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Modeling the snow cover in climate studies 1. Long-term integrations under different climatic conditions using a multilayered snow-cover model

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Abstract. A physically based model of a snow cover, which has been designed for climate studies, was tested under different conditions (Arctic, midlatitudes, alpine regions). This multi layered model can be used for stand-alone experiments as well as for implemention into sophisticated coupled models. One version of the model, which considers a partial snow coverage of the grid cells and an albedo masking effect by vegetation, is formulated for the Hamburg climate model ECHAM [*Roeckner et al.*, 1992]. The key parameters of the snow-cover model are snow albedo, description of liquid water (storage capacity, transmission rate), turbulent fluxes at the snow surface, heat conduction (internal heat conduction and heat exchange with ground), new snow density and density changes due to aging, and the choice of snow layers. Wind drifting processes and an additional form of metamorphism, the wind compaction, have to be considered in areas with high wind speeds. In order to obtain the simulation quality which is necessary for climate studies when using observed forcing data, the precipitation measurements should be corrected and information concerning the precipitation type should be given in the input data sets. A further improvement of the model could be reached by the implementation of a more sophisticated transmission scheme for liquid water, the use of a wind dependent new snow density, and the modification of snow albedo due to different rates of pollution.

1. Introduction

Since snow cover strongly affects both the global hydrological cycle and the atmospheric processes there is a need for reliable snow-cover models in a wide range of fields. From the hydrological point of view, for example, the appearance of snow cover leads to a temporal shift in the runoff. According to the snow-cover duration, this shift can last from hours to days up to a seasonal shift in the runoff from autumn to spring. Unusually deep snow covers, or a shift in the snowmelt period toward a period with intensive rainfall, can lead to heavy floods. Models are important tools for predicting such situations. Furthermore, in areas with a rare observational database, snow-cover models can be useful tools for interpreting satellite pictures, traditionally problematic due to the similar radiation properties of clouds and snow cover in most wavelengths (e.g., if measurements for 1.6 μm are not available), and for reconstructing historical snow data [*Brown and Goodison,* 1994, 1996] The detection of snow anomaly patterns is also important for applied environmental research (e.g. pollution of snow cover as a control parameter of the air quality) and for regional industry [*Gray and Male,* 1981; *Pomeroy and Gray,* 1995].

Considering the snow cover as part of the climate system makes it necessary to focus on both the hydrological properties and the snow-surface parameters determining the lower atmosphere boundary (albedo, temperature). The alteration of radiation properties and turbulent heat fluxes at the Earth's surface directly affects the atmospheric energetics and circulation.These changes can, in turn, have repercussions on the snow cover and thus lead to complex feedback mechanisms. Several early studies [e.g., *Dewey,* 1977; *Namias,* 1985] and recent investigations [e.g., *Groisman et al.*, 1994a, b] have noted a high correlation between changes in air temperature, circulation and snow cover. Since snow cover transforms the high-frequency atmospheric disturbances into low-frequency variability, it seems a suitable indicator for detecting climatic change. This assumption is supported by the observed snow data, which show significant changes on the decadal scale in both local [*Rohrer et al.,* 1993; *L. Mercalli and S.Paludi,* manuscript in preparation; *Brown and Goodison,* 1994] and hemispheric [*Robinson et al.,* 1991] snow parameters, which are caused by the increase in temperature due to the enhanced greenhouse effect, a lower albedo due to higher air pollution, and probably other reasons.

On the other hand, investigations [e.g., *Randall et al.,* 1994] have shown that feedback mechanisms estimated by different atmospheric models cover a wide range of responses to any given snow anomaly. Numerical studies mainly depend upon the complexity of the snow model and the formulation of the atmospheric model. The response simulated by different atmospheric circulation models to the same prescribed snow anomaly ranges from a strong positive feedback to a weak negative feedback.

Despite the obvious importance of snow cover for climatic processes and feedback mechanisms some sophisticated atmospheric models (e.g., the Hamburg global climate model ECHAM [*Roeckner et al.,* 1992]) simulate the snow processes with a relatively simple energy balance model. These models do not resolve the internal snow processes which partly determine the snow surface properties and the atmosphere-snow interface fluxes. The snow aging processes are parameterized by means of changes in the snow albedo only, where the albedo is either formulated as a function of time or temperature. Using a temperature-dependent snow albedo [e.g., *Roeckner et al.,* 1992} has the consequence that the snow is unrealistically renewed when the temperature drops.

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This can lead to inaccuracies up to 0.4 to 0.5 in spring, where the albedo values determine the ablation behavior and the end of the snow-cover period [*Loth and Graf*, this issue].

This paper describes results performed with a multi layered snow-cover model which has been developed for climate studies. With a view to global application, the simulation quality and the main physical parameters were determined under different climatic conditions. The atmospheric and snow observations used in these investigations were taken from a snow-monitoring station in the French Alps (Col de Porte) and four standard synoptic stations (Svalbard Lufthavn, Sodankylä, Gander, Zugspitze). The results of the model performances and the main candidates for the differences between the simulated and observed values are discussed. A summary is given at the end of the paper.

At the end of this section we would like to give some comments concerning the verification potential of the data sets and the goal of this paper. To verify the accuracy of any snow-cover model which is to be used in climate studies, simulated results should be compared with observations of both the hydrological properties (snow mass, duration, snow depth, snowmelt runoff) and the parameters controlling the energy and mass fluxes at the snow surface (albedo, snow-surface temperature, evaporation rate, or the fluxes themselves). During the test phase of this model, however, only one comprehensive data set was available to the authors. We therefore decided to use data sets from synoptic stations for representing different climatic conditions. Snow depth is the only snow parameter which is routinely recorded at these stations. Since other important parameters like snow mass, surface temperature, and albedo are not measured, this study can not show exactly which single inaccuracy is caused by which error. However, by means of a wide range of numerical sensitivity experiments, we were able to identify both the parameters which mainly influence the snow simulations under these different climatic conditions and the main candidates for producing the differences between the observed and simulated values. Thus, despite the existing uncertainties, we could determine the simulation quality, find out the main model deficits and suggest a list of parameters which should be included in both future field measurements programs and snow-model intercomparison projects.

