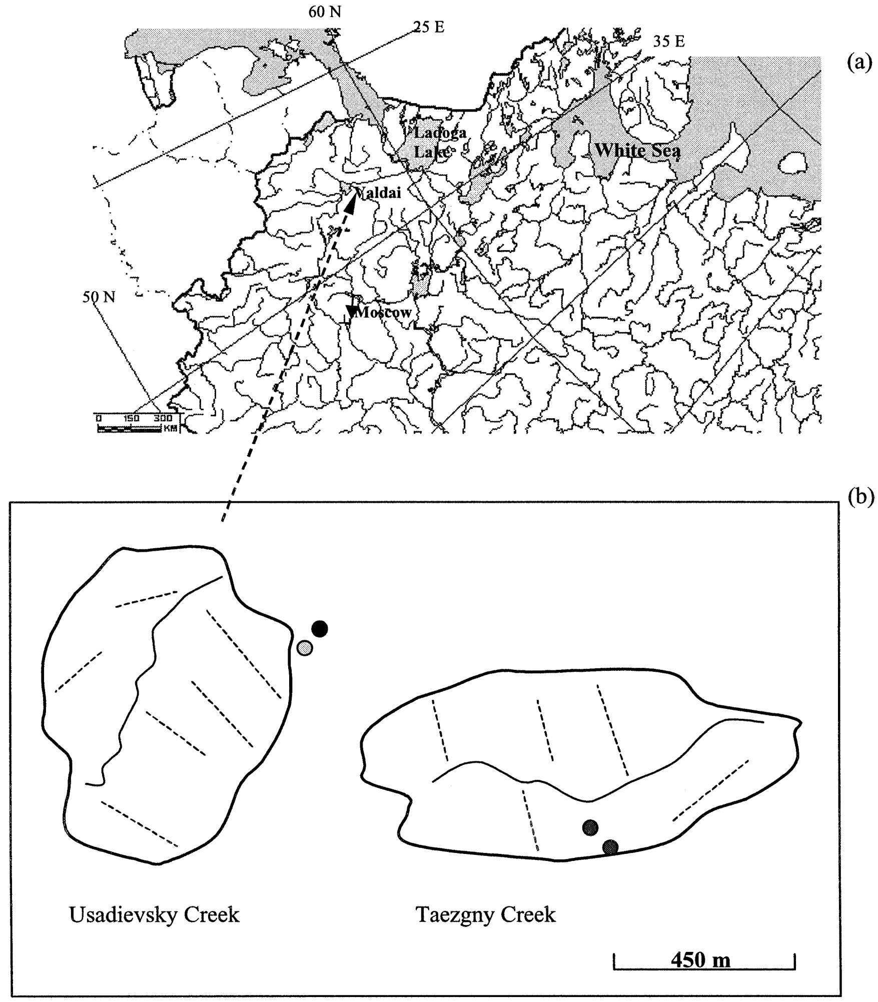

FIG. 1. (a) Location of the Valdai water balance station and (b) scheme of the used observational sites at Usadievsky and Taezgny watersheds: ● = meteorological station; dark-shaded area = subcanopy precipitation; light-shaded area = surface snow evaporation; and ——— = snow measurement transect.

3. Methods

a. Basic coupled energy and mass balance

To calculate the characteristics of snow cover, a system of vertically averaged equations of snow processes in a point has been applied (Kuchment and Gelfan 1996). The system includes a description of temporal change of the snow depth, content of ice and liquid water, snow density, snowmelt, sublimation, refreezing meltwater, and snow metamorphism and is written as follows:

$$\frac{dH}{dt} = \rho_w[X_s\rho_0^{-1} - (S + E_s)(\rho_i i)^{-1}] - V, \quad (1a)$$

$$\frac{d}{dt}(\rho_i iH) = \rho_w(X_s - S - E_s) + S_i, \quad (1b)$$

$$\frac{d}{dt}(\rho_w wH) = \rho_w(X_l + S - E_l - R) - S_i, \quad (1c)$$

where H is the snow depth; i and w are the volumetric content of ice and liquid water, respectively; and X_s and X_l are the snowfall rate and the rainfall rate, respectively. It is assumed that if the temperature of air $T_a \geq 0°C$, then only rainfall occurs, and if $T_a < 0°C$, then only snowfall occurs; S is the snowmelt rate; ρ_w, ρ_i, ρ_s, and ρ_0 are the density of water (1000 kg m^{-3}), ice (917 kg m^{-3}), snowpack, and fresh fallen snow (taken as 70 kg m^{-3}), respectively; the density of snowpack is calculated as $\rho_s = \rho_i i + \rho_w w$; E_l is the rate of liquid water evaporation from snow; E_s is the rate of snow sublimation; S_i is the rate of refreezing of meltwater in snow; R is the meltwater outflow from snowpack; and V is the snowpack compression rate.

Equations (1a)–(1c) were numerically solved by an explicit finite-difference scheme with a time step of 30 min.

The melting rate S is found from the energy balance equation as

$$S = (Q_{\mathrm{sw}} + Q_{\mathrm{lw}} - Q_{\mathrm{ls}} + Q_T + Q_E + Q_P + Q_G)(\rho_w \chi)^{-1}, \quad (2)$$

where Q_{sw} is the net shortwave radiation; Q_{lw} is the incoming longwave radiation; Q_{ls} is the outgoing longwave radiation; Q_T is the sensible heat exchange; Q_E is the latent heat exchange; Q_P is the heat content of liquid precipitation; Q_G is the heat exchange at ground surface; and χ is the latent heat of fusion (333.5 kJ kg^{-1}).

The use of Eq. (2) requires measurements of radiation and of heat exchange components, which are often not available. In the absence of direct meteorological model estimates or observations, energy budget components can be estimated from empirical relationships with commonly observed driving meteorological parameters such as air temperature and humidity, wind velocity, and cloudiness (e.g., U.S. Army, Corps of Engineers 1956; Kuz'min 1961). It is recognized that the accuracy of such relationships is affected significantly by many physiographic and climatic factors (Male and Gray 1981).

For estimation of the energy budget components, we applied the formulas obtained by Kuz'min (1961) and tested by many researchers using snow measurements in different regions of the former USSR including the Valdai region. These formulas are given in the following subsections (radiation fluxes are given in watts per meter squared).

1) Shortwave radiation

Here

$$Q_{\mathrm{sw}} = Q_0(1.00 - r)(1.00 - 0.20N - 0.47N_0), \quad (3)$$

where $Q_0 = 1000\beta$ (Kuz'min 1961) is the shortwave radiation flux under clear-sky conditions for the day and the hour in question; β is the angle of shortwave radiation above the horizontal in radians, calculated as a function of the local latitude (φ), the declination (δ), and the sun's hour angle (ω) by formulas

$$\sin\beta = \sin\varphi \sin\delta + \cos\varphi \cos\delta \cos\omega,$$

$$\delta = 1346 \sin\frac{2\pi}{\Pi}(t_d - 81), \quad \text{and}$$

$$\omega = \frac{\pi}{12}(t_h - 12). \quad (4)$$

Here Π is the number of days in 1 yr; t_d is the number of days from 1 January to the day under consideration; t_h is the local time (in hours from midnight to the hour under consideration); r is the snow albedo, calculated by the empirical relation $r = 1.03 - \rho_s\rho_w^{-1}$, suggested by (Kuchment et al. 1983); and N and N_0 are the total and the lower-level cloudiness (ratiometric), respectively.

2) Net longwave radiation

Here

$$Q_{\mathrm{lw}} = \sigma T_a^4(0.61 + 0.05e_a^{0.5}) \times (1.00 + 0.12N + 0.12N_0) - \varepsilon_s \sigma T_s^4, \quad (5)$$

where σ is the Stefan–Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}); e_a is the vapor pressure in millibars, ε_s is the effective emissivity of the snowpack (taken as 0.99); T_s is the temperature (K) of the snow surface; and T_a is the temperature of air in kelvins.

3) Turbulent exchanges

Using long-term snow evaporation measurements from the Valdai water balance station, Kuz'min (1961) obtained the following empirical formula for the rate of surface (not intercepted or blowing) snow sublimation (in millimeters per day):

$$E_s = (0.18 + 0.098u)(e_s - e_a), \quad (6)$$

where u is the wind velocity at 10-m height in meters per second, e_s is the saturated vapor pressure over the ice in millibars, and e_a is the air vapor pressure at 2-m height.

Following from Eq. (6), the latent heat flux, Q_E in watts per meter squared, can be calculated as

$$Q_E = -\chi_s\rho_w E_s = 32.8(e_a - e_s)(0.18 + 0.098u), \quad (7)$$

where χ_s is the latent heat of sublimation (2834 kJ kg^{-1}).

Coupling Eq. (7) with the Bowen ratio expressed as $Q_T/Q_E = 0.57(T_a - T_s)/(e_a - e_s)$, Kuz'min (1961) found a formula for sensible heat exchange:

$$Q_T = 18.7(T_a - T_s)(0.18 + 0.098u). \quad (8)$$

4) Heat content of liquid precipitation

Here

$$Q_P = \rho_w C_w T_a X_l, \quad (9)$$

where C_w is the specific heat capacity of water (4.18 kJ kg^{-1} °C^{-1}).

Heat exchange Q_G between the melting snow cover and the ground was assumed to be negligible. According to our estimates for the Valdai site, the mean value of Q_G during the cold period (from November to March) does not exceed 10% of the mean net radiation (4 and 48 W m^{-2}, respectively). Our estimations of the winter Q_G are close to the observational data reported in (Kuz'min 1961) for the northwestern part of Russia, where the Valdai station is located.

5) Wind redistribution of snow

In order to estimate the effect of wind redistribution of snow at the open site, we applied the Simplified Blowing Snow Model (SBSM) presented in Essery et al. (1999). The SBSM is a parametric model using measured wind speed, humidity, air temperature, and snowfall amounts to efficiently reproduce the results of the Prairie Blowing Snow Model as described by Pomeroy and Li (2000). This simulation calculates downwind blowing snow transport and in-transit sublimation losses over a fetch distance and then via continuity calculates surface snow erosion and accumulation. The model was run in single-column mode to estimate if snow was being blown from the open sites.

b. *Effect of forest canopy*

Meteorological conditions for snow accumulation and melt on forest floors differ from those in clearings because of the influence of the canopy. Precipitation, air temperature, humidity, wind speed, radiation fluxes, etc. are seldom measured under canopies, creating a considerable problem for estimating snowpack dynamics from meteorological inputs. Two approaches to estimating subcanopy meteorology are commonly used. The first approach is based on simulating the physical mechanisms by which canopies transform radiation, turbulent exchange, and mass fluxes. This approach was realized, for example, by Pomeroy and Dion (1996) for radiation fluxes, by Hedstrom and Pomeroy (1998) for precipitation, and by Nakai et al. (1999) for turbulent fluxes. The second approach involves using empirical relationships between meteorological variables obtained from simultaneous measurements in the open and in the forest. For example, Kuz'min (1961) suggested linear transformations of the values of the incoming shortwave radiation, the net longwave radiation, and the wind speed in the open site, in the respective subcanopy values. Similar approaches were used by Metcalfe and Buttle (1998).