2. Description of the Model

The snow-cover model [*Loth et al.*,1993], the numerical (Figure 1) and physical structure (Figure 2) of which is schematically given, is suitable for stand-alone simulations as well as for implementation within global atmospheric models. Designed as a multi layered model, it is based on fundamental physical principles and resolves the internal snow processes (water-vapor diffusion, transmission and storage of liquid water, heat conduction, melting, freezing, and aging). The snow mass balance is determined by the snowfall intensity (in mass units not depth values per time unit), the rain intensity, the evaporation at the snow surface, and the runoff. The energy budget is influenced by the short-wave and long-wave radiation balances, the turbulent heat fluxes, the ground heat flux, and the heat gain due to freezing of rainwater. The energy fluxes which directly depend on temperature are formulated by means of a linearized Taylor development against the initial value of the time step. This formulation is important, since the coupling between snow cover and atmosphere is strongly influenced by temperature, in particular, by the temperature at the snow surface.

Snow is considered as a mixture of all three phases of water and air. Its porous structure strongly influences the thermal properties of snow. Since the water-vapor pressure is set to the saturation

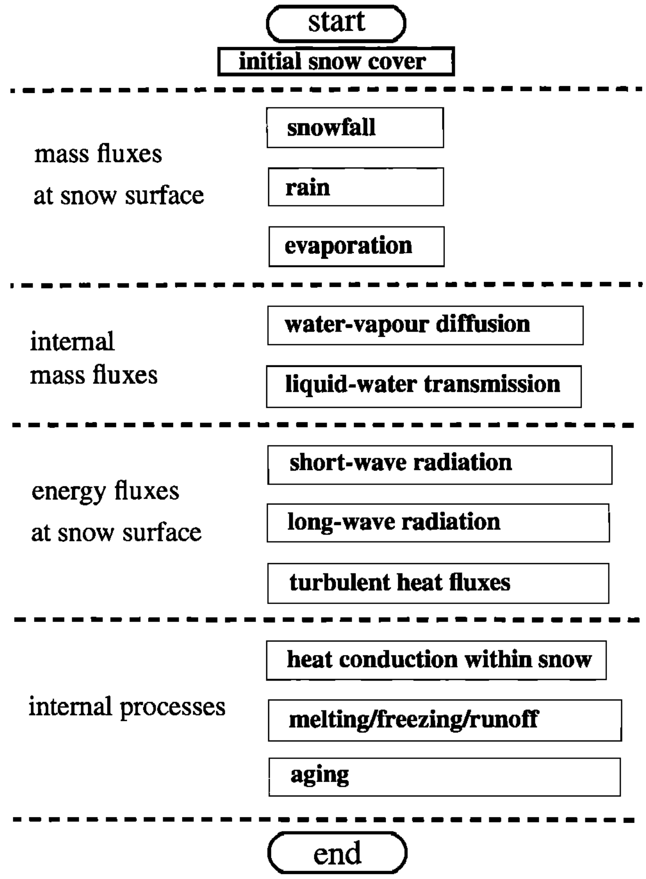


Figure 1. General structure of the multi layered snow-cover model [*Loth et al.*, 1993].

value over ice (for dry snow) or water (for snow with liquid water), the water-vapor concentration is a diagnostic variable and depends on temperature only. The prognostic variables of the model are the depth of snow layers, snow albedo, snow temperature, density of dry snow, and liquid-water content. The number of layers depends on the vertical structure of the snow cover. In general it ranges from two to five layers. The time step is chosen according to the atmospheric model with which it is coupled, or to the frequency of the observations. A snow cover is initialized when snowfall leads to a snow depth of at least 3 mm.

In order to accurately separate the simulation of internal changes in temperature from the simulation of melting or freezing processes, the model distinguishes between cold and wet snow. Owing to the aging processes, snow can be transformed into firn snow and ultimately into layers of compact ice. The three different types of snow, which are considered in the model's physics, are defined as follows: (1) cold snow without liquid water (penetrating water immediately freezes), (2) wet snow with a fixed temperature of $0^{o}C$, and (3) layers of compact ice, having a fixed density of 920 kg/m^3.

The storage capacity for liquid water is parameterized according to *Anderson* [1976] and decreases linearly from 8% to 10% in the case of new snow to 3% to 5% for metamorphosed snow. A vertical redistribution of liquid water is caused by the capillary and gravity forces. It is considered by assuming a transmission proportion of 1% of the liquid-water mass of each layer. This proportion is redistributed throughout the vertical column according to the depth of the layers. The processes of metamorphism (settling, compaction) are parameterized in terms of the relative den-

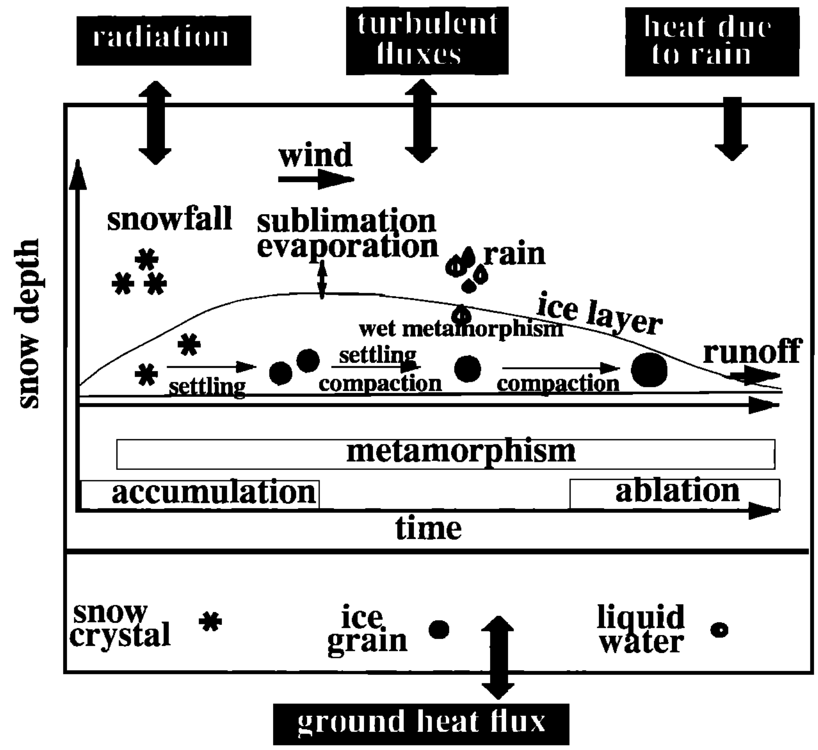


Figure 2. Schematic representation of the internal snow processes and the exchange fluxes between snow cover, atmosphere and ground which have to be considered in climate studies.

sity changes [*Anderson,* 1976]. In wet snow conditions the aging processes of the snow density are accelerated, and to account for this, the aging coefficient is double that of dry snow [*Jordan,* 1991].

Despite the high spectral dependence [*Wiscombe and Warren,* 1980], the snow albedo is described by an integral value. For clear-sky conditions and zenith position of the Sun the snow albedo is parameterized by a linear function of time. The aging coefficient is set to -0.006/d for dry snow and -0.071/d for melting snow covers with a depth below 25 cm [*Gray and Landine,* 1987]. For deep snow covers an exponential approach is used [*Verseghy,* 1991]:

$$\alpha(t) = 0{,}5 + [\alpha(t-1) - 0{,}5]\exp[-(0{,}24\Delta\tau)] \quad (1)$$

The modification of the snow albedo by clouds and/or Sun height is parameterized following *Siemer* [1988]:

$$\alpha_s = \alpha_0 + (1-\alpha_0)\alpha_0^3 \left\{N_{cl}^2 + \left(1 + C_\alpha N_{cl}^2\right)\exp\left[1 - (1 - \sin(e_\alpha))^2\right]\right\} \quad (2)$$

where N_{cl} is the amount of cloud, the coefficient C_α is set to -1.3, and e_α is the minimum of the values $\pi/3$ and the Sun height ϕ. The clear-sky snow albedo α_0 increases due to snowfall by 0.1 per 1 cm new snow depth. The model's maximum value for the clear snow albedo is 0.85.

In the ECHAM version of the multi layered snow model the surface albedo α_{surf} is estimated as function of the snow albedo α_s, the background albedo α_{sb}, and the snow water equivalent (swe) following the approach proposed by *Roeckner et al.* [1992]:

$$\alpha_{surf} = \alpha_{sb} + (\alpha_s - \alpha_{sb})\frac{swe}{swe + s_{crit}} \quad (3)$$

The critical value s_{crit} is set to 0.01 m. A vegetation masking effect on the snow albedo is introduced for forest areas, where the snow albedo α_{sf} is estimated as a function of the fractional forest area a_f (a_f=[0,1]), the albedo of snow-covered forest α_f, and the snow albedo in the forest-free part of the grid cell are α_s using the following formula:

$$\alpha_{sf} = (1 - a_f)\alpha_s + a_f\alpha_f \quad (4)$$

The fractional forest area a_f is taken from a data set compiled by *Matthews* [1983]. The albedo of snow covered forest α_f is set to 0.35.

At the end of each time step, neighboring snow layers are merged if the depth of one layer is less than 3 mm. In order to avoid a high number of layers, neighboring layers are also combined according to the following criteria: (1) if their temperature difference is less than 3 K, (2) if they consist of the same snow type, and (3) if the density difference is less than 150 kg/m^3, whereby the snow layers at the interfaces with the atmosphere and with the ground are excluded from this procedure. If the depth of the snow cover is less than 2 cm, all snow layers are merged at the end of the time step and the snowpack is divided into two snow layers in the middle of the snow depth with the grid points at the interfaces to atmosphere and soil, respectively.

In the ECHAM model version a fractional coverage of a grid square is considered in the estimation of the evaporation, the turbulent heat fluxes, and the surface albedo. In contrast to albedo, which is modified in forest areas only (4) and for which the horizontal inhomogenity of the grid square and the influence of the underlying soil are described by a linear combination of background and snow albedo using the critical value s_{crit} (3), the evaporation as well as the turbulent heat fluxes are separately estimated over four different fractions of the grid square [*Roeckner et al.,* 1992]: (1) c_{sn}, the fraction covered with snow, (2) $(1-c_{sn})c_l$, the fraction covered with water in the skin reservoir, (3) $(1-c_{sn})(1-c_l)(1-c_v)$, the fraction covered with bare soil, and (4) $(1-c_{sn})(1-c_l)c_v$, the fraction covered with vegetation.

The snow-cover fraction c_{sn} depends on the snow water equivalent (swe),

$$c_{sn} = min\left(1, \frac{swe}{swe_{crit}}\right) \quad (5)$$

where the critical snow water equivalent swe_{crit} is set to 0.015 m.

If the snow-cover model is implemented within a general circulation model, the energy and mass fluxes at the snow surface are estimated by the atmospheric model. The energy exchange at the snow base is directly calculated. Temperature diffusion processes are estimated for the entire vertical column consisting of snow and soil layers. A list of the snow-cover model input parameters, as used when coupled to an atmospheric model, is given in Table 1.