A combination of physically based and empirical approaches is proposed to assign inputs to the model of snow accumulation and melt in the forest floor. The approach reflects the current lack of understanding of some of the processes and the need to minimize uncertainty and complexity in the model.

1) SNOW INTERCEPTION, SUBLIMATION, AND UNLOADING

Subcanopy precipitation, X_s^f, was found using a mass balance of intercepted snow, sublimation, melting, and unloading, where

$$X_s^f = X_c - I + U, \qquad (10)$$

and X_c is snowfall to the canopy (considered equal to the open-area snowfall), I is snow interception in the canopy, and U is unloading from the canopy. The snow interception formulation of Hedstrom and Pomeroy (1998) relates interception to leaf area index, tree species, canopy density, air temperature, wind speed, and snowfall. For a single snowfall event into a snow-free canopy, Hedstrom and Pomeroy's algorithm can be simplified to its primary factors, as

$$I = I^*[1 - e^{(-C_p X_c)/I^*}], \qquad (11)$$

where $I^* = S_p \mathrm{LAI}'(0.27 + 46\rho_0^{-1})$; I^* is interception capacity; S_p is the species snow-loading coefficient; LAI′ is effective winter leaf area index (total horizontal area of stems and needles per unit area of ground times the clumping factor); and C_p is the maximum plan area of the snow–leaf contact per unit area of ground.

Schmidt and Gluns (1991) present field measurements that suggest a value for S_p of 5.9 kg m^{-2} for spruce. The value of C_p can be approximated as 1 for conifer canopies (Hedstrom and Pomeroy 1998) where wind speeds exceed 1 m s^{-1}. According to Dzhogan and Lozinskaya (1993), the average value of winter LAI for spruce trees at the Taezgny catchment equals 9.0. However, this value does not include the clumping factor, Ω, which can cause a 62% reduction in LAI for spruce (Gower and Norman 1991) in calculating the effective LAI′ (LAI′ $= \Omega \times$ LAI), used in Hedstrom and Pomeroy's algorithm. This provides LAI′ $= (1 - 0.62) \times 9.0 = 3.4$ for Taezgny, similar to mature spruce stand measurements in Canada (Pomeroy et al. 2002). Thus, with a value for fresh snow density, $\rho_0 = 70$ kg m^{-3}, the snow interception capacity of the stand is $I^* = 0.019$ m. This value was adopted in our model.

Unloading was calculated for frozen conditions as

$$U = (1 - e^{-\alpha t})(I - E_t^i), \qquad (12)$$

where α is an unloading rate coefficient (per second) found by Hedstrom and Pomeroy (1998) to be 0.000 646 7 s^{-1} in cold conditions, t is elapsed time between snowfall events, and E_t^i is sublimation of intercepted snow over elapsed time t. Hedstrom and Pomeroy found wind speed to be unimportant for unloading in frozen conditions

Snow sublimation models (e.g., Pomeroy and Gray 1995) presume a thermodynamic equilibrium at the "ice bulb" temperature. This temperature is controlled by the water vapor deficit with respect to ice, air temperature, and ventilation rate (see, e.g., Iribarne and Godson 1981). For conditions where the wet-bulb/ice-bulb temperature exceeds 0°C and wind speed is greater than 0.5 m s^{-1} then intercepted snow in the canopy is considered to be sufficiently ventilated to be isothermal at 0°C and therefore subject to melt. Pomeroy and Gray (1995) discuss the mechanisms for unloading of intercepted snow, all of which increase dramatically for wet-snow conditions. For this reason, U is set equal to $I - E_t^i$ when the wet-bulb temperature exceeds 0°C for 3 h or more and wind speed is greater than 0.5 m s^{-1}. This is consistent with the unloading criterion of Storck et al. (2002), is physically meaningful, and is computationally simple.

Calculation of canopy snow sublimation with bulk transfer equations such as Eq. (6) is inappropriate for a vast number of reasons (Parviainen and Pomeroy 2000); however, for nonmelting conditions, the rate of sublimation of intercepted snow E^i can be found from the energy balance of intercepted snow:

$$E^i \rho_w \chi_s = -(R_{\mathrm{net}}^i + Q_T^i + Q_P), \qquad (13)$$

where R_{net}^i is the net radiation absorbed by the inter-

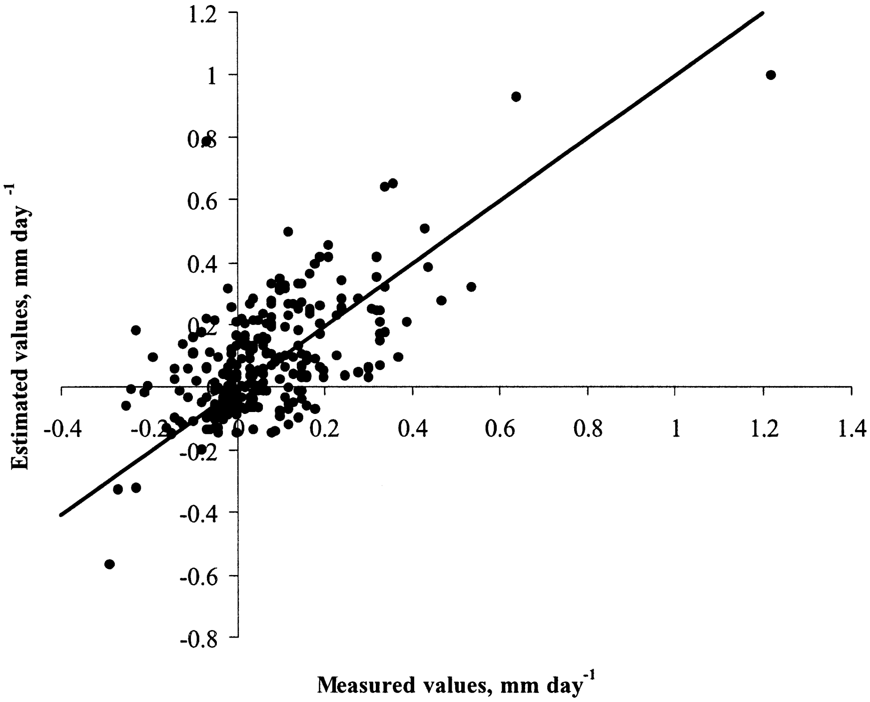

FIG. 2. Estimated vs measured surface snow evaporation in open sites (Valdai station; 1971–73).

cepted snow, and Q_T^i is the sensible heat exchange for intercepted snow.

2) SUBCANOPY RADIATION

The net radiation, $R_{\rm net}^i$, absorbed by the intercepted snow may be expressed as

$$R_{\rm net}^i = Q_{\rm sw}[1 - r_c - k_{\rm sw}(1 - r)] + Q_{\rm lw} + Q_{\rm ls} - 2Q_{\rm lc}, \quad (14)$$

where r_c is the canopy albedo [according to Pomeroy and Dion (1996) it is 0.12]; $k_{\rm sw}$ is the transmissivity through the canopy; $Q_{\rm ls}$ is the upward longwave radiation from snow on the forest floor, calculated as

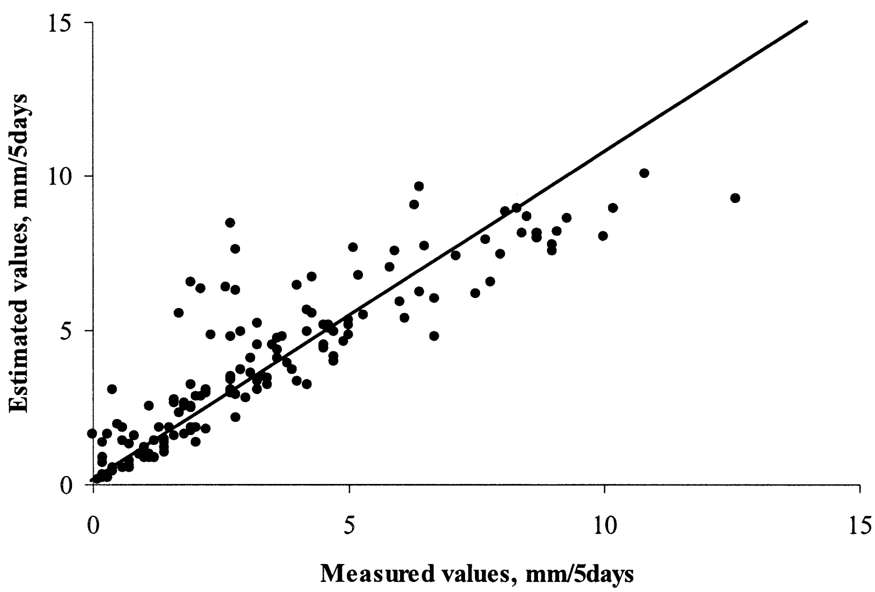

FIG. 3. Estimated vs measured water equivalent of intercepted snow (Taezgny catchment; 1970–79).

TABLE 1. List of the parameters adopted by the model of snow accumulation and melt.