In stand-alone simulations the input data set consists of atmospheric and soil measurements. In addition, local information concerning the properties of the soil and vegetation at the measuring site are needed. In particular, the type of soil/vegetation, the heat conductivity, and the heat capacity of soil are necessary. Initial profiles of soil temperature can be provided as measurements or climatological values. The parameters of the input data set are given in Table 2.The energy and mass fluxes at the snow surface are estimated as follows: The incoming short-wave radiation Q_G is calculated by means of a δ-Eddington radiation-transport model [*Schult,* 1991]. The influence of clouds, which lead to a spectral shift of the incident radiation and a different proportion of diffuse to direct radiation, is considered using the approach

$$Q_G = Q_{G0}(1 - f_{cl}N_{cl}) \quad (6)$$

Q_{G0} and N_{cl} are the incoming short-wave radiation at the snow surface for clear-sky conditions and the total amount of cloud, respectively. The coefficient f_{cl}, which depends on the type of clouds, is set to 0.14 for high and medium cloud levels and to 0.67 for low clouds.

The downward directed long-wave radiation is calculated as a sum of the clear-sky radiation and the radiation emitted by clouds. For clear-sky conditions the incoming radiation is a function of the atmospheric water-vapor pressure and the air temperature [*Idso,* 1981]. The radiation emitted by clouds depends on several factors: the temperature at the cloud base, the emissivity of the clouds, the transmissivity of the atmosphere in the water-vapor window 8 μm to 14 μm, and the amount of cloud [*Kimball et al.,* 1982]. The emissivity of clouds is set to 1 for low clouds, 0.75 for medium clouds, and 0.50 for high clouds [*Siemer,* 1988]. A detailed description of the incoming long-wave radiation parameterization is given by *Loth et al.* [1993].

Table 1. Input Parameters of the Snow-Cover Model for Cases Coupled With Atmospheric Models

Symbol	Input Parameter
Q_G	incoming short-wave radiation
Q_{la}	incoming long-wave radiation at snow surface
Q_h	turbulent flux of sensible heat
M_{vS}, Q_e	turbulent water-vapor flux and turbulent flux of latent heat
M_s, M_r	rate of snowfall and rain intensity
T_B	temperature of uppermost soil layer
dQ_h/dT_s	derivative of sensible heat flux with respect to snow surface temperature
dM_{vS}/dT_s, dQ_e/dT_s	derivative of turbulent water-vapor flux and latent heat flux with respect to snow surface temperature

Table 2. Input Parameters of the Snow-Cover Model in Stand-Alone Simulations

Symbol	Input Parameter
T_A	air temperature at 2 m
v	dew-point temperature or relative humidity at 2 m
w	wind speed at 10 m
p_A	air surface pressure
N_{cl},I_{clG}	cloud parameters (total amount of clouds, type of clouds)
P	intensity of precipitation
Q_G	incoming short-wave radiation
T_B	soil temperature

The turbulent fluxes at the snow surface are estimated in accordance with the Monin-Obukhov similarity theory, taking into account different roughness lengths for momentum, heat and water vapour. The dimensionless stability functions $\psi_M(z/L)$, $\psi_H(z/L)$ and $\psi_L(z/L)$ are parameterized according to *Beljaars and Holtslag* [1991]. Here, z and L are the reference level and the Obukhov length, respectively. The reference level is set to 2 m. The wind speed is extrapolated from 10 m to 2 m assuming a logarithmic wind profile. The proportions of the roughness length for heat and for water vapour relative to that of momentum are taken from *Andreas* [1987]. They are determined in dependence with the Reynolds friction number. The roughness length for momentum is set to 10^{-4} m [*Wieringa,* 1993].

The ground heat flux is determined by the vertical gradient of the soil temperature and the thermal conductivity of soil. The thermal conductivity of soil ranges between 0.08 W/(mK) for dry moor and 3.35 W/(mK) for granite [*Linke and Baur,* 1970]. It is a function of both type and liquid-water content of the soil. The default value of the model is set to 0.3 W/(mK), which is typical for soils of dry sand and loam. For snow on sea ice or land ice the thermal conductivity of the ground is set to 2.26 W/(mK).

The type of precipitation depends upon the air temperature at 2 m [*Wilhelm,* 1975]. A mixture zone with a coexistence of both types of precipitation is assumed for the interval between -1°C and 4°C. The temperature of all the following hydrological components, falling precipitation, in snow cover infiltrating rainwater, and accumulated new snow, is set to the wet-bulb temperature of air. In the case of snowfall, a fresh-snow layer is opened when the new snow depth exceeds 3 mm. Otherwise, the precipitation is added to the uppermost snow layer. The new snow density is estimated as function of the wet-bulb temperature using a parameterization proposed by *Anderson* [1976]. For wet-bulb temperatures below -15°C, for which the parameterization does not provide snow density values, we use a linear interpolation between -15°C, 50 kg/m^3 and -30°C, 30 kg/m^3. The minimum snow density is set to 30 kg/m^3, also for lower wet-bulb temperatures, since wind rounding activities remove the fine-structured parts of the floating snowflakes and thus avoid smaller densities of the accumulated snow in the presence of wind movement. Very small new snow densities of about 10 kg/m^3 are observed under calm cold conditions [e.g., *Wilhelm,* 1975]. However, since the dependence of the new snow density on the wind speed is not taken into account in the standard version of the model, the assumption of a minimum snow density seems to be a good approximation.

3. Description of the Data Sets

The geographical position of the stations from which data sets are used is shown in Figure 3.The characteristics of these different locations are briefly outlined below.

3.1. Snow Monitoring Station Col de Porte

A comprehensive data set including most of the climatically relevant snow parameters like mass and depth of snow cover, runoff, albedo, and snow surface temperature is taken by the Centre d'Etude de la Neige in Grenoble at Col de Porte (45 °N, 6 °E, 1320 m) in the French Alps. This site is characterized by a continuous snow cover in winter, usually lasting from late autumn (late November to the beginning of December) to late spring (end of April to mid-May), loamy soil, grass vegetation and some relatively tall trees at the edge of the measuring field [*Brun et al.*, 1992]. We analyzed the simulation results for two different periods (1988-1989, 1993-1994). Since the results are similar, we only focus on the data set which covers the period December 17, 1988 to May 8, 1989, in this paper.

The measuring frequency for both the atmospheric and the snow data is 1 hour. The snow depth was measured by means of two different methods: hourly with an ultrasonic sensor and once a week from a manual snow pit. According to *Brun et al.* [1992], the differences between the two measured values were caused by the compaction of the underlying grass vegetation and the fact that the ultrasonic sensor may have been more exposed to the Sun than the pit measurements. Further, locally different accumulation rates due to a horizontal redistribution of previously deposited snow may have contributed to the deviations in the measurements.

In addition to the input data necessary for model runs (Table 2), the incoming short and long wave radiation was measured. This provides the advantage to reduce the simulation uncertainty in the snow-independent part of the model.

Since the mass of snowfall and rain are determined separately, assessing a snow-rain criterion, which at midlatitudes may drastically increase the model inaccuracy [e.g., *Loth et al.*, 1993], also becomes unnecessary. In order to correct the systematic measuring error, the precipitation data recorded for Col de Porte were increased by 15% [E.Brun, personal communication]. For comparison reasons, the simulation results are presented for both cases with and without a correction of the precipitation measurements.

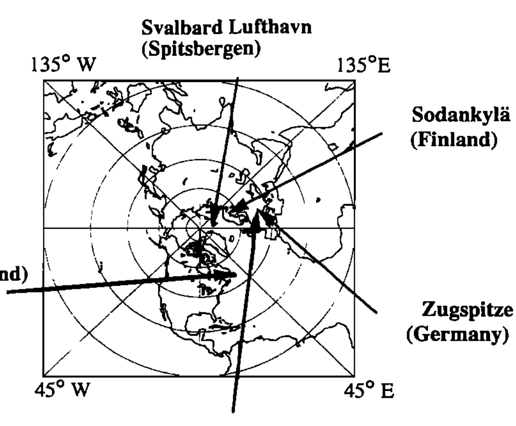


Figure 3. Measuring stations used for the verification of the multi layered snow-cover model.

Table 3. Initial Snow Profile Used in the Simulations for Col de Porte

Depth, cm	Density, kg/m^3	Temperature, K	Liquid Water, kg/m^3
5	200	270.15	0
10	240	270.15	0
5	300	270.15	0
5	320	270.15	0

Because of the lack of profile measurements on December 17, 1988, assumptions had to be made concerning the initial snow conditions. Although we could not check the initial values directly, we could compare the measured and simulated snow profiles on December 23, 1988. To test the sensitivity to the initial snow conditions, we used different temperature and density profiles. We also tested the possibility of taking the measured air temperature at 2 m and the wet-bulb temperature to initialize the snow-surface temperature. The simulations which are presented in this paper were initialized using a four-layer snowpack of dry snow at a temperature of 270.15 K (Table 3).

Unlike for the simulations with the data sets from the synoptic stations, for which we prescribe the soil temperature at 5 cm depth, we coupled the snow-cover model with the ECHAM soil model [*Roeckner et al.*, 1992] for the Col de Porte calculations. This soil model uses five thermal layers with depths of 6.5 cm, 25.4 cm, 91.3 cm, 290.2 cm, and 5700 cm and a bucket model for the soil hydrology. No-flux conditions are assumed for the lower boundary of the soil. The thermal soil characteristics depend on the soil type only and were set to the values of loamy soil in the case of Col de Porte. The measured values of the initial soil temperature (at 20 cm and 50 cm) were linearly interpolated for the upper three soil layers, whereas the initial temperature of the lower two soil layers were set to the climatological ECHAM value for this region. Taking into account the influence of the gras vegetation, we assumed the existence of a thin air layer between the top of the grass and the soil surface. The thermal conductivity at the snow-soil interface was set to that of air (0.0242 W/(mK)).

3.2. Four Synoptic Stations

In order to test the snow-cover model under conditions different from those of Col de Porte, we analyzed data from four synoptic stations, each representing special types of snow or climatic conditions. The data sets analyzed for the synoptic stations cover the period September 1, 1986, to June 1, 1990. The values of air and dew point temperature, the wind speed, and the surface pressure are determined hourly. Observations of the cloud properties (type and total amount of cloud) are taken at 3-hour intervals. Precipitation is recorded every 6 hours. The snow depth is measured daily. Since the soil temperature was not measured, assumptions had to be made concerning the lower boundary of the snow model. The sensitivity of the snow-cover simulations to the soil temperature were tested for each station. It will be briefly described below when we discuss the simulation uncertainties for these stations.

Sodankylä (Finland). Winter conditions are determined by polar night and low temperatures. Because of the low temperatures, the whole winter precipitation falls as snow, and melting processes only occur in spring and early summer. Missing shortwave radiation in winter leads to a separation of the thermal and the radiative processes during the cold season.

Svalbard Lufthavn (Spitsbergen). As at Sodankylä, winter conditions are determined by polar night and low temperatures.

Frequently, high winds cause snow drifting and wind compaction of the deposited snow, effects which are of minor importance at Sodankylä. Despite its relatively northern geographical position, the air pollution at Svalbard Lufthavn is similar to that of the midlatitudes.

Gander (Newfoundland). Gander is a maritime midlatitude station, which is affected by snowstorms and has a relatively deep snow cover compared with other lowland midlatitude locations.

Zugspitze (German Alps). At Zugspitze, extreme conditions occur due to the station's high altitude. Snowfall and temperatures below the freezing point can occur throughout the whole year. Compared to Col de Porte, the mean wind speed is significantly higher, the mean temperatures are lower, and the snow cover lasts longer.