Mathematical symbol	Physical meaning	Numerical value	Source
	Model parameters calculated by empirical formulas		
I^*	Snow interception capacity	0.019 m	Hedstrom and Pomeroy (1998)
k_{sw}	Transmittance	Depending on solar angle and LAI (0.02–0.06 in this study)	Pomeroy and Dion (1996)
r	Snow albedo	Depending on snow density (0.62–0.96 in this study)	Kuchment et al. (1983)
k_u	Coefficient of wind shield	Depending on canopy coverage (0.14 in this study)	Kuz'min (1954)
	Model parameters taken from observations		
C_c	Canopy coverage	0.65	Fedorov (1977)
LAI′	Effective leaf area index	3.4	Pomeroy et al. (2002)
C_p	Maximum plan area of the snow–leaf contact per unit area of ground	1.0	Hedstrom and Pomeroy (1998)
S_p	Species snow-loading coefficient	5.9 kg m^{-2}	Schmidt and Gluns (1991)
r_c	Canopy albedo	0.12	Pomeroy and Dion (1996)
ε_c	Emissivity of canopies	0.96	Price and Petzold (1984)
ρ_0	Density of fresh-fallen snow	70 kg m^{-3}	Pomeroy and Gray (1995)

$\varepsilon_s \sigma T_s^4$ [see Eq. (5)]; Q_{lc} is the longwave radiation emitted by the canopy (upward and downward), calculated as $\varepsilon_c \sigma T_c^4$, where T_c is the temperature of canopy (K) and is assumed equal to air temperature; and ε_c is the emissivity of the canopy and is taken as 0.96 (Price and Petzold 1984).

Net radiation flux, R_{net}^f, into the snowpack on the forest floor is calculated as (e.g., Koivusalo and Kokkonen 2002)

$$R_{net}^f = Q_{sw}(1 - r)(1 - C_c + k_{sw}C_c) + C_cQ_{lc} + (1 - C_c)Q_{lw} - Q_{ls}, \qquad (15)$$

where C_c is the canopy coverage. According to (Fedorov 1977), $C_c = 0.65$ for the Taezgny catchment.

The component Q_T^i was calculated by Eq. (8), where T_s was presumed equal to the ice-bulb temperature.

Solar radiation transmission was modeled using results of Pomeroy and Dion (1996) from a Canadian pine forest:

$$k_{sw} = \exp(-Q_{ext}\mathrm{LAI}' \sin^{-1}\beta), \qquad (16)$$

where Q_{ext} is the extinction efficiency estimated as $Q_{ext} = 1.08\ \beta \cos(\beta)$ (β is in radians); this formula gives an effective multiple reflection canopy transmissivity appropriate for planophile evergreen canopies. The value of k_{sw} calculated from Eq. (16) under the assigned value of LAI′ = 3.4 reaches 0.047 when the solar angle is 40°. Kuz'min (1961) reported a very close value of k_{sw} (0.042) measured in the Taezgny catchment under the same conditions.

3) SUBCANOPY TURBULENT FLUXES

Detailed studies (Kuz'min 1954, 1961; Fedorov 1977) were carried out in the Valdai station in order to compare meteorological variables measured in the same sites as we used in this paper, namely, Usadievsky and Taezgny Creek basins. The main objective was to modify Eqs. (5)–(8) to take into account the canopy effect. It was shown that the air temperature during November–April at the height of 2 m above a spruce forest floor differed only slightly from one in an open site 5 km away. The mean deviation was +0.2°C (forest temperature is higher), with 75% of deviations between ±0.5°C. The 2-m water vapor pressure [see Eq. (5)] in the forest was also similar to that in the open; mean deviation was +0.1 mb, with 93% of deviations within ±0.5 mb. Importantly, the differences ($T_a - T_s$) and ($e_a - e_s$) [see Eqs. (7) and (8)] calculated for the forest are quite similar to those in the open.

The wind speed near the forest floor, u^f, in the Taezgny catchment was less than that in the open site, u, by a factor varying from 3 to 12 and depending on canopy density and forest species. Kuz'min (1954) called this factor the "coefficient of wind shield," k_u. Based on a great body of data, Kuz'min found a relationship between k_u and the canopy density of the spruce forest, C_c. The relationship may be approximated by the formula

$$k_u = 0.56 \exp(-2.25C_c), \quad \text{for } 0.2 < C_c < 0.9. \qquad (17)$$

For the Taezgny catchment $C_c = 0.65$; consequently $u^f = k_uu = 0.14u$. The value u^f was substituted for u in Eqs. (7) and (8) when calculating latent and sensible heat fluxes for subcanopy snowpack.

4) SUBCANOPY SNOWMELT

The melting rate S^f on the forest floor can be found from the energy balance equation as Eq. (2), taking into account the changes in subcanopy air temperature, humidity, and wind speed and comparing with the changes measured in the open site. The subcanopy snow char-

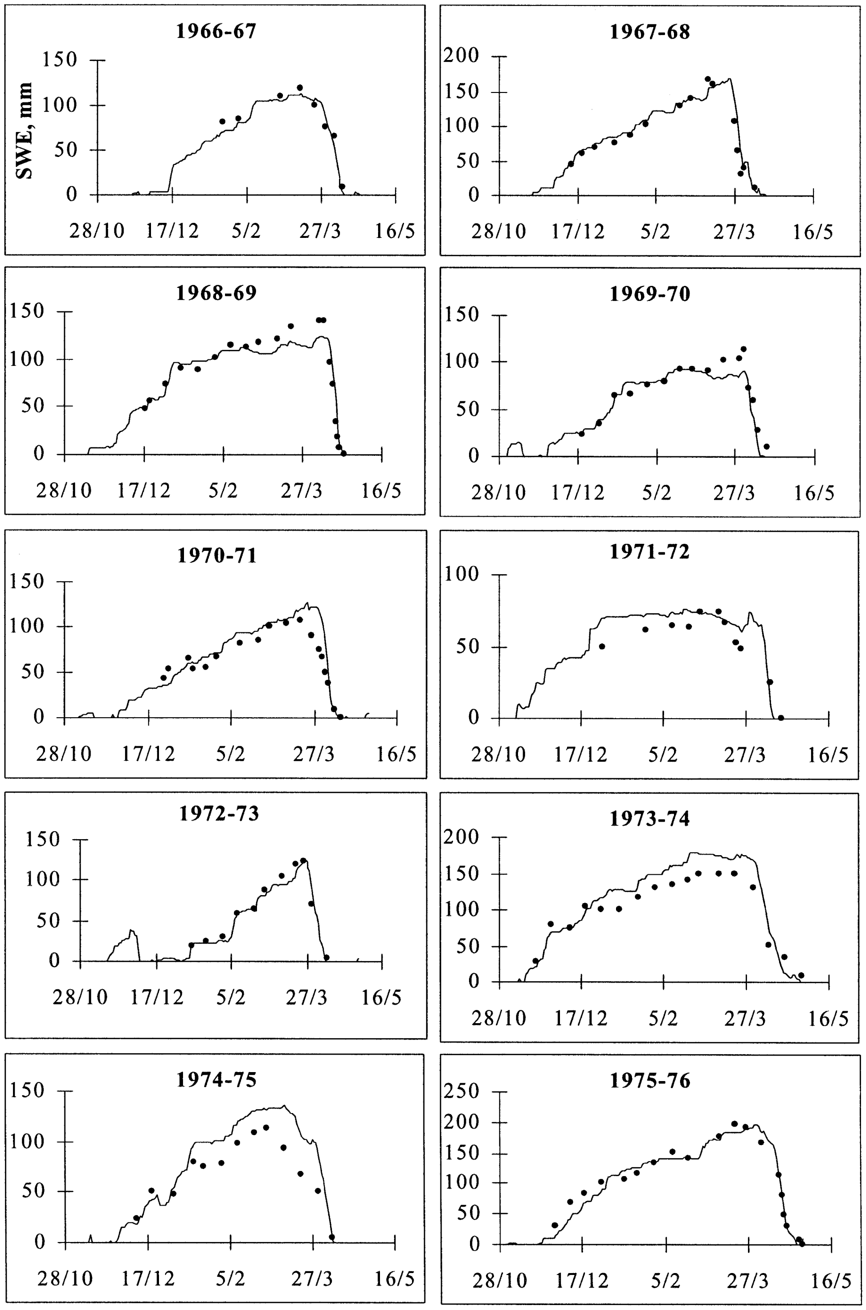

FIG. 4. Measured (points) and calculated (line) SWE in the Usadievsky catchment (open).

acteristics were calculated by Eqs. (1a)–(1c) with the following modifications of the input variables:

- subcanopy precipitation was calculated from a mass balance accounting for the interception, unloading, and intercepted snow sublimation processes [Eqs. (10)–(12)];
- subcanopy net radiation was calculated accounting for the effect of solar angle on extinction efficiency and hence solar radiation transmission and the longwave emission by the canopy [Eqs. (14)–(15)];
- subcanopy wind speed was calculated accounting for the wind shield factor [Eq. (17)];
- subcanopy values of the air temperature and humidity were presumed equal to those measured in the open.

4. Results

a. Snow surface sublimation

Equation (6) was evaluated using daily in situ snow-pan measurements of surface snow evaporation from

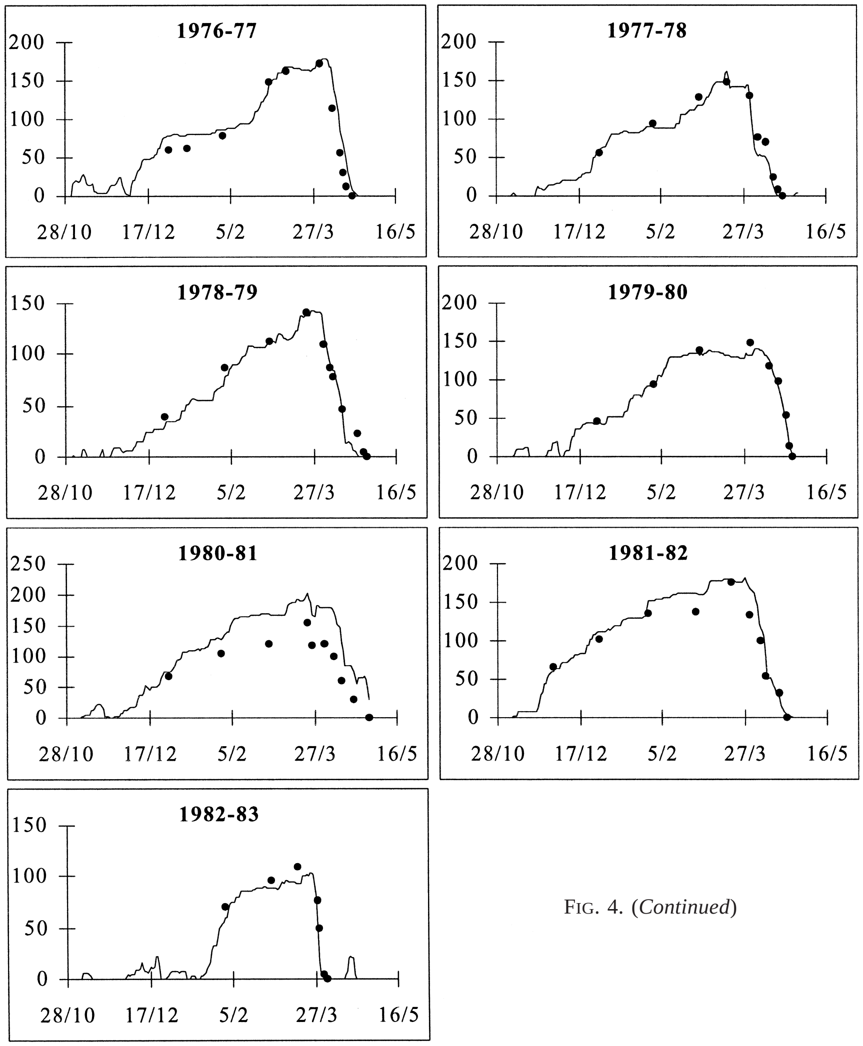

FIG. 4. (*Continued*)

open sites at the Valdai station from 1 February to 31 March for 1970–73. A comparison of estimated and measured values is presented in Fig. 2 and displays several features:

- measured average daily surface snow sublimation in late winter was small (0.17 mm day^{-1}), with 28% of days displaying condensation rather than sublimation;
- the mean error of estimate of surface sublimation was -0.02 mm over the period of evaluation.