3.3. Uncertainties of This Study When Using Synoptic Data

Besides the measuring errors, uncertainties occur due to the estimation of the interface fluxes of the snow cover. These uncertainties are not a direct part of the snow-cover model, but they do influence the accuracy of the simulated snow cover. In the absence of observations, the incoming short- and long- wave radiation are estimated with state-of-the-art parameterizations (cf. section 2). The estimated clear- sky component of the downward directed short wave radiation has been verified by *Schult* [1991]. The parameterization of the incoming long wave radiation has been tested against other state-of-the-art parameterizations by *Yang et al.* [1997]. Both studies show reasonable results. Since there is no modification in the schemes and these parameters as well as the choice of the precipitation type are estimated by the atmospheric part in a coupled model without any direct impact of the snow cover, we assume that these components are reasonable. In contrast, the turbulent fluxes and the uncertainties which occur due to errors in the estimated ground heat flux directly depend on the snow properties and provide a direct interaction with the atmosphere when the snow model is implemented into an atmospheric model.

Since we only use one parameter for the model verification at the synoptic stations, an error compensation cannot be excluded. In order to identify the important simulation parameters under the different conditions and to find out the contribution of each error candidate to the inaccuracies, we have performed a range of sensitivity studies taking into account the probable uncertainties. Unlike for Col de Porte, where we used corrected precipitation data for all sensitivity tests, we did not correct any potential error for the synoptic stations. We analyzed the influence of a potential error candidates on the simulation results and the way in which the particular parameter should change in order to correct the entire inaccuracy.

Uncertainties in the snow mass balance can originate from the uncertainties of the snow-rain criterion, an imprecise correction of the precipitation measurements, simulation uncertainties in liquid-water parameters of the model (storage capacity and transmission rate) as well as errors in the evaporation estimation. All these components should be measured for the verification of a snow-cover model. Since at synoptic stations the snow depth, and not the snow mass, is measured, all uncertainties concerning the snow density are superimposed, if the simulated snow depth changes differ from those observed. The uncertainties concerning the snow density are inaccuracies in the new snow depth and in the metamorphism rates. Observed snow depth changes can also be caused by a snow redistribution, which is not included in the model.

If only observed snow depths are available for the verification of any snow model, uncertainties in the estimated energy balance cannot be directly identified. However, a simulation error in the energy balance has to be assumed if the observed and modeled decrease in snow depth differ and it is known that melting occurred in particular in cases in which the ablation period is shifted although the simulated and observed snow depth were similar just before the melting. Detected uncertainties in the estimated energy balance may originate from inaccuracies in the net short-wave radiation (albedo or solar radiation), net long-wave radiation (surface temperature, incoming long-wave radiation, emissivity), and turbulent heat fluxes (surface temperature, estimation of the stability functions). The snow-surface temperature is, in turn, influenced by the energy budget of the snow cover but also is determined by the internal heat conduction, the liquid-water content of the different layers, and the snow mass. The lower boundary (ground heat flux) directly influences the temperature and liquid water at the snow base. It might directly influence the snow mass due to melting processes at the snow base and even the snow surface due to the internal snow processes (heat conduction and liquid-water transmission).

An error in the incoming short wave radiation can be compensated by error in the estimated snow albedo. The short wave radiation is mainly influenced by clouds, whereas the albedo is a function of the grain size (parameterized as function of time), the liquid water and the pollution rate. In the case of thin or patchy snow covers, the albedo is determined by the underlying soil. In vegetated areas the snow albedo is masked by the vegetation. The influence of all these parameters cannot be exactly determined in sensitivity studies in the absence of measurements of solar radiation and albedo. The assumed uncertainties in the albedo can be superimposed by errors in the estimated solar radiation. This problem is of importance for midlatitude stations, where the short-wave radiation strongly influences the snow-cover development throughout the whole winter season and during the ablation period in Arctic regions.

4. Results

4.1. Model Consistency (Col de Porte)

The model simulates the snow cover properties fairly accurately. The comparison of observed and simulated snow depth values (Figure 4a) shows that both the shape of the snow depth curve and the duration of the snow cover are in good agreement. The simulated water equivalent (Figure 4b), which expresses the snow mass in water column depth, as well as the calculated runoff (Figure 5) document a relatively precise estimation of the snow mass balance. The two snow parameters which control the energy exchange between the atmosphere and the snow cover, namely, snow surface temperature (Figure 6) and snow albedo (Figure 7), are also well reproduced. These results lead to the assumption that the model is physically consistent.

The differences between model results and observations are caused by both simulation and measurement uncertainties. The main simulation error is an inaccuracy in the liquid-water processes which results in an accelerated runoff (see below). The highest measuring uncertainties seem to be an accurate determination of the precipitation. The measuring error of precipitation, which results in negatively biased measured values, usually ranges between 10% and more than 100% and depends mainly on the type of the precipitation, the wind speed and the measuring device [*Goodison and Yang*, 1996]. As was mentioned above, for comparison, the calculated curves are shown for the simulation without any correction of the precipitation data (thick grey line) and when using the correction values of 15% (solid black line). The correc-

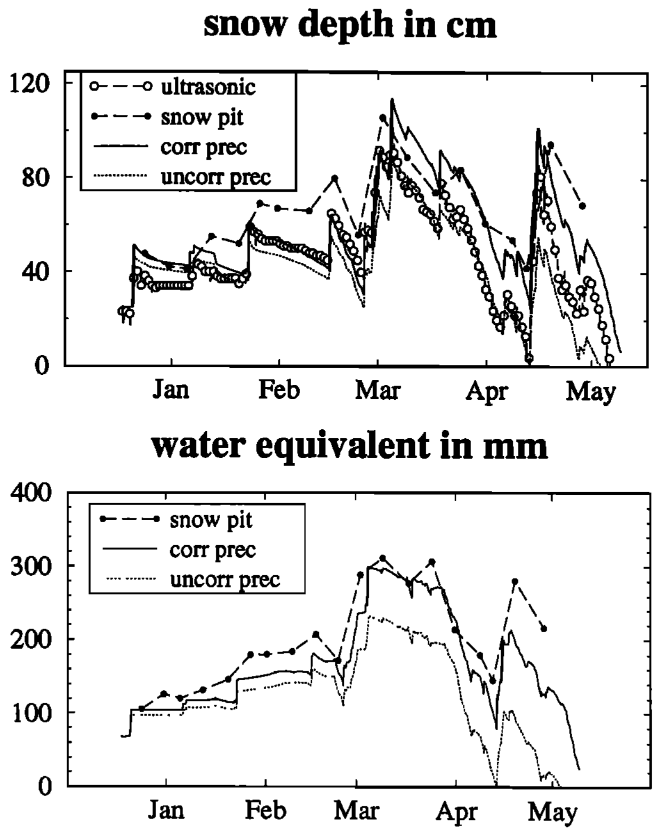


Figure 4. Measured and simulated values of snow depth and water equivalent at Col de Porte for the time period December 17, 1988, to May 8, 1989.