Given the relatively small effect of surface sublimation on the mass balance and the small long-term error of evaluation, it is felt that Eq. (6) does not represent a significant error in snow accumulation and melt modeling for these catchments.

b. *Blowing snow*

The blowing snow simulation with fetch set to 1000 m over 17 seasons estimated a mean seasonal snow erosion and subsequent transport of 0.201 mm SWE with a maximum seasonal erosion for transport was 0.248 mm SWE. Such small values of snow transport resulted from the very low wind speeds at this site, which exceed 3 m s^{-1} only a few times per winter and hence did not often exceed the threshold condition specified by Li and Pomeroy (1997). Mean winter wind speed for the Valdai station is 2.6 m s^{-1}. It was therefore presumed that there was not significant wind redistribution of snow at the open site and that neither blowing snow transport nor sublimation would require further analysis in this simulation.

c. *Snow interception*

The calculated areal water equivalent of intercepted snow was compared with one derived from a mass balance of measurements of 5-day totals of subcanopy and above-canopy precipitation in the Taezgny catchment during the winters of 1970–79. The results of this comparison are shown in Fig. 3. The multiyear mean root-mean-square error between calculated and mea-

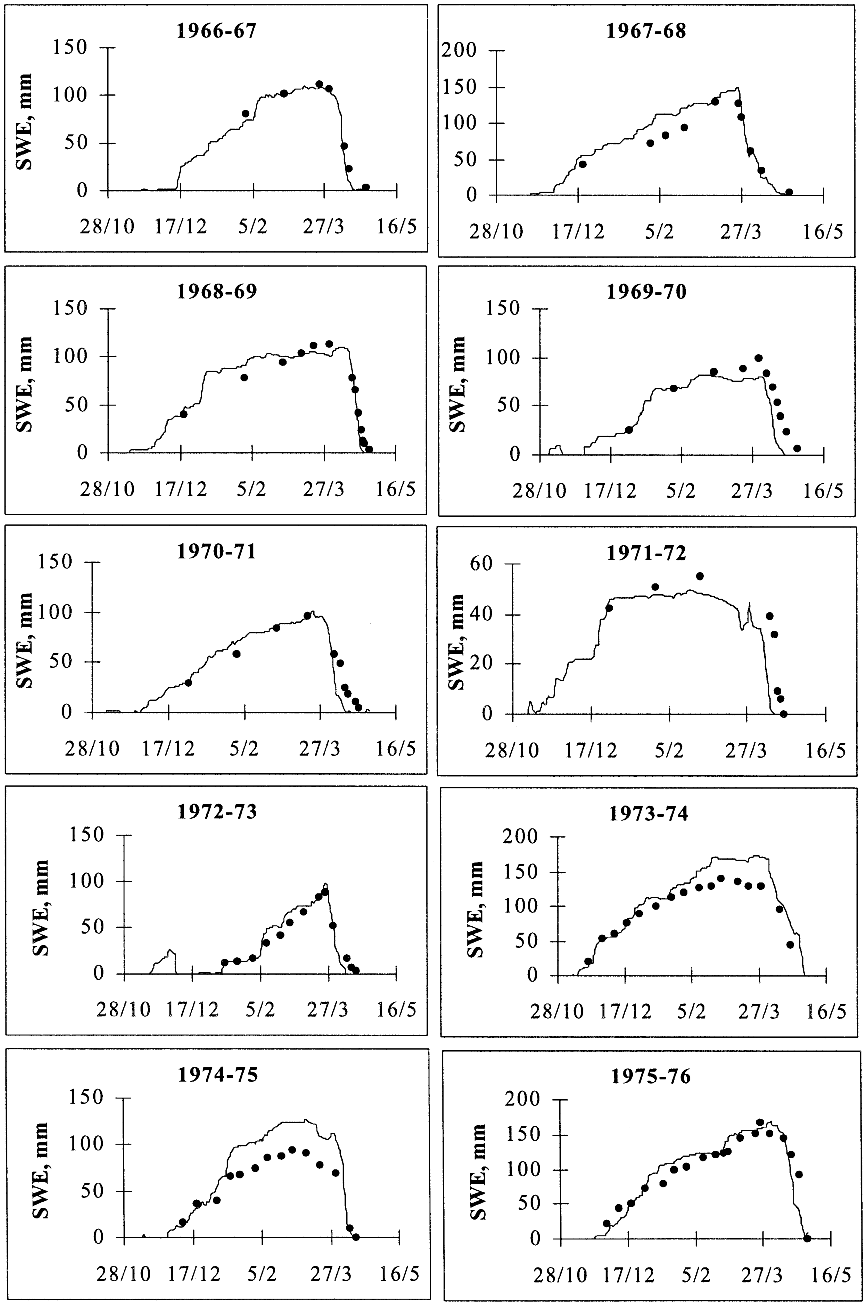

FIG. 5. Measured (points) and calculated (line) SWE in the Taezgny catchment (forested).

sured values was 1.27 mm; maximum values were for the winters of 1975 and 1976 (1.64 and 1.68 mm, respectively). Both measured and modeled maximum snow interception were an order of magnitude larger than that normally found for rainfall interception, emphasizing the need for a distinctive treatment of snow interception in hydrological models. This is a similarly good fit to measurements as for the original evaluation of Hedstrom and Pomeroy (1998), suggesting that the algorithm is robust and can be applied to the Valdai forested catchment with confidence.

d. Model parameterization, simulations, and performance

The parameters adopted for the model (except for well-known physical constants such as Stefan–Boltzmann constant, density of water, etc.) are shown in Table

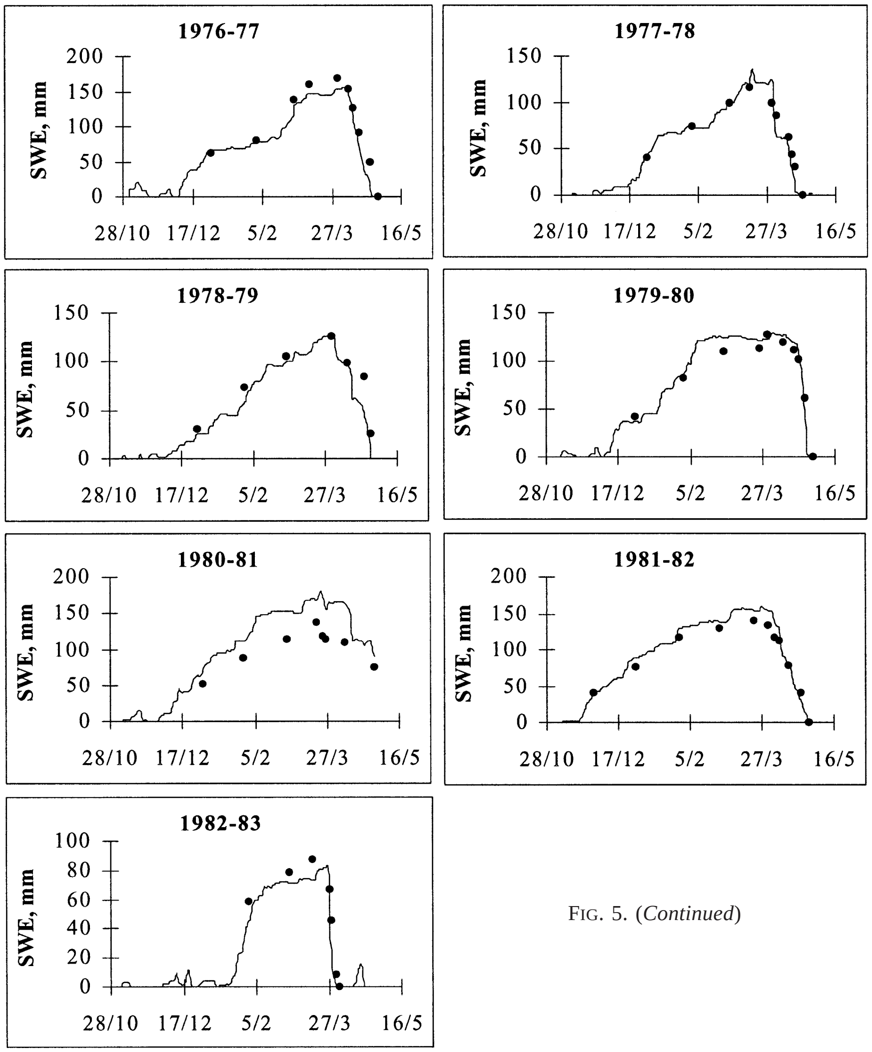

FIG. 5. (*Continued*)

1. The values of all parameters were either taken from observations in Russian and Canadian forests or calculated by empirical formulas base on these data. It should be stressed that the model does not contain any parameters that are derived by a model calibration process.

The complete model was run at a 30-min time step for 17 seasons (from 1 November to 30 April 1966–83) using 3-hourly meteorological data measured at the open field site. Model results were evaluated using measured SWE values averaged along 11 snow courses distributed through the Usadievsky and the Taezgny catchments. Seasonal dynamics of the measured SWE were compared with calculated values and are shown in Figs. 4 and 5.

The model results were assessed in terms of the mean error (ME, observed − modeled) and the root-mean-square error (rmse) of estimates of SWE (Table 2). Using

TABLE 2. Comparison of simulated and observed SWE (Valdai water balance station; 1966–82).

Catchment	No. of values	Observed mean (mm)	Simulated mean (mm)	Observed std dev (mm)	Simulated std dev (mm)	ME	Rmse
	All data						
Taezgny (forested)	81	73.4	77.8	43.0	53.1	−4.4	18.1
Usadievsky (open)	87	77.6	77.1	45.9	53.7	0.6	19.5
	Seasonal maximum SWE						
Taezgny	17	117.5	118.9	29.4	36.2	−1.4	15.5
Usadievsky	17	136.8	140.0	31.1	37.1	−3.2	18.7

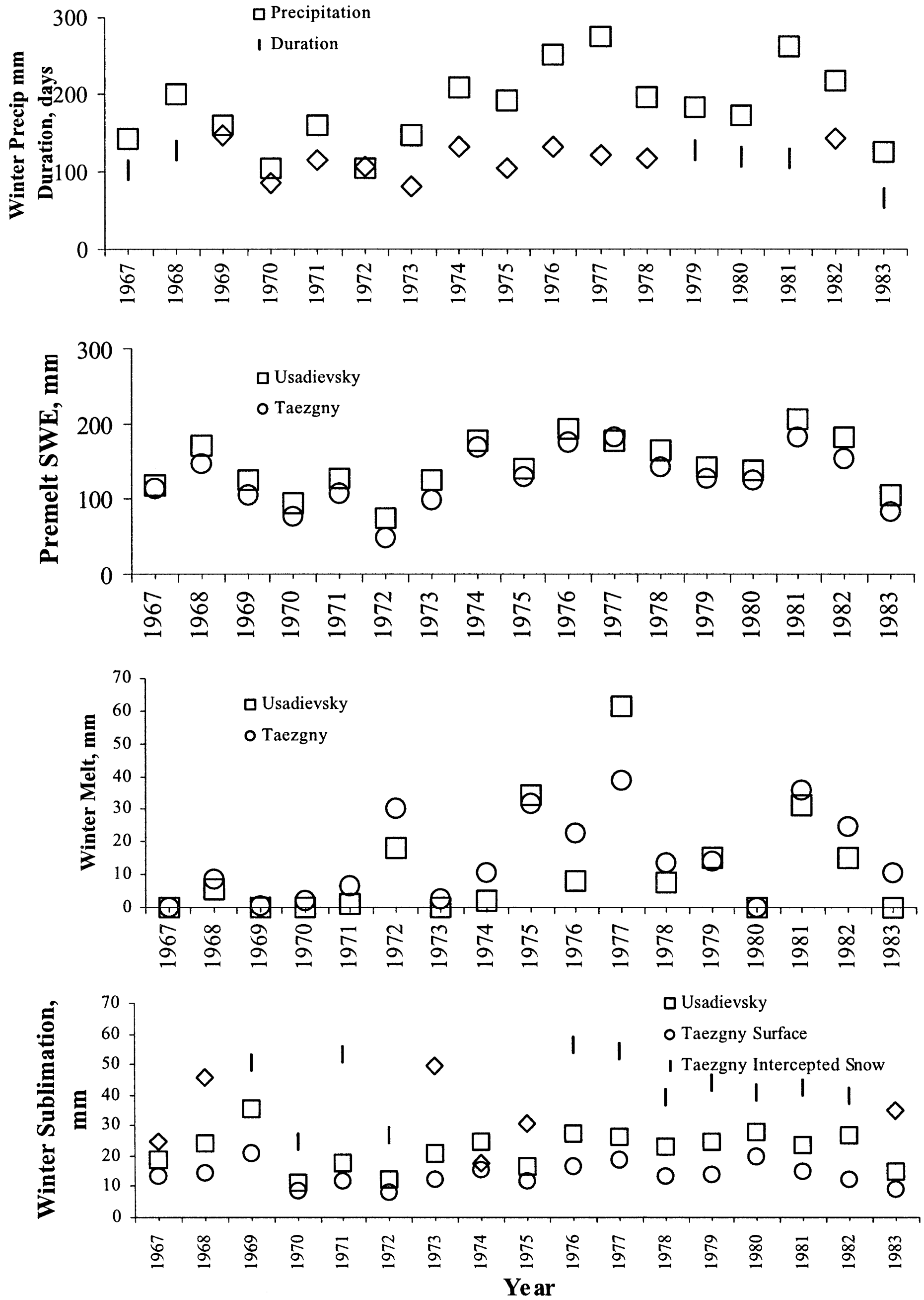

FIG. 6. Observed season precipitation, calculated duration of snow accumulation period, premelt SWE, overwinter melt, and overwinter sublimation from surface and intercepted snow for Usadievsky (open) and Taezgny (forested) catchments.

all seasonal SWE observations from 17 seasons (87 and 81 values for the open and the forested catchment, respectively), MEs are 0.57 and −4.35 mm for these catchments. The MEs of seasonal maximum SWE are −3.18 mm for the open and −1.33 mm for the forested catchment. Rmses of seasonal maximum SWE are 18.72 and 15.51 mm, respectively. The long-term average areal standard deviation of the measured maximum SWE is 40 mm for the open catchment and 27 mm for the forested one. These results suggest that the model satisfactorily simulates SWE both in the open and in the forested catchment.

5. Analysis and discussion

The results were divided into an accumulation and a melt period for analysis of the processes responsible for

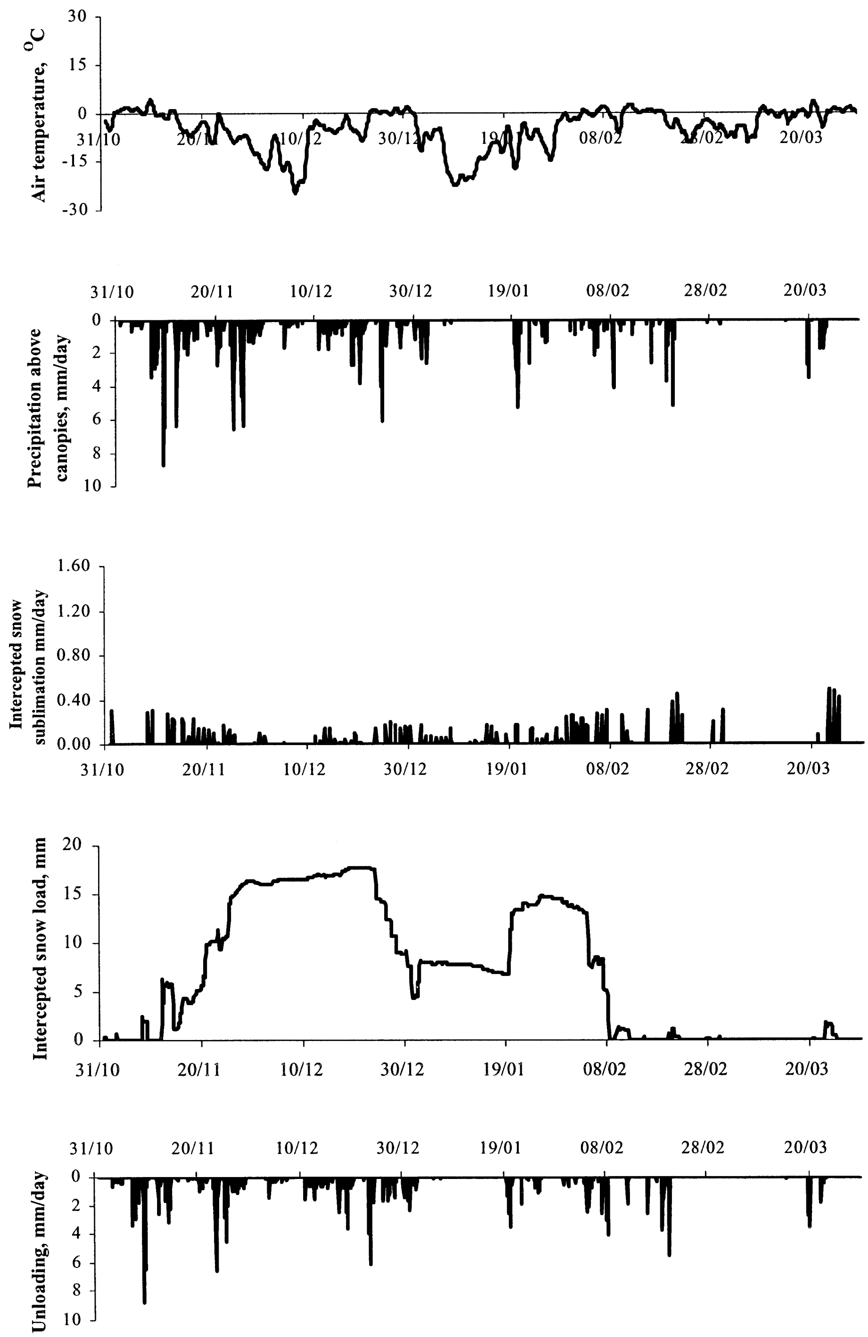

FIG. 7. Dynamics of measured air temperatures, precipitation, estimated interception, sublimation of intercepted snow, and unloading (Taezgny catchment; 1 Nov 1973–1 Apr 1974).

the differences in accumulation and melt in the two catchments. The accumulation period was defined as from the beginning of snow accumulation on the ground until the beginning of a reasonably monotonic decline in SWE from some peak, as evidence of the spring melt. The accumulation period therefore included midwinter melts, and the melt period included accumulation events.

a. Snow accumulation period

Figure 6 shows accumulation period precipitation (observed rain plus snow), modeled period duration, maximum premelt SWE, overwinter melt, and overwinter sublimation from surface and intercepted snow over the years of simulation. Winter precipitation and accumulation season duration in the catchments were