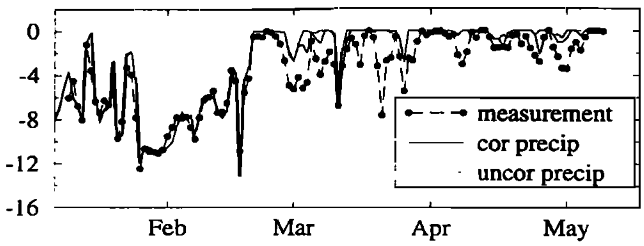


Figure 6. Daily mean values of measured and simulated snow-surface temperature (in degrees Celsius) at Col de Porte for January 8, 1989, to May 8, 1989.

tion is, of course, only an approximation. To be more precise, each single value should be corrected separately. In both snow depth and water equivalent, the correction leads to a general shift in the calculated curves toward higher values, whereby the differences between the curves increase with time (Figure 4). The two estimated curves of runoff (Figure 5), surface temperature (Figure 6), and snow albedo (Figure 7) also differ, but the changes due to the correction of the precipitation data are small. Differences in mean daily surface temperature, which range between 0.1 K and 2 K, occur at the beginning of March, in mid-March, in mid-April, at the end of April, and at the beginning of May. The estimated albedo only changes (from the value for snow of about 0.68 to the surface albedo which is set to 0.2) because of a short-term disappearance of snow in April and the earlier end of ablation in May in the case of using uncorrected precipitation data. The runoff behavior of the model is very similar in both simulations. Owing to the smaller accumulated snow mass in the case of using uncorrected precipitation data, the modeled runoff is + 2 mm to 6 mm higher at the time of initial melting from February to the beginning of April and increases by about 20 mm in mid-April. The following discussion is confined to the simulation using the corrected precipitation data.

Throughout the whole period, the calculated snow depth (Figure 4a) either ranges between the two observed values or it fits one of the measured curves. The model results correspond well with the snow pit measurements from December to mid-January and from the end of March to the end of April, a time during which the ultrasonic sensor measured smaller values (by -2 cm to -10 cm and -10 cm to -20 cm, respectively). In contrast, from mid-January to the first third of March the simulated curve is nearly identical to the measurements of the ultrasonic sensor, and the values taken by the snow pit exceed both by 10 cm to 20 cm. At the beginning of March the simulations exceed both measurements by about 20 cm.

Inaccuracies in the simulated snow depth mainly result from uncertainties in the liquid water processes, which lead to a wetter bottom layer than observed (Figure 8) and, in the end, to an accelerated runoff in comparison with the measurements (Figure 5). This can be caused by uncertainties in the storage capacity for liquid water or an erroneous simulation of the heat conduction at the snow base, which leads to errors in the rate of phase transformation and/ or an erroneous assumption of the soil temperature. An improvement in the simulation occurs if the parameterization of the storage capacity for liquid water is changed for densities greater than 200 kg/m^3 to 10% for densities greater than 200 kg/m^3, 20% for densities above 300 kg/m^3, and 30% for densities above 400 kg/m^3.

The simulation uncertainties are also influenced by the estimation of the new snow density, which together with the precipitation data determines the new snow depth. The estimated new snow density may differ from reality, since the dependence of the new snow density on the wind speed and the possibility of mixing falling with redistributed snow, which leads to higher snow densities

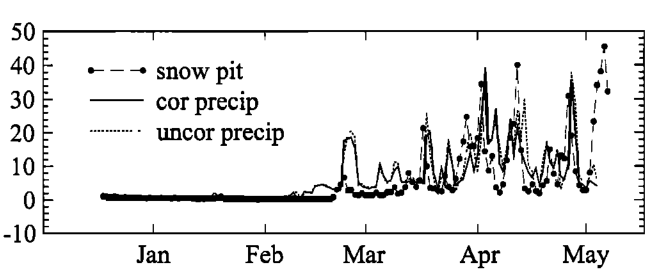


Figure 5. Measured and simulated runoff (daily sums in millimeters) at Col de Porte for December 17, 1988, to May 8, 1989.

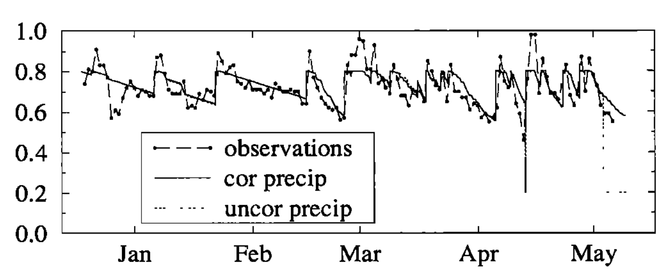


Figure 7. Observed and simulated values of snow albedo at Col de Porte for the time period December 17, 1988, to May 8, 1989.