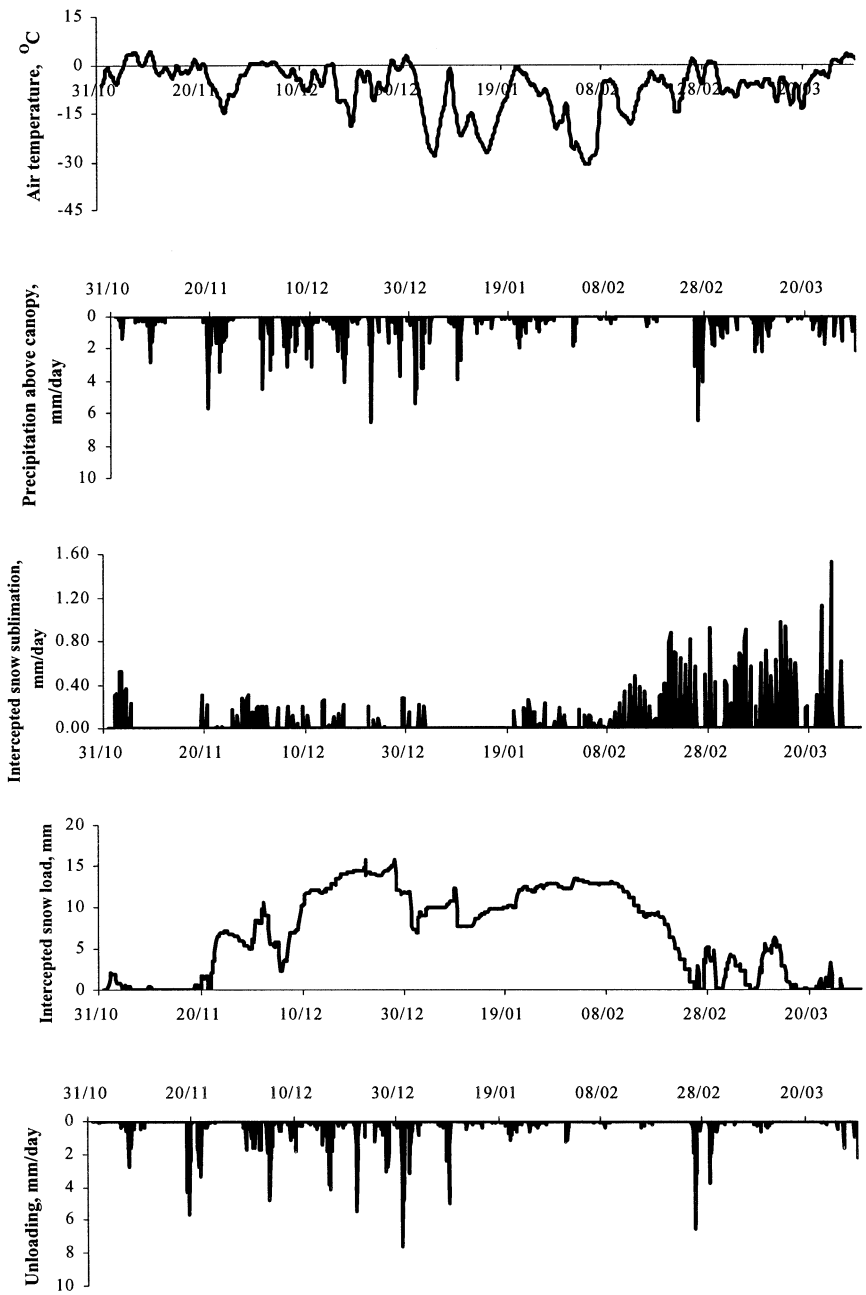

FIG. 8. Dynamics of measured air temperatures, precipitation, estimated interception, sublimation of intercepted snow, and unloading (Taezgny catchment; 1 Nov 1975–1 Apr 1976).

virtually identical and cannot be distinguished in Fig. 6. Accumulation season duration varied from about 2–5 months with a mean of 114 days for the open and 118 days for the forest, and duration length varied positively with winter precipitation. Maximum SWE was less than winter precipitation because of some rain events, melt, and sublimation. The maximum premelt SWE in the agricultural Usadievsky basin was on average 12% higher than that in the forested Taezgny basin. This difference changed widely from year to year; sometimes the SWE in the open catchment was 36% higher than that in the forested (1972), but in other years it was less than 10% (1967, 1974, 1975, and 1980). In 1977 the SWE in the forested catchment was 2% higher than in the open one. The differences in maximum SWE were due to differences in overwinter sublimation and melt.

As evident in Fig. 6 there were no large differences (about 6 mm on average) between winter ablation (melt + surface sublimation) of surface snow during midwinter periods for the forested and open catchment. The

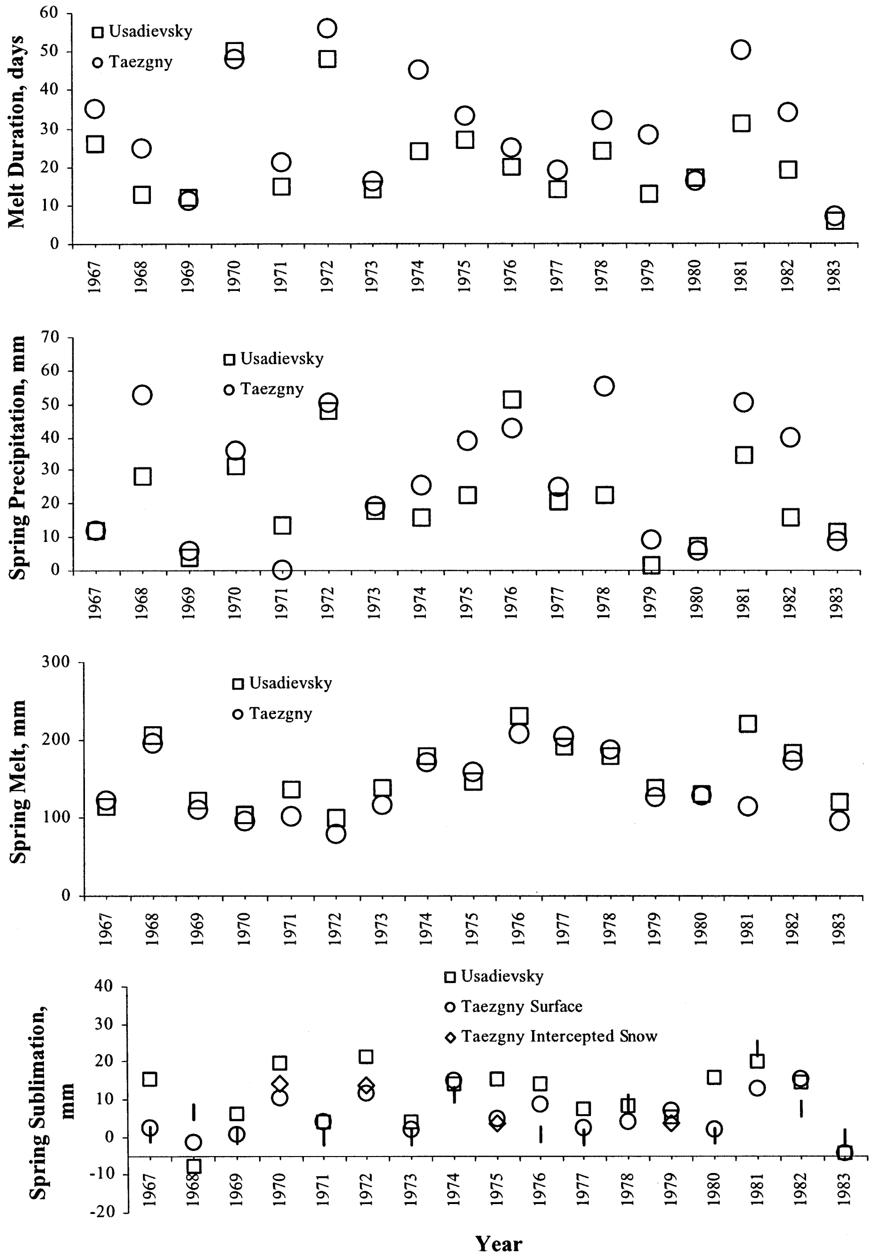

FIG. 9. Melt season calculated duration, observed precipitation, calculated sums of melt, and sublimation from surface and intercepted snow for Usadievsky (open) and Taezgny (forested) catchments.

multiyear average winter ablation rate was 0.29 mm day^{-1} for the open site and 0.25 mm day^{-1} for the forested one (the standard deviations were 0.14 and 0.10 mm day^{-1}, respectively). The average overwinter melt on the forest floor was close to that on the open surface; sublimation from the forest floor was 8 mm smaller. The small difference in estimates of midwinter ablation between the forest and the open sites are due to compensating processes. Longwave and turbulent fluxes are the primary contributors to midwinter ablation of surface snow, as shortwave fluxes are small at this time of year in high latitudes. Less intensive turbulent exchange in the forest, because of the wind speed reduction, is largely compensated by the additional source of incoming longwave radiation emitted by the canopy when compared to the open site.

The main reasons for differences in maximum SWE between the open and forested catchments were the interception and subsequent sublimation of intercepted snow. This is consistent with observations in the Canadian boreal forest (Pomeroy et al. 1998). Cumulative losses due to sublimation of intercepted snow in Taez-

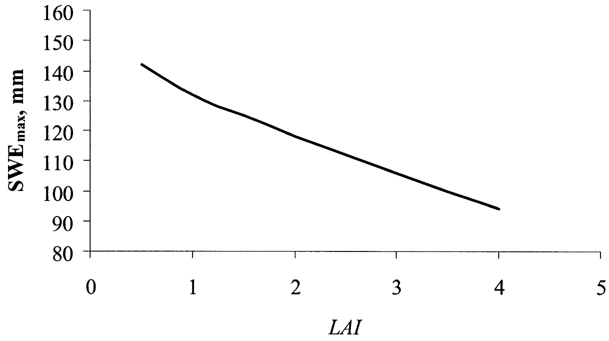

FIG. 10. Sensitivity of the calculated mean maximum SWE to changes of LAI.

gny catchment were 39 mm (21% of precipitation) on average. This seasonal sublimation loss is similar to Pomeroy et al.'s estimates in Canada and close to experimental estimates by Fedorov (1977). On the basis of 19 yr of field observations, he estimated overwinter interception losses in the Taezgny catchment as 18% of precipitation. There is a weak positive correlation between cumulative precipitation above the canopy and intercepted snow sublimation loss, with a correlation coefficient of 0.52. Other factors however have a strong influence on sublimation loss. According to our results, interception losses varied substantially from year to year; the minimum value was 20 mm for 1973–74, and the maximum value was 57 mm for 1975/76. Figures 7 and 8 show the overwinter dynamics of air temperature, precipitation, intercepted snow load, unloading, and sublimation during the periods of snow accumulation (from 1 November to 1 April) for 1973/74 and 1975/76, respectively. Above-canopy precipitation for these periods was 22 mm higher in 1975/76 (263 mm) than in 1973/74 (241 mm). The winter of 1975/76 (Fig. 8) was almost 2.4°C colder and had fewer melt events. This resulted in differing snow interception regimes for the two seasons. Figure 7 shows that in 1973/74 intercepted snow load dropped nearly to zero in early February as a result of a midwinter thaw. However, in 1976 (Fig. 8) snow remained intercepted until mid-March. There were no severe thaws in winter and early spring in 1976; therefore, the reduction in intercepted snow load was caused primarily by sublimation. These differences demonstrate the value of explicitly calculating snow interception, unloading, and sublimation dynamics as opposed to adopting less physically based methods.

b. Snowmelt period

Figure 9 shows the interseasonal variation of spring melt duration, precipitation, melt, and sublimation over the years of simulation for the two catchments. Spring melt and sublimation were slightly complementary rather than positively associated. In the snowmelt period, interception sublimation losses played a less important role than in midwinter because there were relatively few snowfalls in spring, and warm temperatures caused relatively rapid unloading. Cumulative sublimation of surface snow was also small. In most years open environment surface sublimation exceeded that of forest surface or intercepted sublimation. In only three springs did intercepted snow sublimation exceed surface sublimation. In two melt seasons, 1968 and 1983, net condensation from vapor to surface snow occurred in the forest catchment. Average sublimation rates in spring were different from those in midwinter. For the open site, sublimation rates increased by more than twofold from 0.19 mm day^{-1} in winter to 0.40 mm day^{-1} in spring. In the forest, surface snow sublimation did not change significantly, from 0.12 mm day^{-1} in winter to 0.14 mm day^{-1} in spring.

The duration of melt was positively associated with spring precipitation, melt, and sublimation, as one would expect. Average calculated ablation (melt + sublimation) rate during the spring period in the open site was 3.2 mm day^{-1} larger than in the forested one, due largely to the melt rate being 9.1 mm day^{-1} in the open and 6.1 mm day^{-1} in the forest. The ratio of ablation rates is close to that of the degree-day factors recommended for open and forested landscapes in this region (e.g., Apollov et al. 1974). Average calculated snowmelt durations in the open and the forested catchment were 22 and 30 days, respectively. This difference in duration is within 2 days of that observed (average 10 days) and is similar to that reported by Pomeroy and Granger (1997) for the western Canadian boreal forest despite the more continental climate in Saskatchewan, Canada. The differences in melt duration at Valdai are suppressed somewhat because of a positive association between snow accumulation and melt rate; lower melt rates in the forest were partly compensated for by lower snow accumulation. If premelt SWE had been identical at both sites then melt would have been prolonged by almost 2 weeks in the forest.

c. Sensitivity of premelt SWE to forest cover

Subcanopy snow-cover sensitivity analyses were carried out to investigate the effect of variation in LAI′ on the snow dynamics. The value of LAI′ was varied from 0.5 to 4.0, as were the parameters that are dependent on LAI′, such as canopy coverage, C_c (Pomeroy et al. 2002); transmittance, k_{sw} [Eq. (16)]; and coefficient of windshield, k_u [Eq. (17)]. Figure 10 presents the sensitivity of the modeled 17-yr-averaged seasonal maximum SWE to LAI′, varying from that typical of an open pine canopy to a dense spruce canopy. The analysis indicates that modeled peak premelt SWE is moderately sensitive to large changes in the derived canopy parameters, declining linearly by approximately 30% for an eightfold increase in leaf area. This sensitivity of snow accumulation to canopy is less than that found in western Canadian boreal forests, where midwinter sublimation may play a relatively larger role because of the drier climate and higher winter insolation levels. The average maximum accumulation for the open catchment was 146 mm, suggesting that in this climate (where blowing snow is unimportant), low LAI′ forests will have similar maximum snow accumulation to open sites. This is consistent with the observations of forest effects on maximum snow accumulation found in a literature survey of North American measurements by Pomeroy and Gray (1995) and predictive equations by Kuz'min (1960) and Pomeroy et al. (2002).

6. Conclusions

A comprehensive model of snow accumulation and ablation processes was shown to provide satisfactory simulations of snow water equivalent during winter accumulation and spring melt periods for both an agricultural and a forested catchment at Valdai in northwestern Russia. Mean errors over 17 seasons in estimating maximum snow accumulation were less than 5 mm (out of typical 150 mm), and for estimating SWE over the accumulation and ablation seasons, less than 4 mm. The model synthesized Russian and Canadian research and simulated the physical processes of snow redistribution, interception, sublimation, and the energy balance for melt without calibration from snow observations.

From the model results several conclusions can be made about the snow accumulation season in this region of Russia. Surface snow evaporation was found to be small and even in late winter averaged 0.17 mm day^{-1} with 28% of days undergoing condensation rather than sublimation. Model estimates were sufficiently accurate when compared to open-area observations. Blowing snow transport was negligible in both catchments because of low wind speeds. Interception of snowfall ranged up to 13 mm per 5-day period in the forested catchment and was well predicted by the model compared to interception loss estimated from a mass balance of snowfall and snow accumulation measurements on the ground. Intercepted snow loads in the canopy in excess of 15 mm persisted over much of the winter periods. Sublimation of intercepted snow was the major process contributing to differences in snow accumulation and melt between open and forested catchments, resulting in a loss of 39 mm, equivalent to 21% of seasonal precipitation. This contributed to the agricultural Usadievsky catchment having 13% higher premelt SWE than the forested Taezgny catchment, despite there being on average 5 mm more midwinter ground snow ablation in the agricultural catchment.

During the melt season total surface snow sublimation losses were smaller than in winter. However, average sublimation rates in spring were higher than those in midwinter: for the open site, sublimation rates increased by more than two times; in the forest, surface snow sublimation did not increase significantly (about 17%).

Average melt rates were 3 mm day^{-1} (33%) higher in the open than in the forest catchment. Higher melt rates in the open catchment were only partially compensated by higher accumulation there, leading to an 8-day difference in snowmelt duration between sites.

A sensitivity analysis of the effect of forest leaf area on premelt SWE showed only a moderate sensitivity, with a 30% decrease in SWE for an eightfold increase in leaf area. The results here show a smaller difference between open and forested environments in modeled snow accumulation and melt rates than has been observed in continental boreal forests in western Canada. It is suggested that differences in climate will have a more important role than differences in forest canopy in accounting for these differences. That these differences have been adequately modeled over a long time series of data with algorithms from both Canadian and Russian experience suggests that there may be potential for large-scale transferability of the simulation techniques to cold regions' forest in general.

Acknowledgments. The authors wish to acknowledge the support of the NATO Collaborative Linkage Grants scheme and a Natural Environment Research Council (UK) standard grant. This paper is a contribution to PUB, the Decade for Prediction of Ungauged Basins, International Association of Hydrological Sciences. The perseverance and fortitude of the many field personnel making winter measurements at the Valdai research basins have provided an important basis for the research of later generations and should be remembered.

APPENDIX

Notation

The following symbols and units are used in this paper:

C_c = Canopy coverage
C_p = Maximum plan area of the snow–leaf contact per unit area of ground
C_w = Specific heat capacity of water (J kg^{-1} °C^{-1})
e_a = Air vapor pressure (mb)
E_l = Rate of liquid water evaporation from snowpack (m s^{-1})
E_s = Rate of snow sublimation (m s^{-1})
E^i = Rate of sublimation of intercepted snow (m s^{-1})
E^i_t = Sublimation of intercepted snow over elapsed time t (m)
e_s = Saturated vapor pressure over the ice (mb)
H = Snow depth (m)
I = Snow interception in the canopy (m)
I^* = Interception capacity (m)
i = Volumetric ice content of snow
k_{sw} = Transmissivity of radiation through the canopy
k_u = Coefficient of wind shield

LAI	=	Leaf area index
LAI′	=	Effective winter leaf area index
ME	=	Mean error
N	=	Total cloudiness (fraction of unit)
N_0	=	Lower level cloudiness (fraction of unit)
Q_E	=	Latent heat exchange (W m^{-2})
Q_{ext}	=	Extinction efficiency
Q_G	=	Heat exchange at ground surface (W m^{-2})
Q_{lc}	=	Longwave radiation emitted by the canopy (W m^{-2})
Q_{ls}	=	Outgoing longwave radiation (W m^{-2})
Q_{lw}	=	Incoming longwave radiation (W m^{-2})
Q_P	=	Heat content of liquid precipitation (W m^{-2})
Q_{sw}	=	Net shortwave radiation (W m^{-2})
Q_T	=	Sensible heat exchange (W m^{-2})
Q_T^i	=	Sensible heat exchange for intercepted snow (W m^{-2})
Q_0	=	Shortwave radiation flux under clear-sky conditions (W m^{-2})
R	=	Meltwater outflow from snowpack (m s^{-1})
r	=	Snow albedo
r_c	=	Canopy albedo
Rmse	=	Root mean square error
R_{net}^f	=	Net radiation flux into the snowpack on the forest floor (W m^{-2})
R_{net}^i	=	Net radiation absorbed by the intercepted snow (W m^{-2})
S	=	Snowmelt rate (m s^{-1})
S^f	=	Snowmelt rate on the forest floor (m s^{-1})
S_i	=	Rate of refreezing of meltwater in snowpack (kg m^{-2} s^{-1})
S_p	=	Snow loading coefficient (kg m^{-2})
SWE	=	Snow water equivalent (m)
t	=	Time (s)
T_a	=	Air temperature (°C)
T_c	=	Temperature of canopy (°C)
t_d	=	Number of days from 1 January to the day under consideration
t_h	=	Local time (h) from midnight to the hour under consideration
T_s	=	Temperature of the snow surface (°C)
U	=	Unloading from the canopy (m)
u	=	Wind speed (m s^{-1})
u^f	=	Wind speed near the forest floor (m s^{-1})
V	=	Snowpack compression rate (m s^{-1})
w	=	Volumetric liquid water content of snow
X_l	=	Rainfall rate (m s^{-1})
X_s	=	Snowfall rate (m s^{-1})
X_s^f	=	Subcanopy precipitation (m)
α	=	Unloading rate coefficient (s^{-1})
β	=	Angle of shortwave radiation above the horizontal (radians)
δ	=	Declination (radians)
ε_c	=	Emissivity of the canopy
ε_s	=	Effective emissivity of the snowpack
Π	=	Number of days in 1 yr
ρ_i	=	Density of ice (kg m^{-3})
ρ_s	=	Density of snowpack (kg m^{-3})
ρ_w	=	Density of water (kg m^{-3})
ρ_0	=	Density of fresh-fallen snow (kg m^{-3})
σ	=	Stefan–Boltzmann constant (W m^{-2} K^{-4})
φ	=	Local latitude (radians)
χ	=	Latent heat of fusion (J kg^{-1})
χ_s	=	Latent heat of sublimation (J kg^{-1})
ω	=	Sun's hour angle (radians)
Ω	=	Clumping factor

REFERENCES

Adams, W. P., and B. F. Findlay, 1966: Snow measurements in the vicinity of Knob Lake, central Labrador-Ungava. *Proc. 25th Eastern Snow Conf.,* 110–139.

Apollov, B. A., G. P. Kalinin, and V. D. Komarov, 1974: *Hydrological Forecasting* (in Russian). Hydrometioizdat, 419 pp.

Barry, P. J., 1991: Discussion paper on the report by H. G. Jones. *Seasonal Snowpacks: Processes of Compositional Change,* T. D. Davies, M. Tranter, and H. G. Jones, Eds., NATO ASI Series, Vol. G(28), Springer-Verlag, 229–235.

Davis, R. E., J. P. Hardy, W. Ni, C. E. Woodcock, R. Jordan, J. C. McKenzie, and X. Li, 1997: Variation of snow cover ablation in the boreal forest—A sensitivity study on the effects of conifer canopy. *J. Geophys. Res.,* **102,** 29 389–29 395.

Dzhogan, L. Ya., and E. A. Lozinskaya, 1993: A method for averaging leaf relative area in mesoscale assessment of heat and moisture exchange between the underlying surface and the atmosphere (in Russian). *Water Resour.,* **20,** 693–700.

Essery, R., L. Li, and J. W. Pomeroy, 1999: Blowing snow fluxes over complex terrain. *Hydrol. Processes,* **13,** 2423–2438.

——, J. W. Pomeroy, J. Parviainen, and P. Storck, 2003: Sublimation of snow from coniferous forests in a climate model. *J. Climate,* **16,** 1855–1864.

Faria, D. A., J. W. Pomeroy, and R. L. H. Essery, 2000: Effect of covariance between ablation and snow water equivalent on depletion of snow-covered area in a forest. *Hydrol. Processes,* **14,** 2683–2695.

Fedorov, S. F., 1977: *Study of the Water Balance Components in the Forest Zone of the European Part of the USSR* (in Russian). Hydrometeoizdat, 264 pp.

Golding, D. L., and R. H. Swanson, 1986: Snow distribution patterns in clearings and adjacent forest. *Water Resour. Res.,* **22,** 1931–1940.

Gower, S. T., and J. M. Norman, 1991: Rapid estimation of leaf area index in conifer and broad-leaf plantations. *Ecology,* **72,** 1896–1900.

Hardy, J. P., R. E. Davis, R. Jordan, X. Li, C. E. Woodcock, W. Ni, and J. C. McKenzie, 1997: Snow ablation modeling at the stand scale in a boreal jack pine forest. *J. Geophys. Res.,* **102** (D24), 29 397–29 405.

——, ——, ——, ——, and ——, 1998: Snow ablation modelling in a mature aspen stand of the boreal forest. *Hydrol. Processes,* **12,** 1763–1778.

Hedstrom, N. R., and J. W. Pomeroy, 1998: Measurements and modelling of snow interception in the boreal forest. *Hydrol. Processes,* **12,** 1611–1625.

Iribarne, J. V., and W. L. Godson, 1981: *Atmospheric Thermodynamics.* 3d ed. D. Reidel, 259 pp.

Jeffrey, W. W., 1970: Snow hydrology in the forest environment. *Snow Hydrology: Proc. of a Workshop Seminar,* Ottawa, Ontario, Canada, Canadian National Committee for the International Hydrological Decade, 1–19.

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

Komarov, V. D., 1959: *Snowmelt River Runoff in the European Part of the USSR* (in Russian). Hydrometeoizdat, 294 pp.

Kuchment, L. S., and A. N. Gelfan, 1996: The determination of the snowmelt rate and the meltwater outflow from a snowpack for modeling river runoff generation. *J. Hydrol.,* **179,** 23–36.

——, V. N. Demidov, and Yu. G. Motovilov, 1983: *River Runoff Formation (Physically-Based Models)* (in Russian). Nauka, 216 pp.

Kuz'min, P. P., 1954: Influence of the forest on snow melt (in Russian). *Trans. State Hydrol. Inst.,* **42** (96), 3–68.

——, 1960: *Snow Accumulation and Methods of Estimating Snow Water Equivalents* (in Russian). Hydrometeoizdat, 171 pp.

——, 1961: *Process of Snow Melting* (in Russian). Hydrometeoizdat, 345 pp.

Li, L., and J. W. Pomeroy, 1997: Estimates of threshold wind speeds for snow transport using meteorological data. *J. Appl. Meteor.,* **36,** 205–213.

Link, T. E., and D. Marks, 1999: Point simulation of seasonal snow cover dynamics beneath boreal forest canopies. *J. Geophys. Res.,* **104** (D22), 27 841–27 857.

Liston, G., and M. Sturm, 1998: A snow-transport model for complex terrain. *J. Glaciol.,* **44,** 498–516.

Lundberg, A., I. Calder, and R. Harding, 1998: Evaporation of intercepted snow: Measurements and modelling. *J. Hydrol.,* **206,** 151–163.

Male, D. H., and D. M. Gray, 1981: Snowcover ablation and runoff. *Handbook of Snow, Principles, Processes, Management and Use,* D. M. Gray and D. H. Male, Eds., Pergamon, 360–436.

Meiman, J. R., 1970: Snow accumulation related to elevation, aspect and forest canopy. *Snow Hydrology: Proc. of a Workshop Seminar,* Ottawa, Ontario, Canada, Canadian National Committee for the International Hydrological Decade, 35–47.

Metcalfe, R. A., and J. M. Buttle, 1998: A statistical model of spatially distributed snowmelt rates in a boreal forest basin. *Hydrol. Processes,* **12,** 1701–1722.

Miller, D. A., 1964: Interception processes during snowstorms. Dept. of Agriculture, Forest Service Res. Paper PSW-18, 24 pp.

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

Ohta, T., K. Suzuki, Y. Kodama, J. Kubota, Y. Kominami, and Y. Nakai, 1999: Characteristics of the heat balance above the canopies of evergreen and deciduous forests during the snowy season. *Hydrol. Processes,* **13,** 2382–2394.

Parviainen, J. W., and J. W. Pomeroy, 2000: Multiple-scale modelling of forest snow sublimation: Initial findings. *Hydrol. Processes,* **14,** 2669–2681.

Pomeroy, J. W., and D. M. Gray, 1995: Snow accumulation, relocation and management. National Hydrology Research Institute, Environment Canada, Science Rep. 7, 144 pp.

——, and K. Dion, 1996: Winter radiation extinction and reflection in a boreal pine canopy—measurements and modelling. *Hydrol. Processes,* **10,** 1591–1608.

——, and B. E. Goodison, 1997: Winter and snow. *The Surface Climates of Canada,* W. G. Bailey, T. R. Oke, and W. R. Rouse, Eds., McGill-Queen's University, Press, 68–100.

——, and R. J. Granger, 1997: Sustainability of the western Canadian boreal forest under changing hydrological conditions. I. Snow accumulation and ablation. *Sustainability of Water Resources under Increasing Uncertainty,* D. Rosjberg et al., Eds., IAHS Publication 240, IAHS Press, 237–242.

——, and L. Li, 2000: Prairie and Arctic areal snow cover mass balance using a blowing snow model. *J. Geophys. Res.,* **105,** 26 619–26 634.

——, D. M. Gray, and P. G. Landine, 1993: The Prairie Blowing Snow Model: Characteristics, validation, operation. *J. Hydrol.,* **144,** 165–192.

——, J. Parviainen, N. R. Hedstrom, and D. M. Gray, 1998: Coupled modelling of forest snow interception and sublimation. *Hydrol. Processes,* **12,** 1592–2337.

——, D. M. Gray, N. R. Hedstrom, and J. R. Janowicz, 2002: Prediction of seasonal snow accumulation in cold climate forests. *Hydrol. Processes,* **16,** 3543–3558.

Popov, E. G., 1963: *Problems of the Theory and the Practice of Hydrological Forecasting* (in Russian). Hydrometeoizdat, 395 pp.

Price, A. G., and T. Dunne, 1976: Energy balance computations of snowmelt in a subarctic area. *Water Resour. Res.,* **12,** 686–694.

——, and D. E. Petzold, 1984: Surface emissivities in a boreal forest during snowmelt. *Arct. Alp. Res.,* **16,** 45–51.

Schmidt, R. A., and D. R. Gluns, 1991: Snowfall interception on branches of three conifer species. *Can. J. For. Res.,* **21,** 1262–1269.

Shutov, V. A., 1994: Spatial distribution of snow water equivalent for watersheds located in the forest zone (in Russian). *Meteor. Hydrol.,* **9,** 85–92.

Storck, P., D. P. Lettenmaier, and S. M. Bolton, 2002: Measurement of snow interception and canopy effects on snow accumulation and melt in a mountainous maritime climate, Oregon, United States. *Water Resour. Res.,* **38,** 1223, doi:10.1029/2002WR001281.

Subbotin, A. I., 1966: *Snowmelt and Rainfall Runoff* (in Russian). Hydrometeoizdat, 378 pp.

Suzuki, K., T. Ohta, H. Miya, and S. Yokota, 1999: Seasonal variations of heat balance components over a Japanese red pine forest in snowy northern Japan. *Hydrol. Processes,* **13,** 2409–2422.

Swanson, R. H., 1988: The effect of in situ evaporation on perceived snow distribution in partially clearcut forests. *Proc. 56th Western Snow Conf.,* Kalispell, MT, Western Snow Conference, 87–92.

U.S. Army, Corps of Engineers, 1956: Snow hydrology. Summary Rep. of the Snow Investigations, North Pacific Division, Portland, OR, 437 pp.

Vershinina, L. K., 1972: Snow accumulation in the open and forested river basins (in Russian). *Trans. State Hydrol. Inst.,* **194,** 93–119.

Woo, M., and M. A. Giesbrecht, 2000: Simulation of snowmelt in a sub-Arctic spruce woodland: 1. Tree model. *Water Resour. Res.,* **36,** 2275–2285.