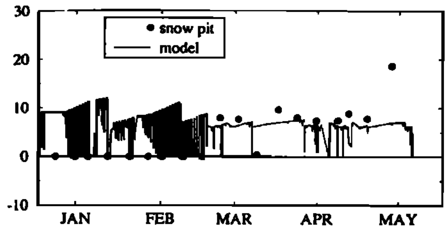


Figure 8. Liquid-water content (in kg/m^3) at the snow base at Col de Porte for 1988-1989. The measurements were taken by a manual snow pit.

of the accumulated snowpack (usually 150 kg/m^3 to 350 kg/m^3) compared to pure new snow conditions (typically, 50 kg/m^3 to 100 kg/m^3), are not considered in the model.

Concerning the measuring uncertainties, in addition to the precipitation measuring error, locally different accumulation rates, and a redistribution of already deposited snow may considerably decrease the reliability of the measurement. The overestimation of snow depth at the beginning of March may probably be caused by the occurrence of a horizontal redistribution of snow. The compaction of the vegetation below the snow cover, which leads to a negative bias of the snow depth measurements (up to -2 cm), seems to be of less importance.

As with the snow depth over some time intervals of the simulation period, the simulated water equivalent (Figure 4b) is seen to be systematically smaller than observations, when compared with the actual pit measurements. The differences between the simulated and observed values are between -10 mm and -20 mm from December to mid-April. They increase to about -50 mm during the ablation period. An analysis of the runoff data (Figure 5) shows that these inaccuracies (like, in part, the above-mentioned differences in snow depth) are caused by an accelerated runoff in early spring (February, March) and at the beginning of the melting period.

The snow-surface temperature (Figure 6), which is an important parameter describing the energy exchange between the snow cover and the atmosphere, is also estimated fairly accurately. In winter the mean daily differences between the observed and simulated values are very small (1 K and 2 K). Larger differences (4 K to 8 K) occur in spring and are caused by a wetter surface layer in comparison with the observed conditions, resulting in an overestimation of the mean daily surface temperature. These deviations are drastically reduced by using an improved parameterization of the liquid-water transmission. We implemented a depth-dependent redistribution function for liquid water in the new model version, assuming that, in the absence of rain or melting processes at the snow surface, the snow layers near the surface dry out due to gravity forces. One example which demonstrates the sensitivity of the model to this parameterization is shown in part 2 of this study [*Loth and Graf*, this issue].

The calculated snow albedo corresponds very well with observations (Figure 7). The changes in albedo due to both dry and wet snow metamorphism are well captured, and the assumption of linear aging under cold conditions and an exponential decrease in albedo during the occurrence of melting processes leads to a high correlation with observations. Only the extreme albedo values following new snow events are not estimated as well. This could be attained by using an exponential function for the albedo aging of cold snow as also introduced by e.g., *Verseghy* [1991]. In the days after snowfall, the processes of destructive metamorphism lead to rapid changes in the structure of the crystals. The transformation of the bizarre snow surface to round snow grains, which can be described by an exponential function, reduces the albedo far more than the aging processes of the following periods (settling, compaction, etc.). Since the albedo is already well simulated by the model, this further improvement has not been implemented.

4.2. Arctic Conditions

The climatic conditions at the Arctic stations of Svalbard Lufthavn and Sodankylä are typical for Eurasian land areas north of 60° N. The air temperature, the daily mean values of which range between -12°C and -17°C in winter, has annual extreme values of-30°C to -40°C. At Sodankylä (Finland) polar night occurs from the end of November to mid-January, and the mean precipitation rate is about 500 mm/yr. At Svalbard Lufthavn (Spitsbergen), where the mean precipitation rate of about 180 mm/yr. documents relatively dry conditions, polar night occurs from the end of October to the end of February. Despite the low mean winter temperature, warm air advection can lead to temperatures near the freezing point. The high winds, which cause snow drifting and wind compaction of the deposited snow, have mean values of 5 m/s to 15 m/s and maximum speeds ranging between 35 m/s and 40 m/s.

In the Arctic simulations for which the results are shown in this section, the soil temperature at 5 cm depth was set to the snow base temperature of the previous time step. With this assumption,

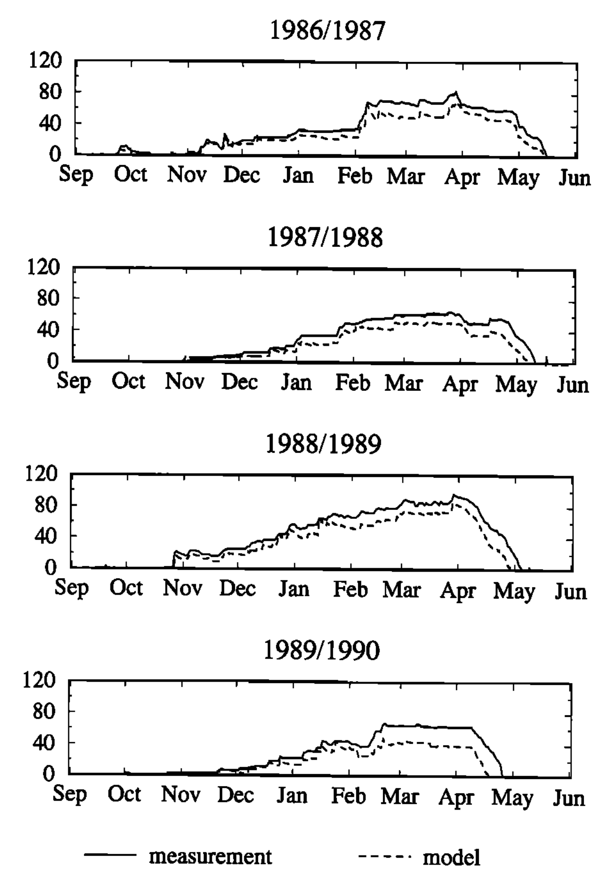


Figure 9. Observed and simulated snow depth (in centimeters) at the Finish station Sodankylä for 1986-1990.

the temperature of snow-covered soil cannot exceed 0°C, which is true if the soil is frozen during the whole snow period, and each simulated change in the snow bottom temperature occurs time shifted in the upper soil, which might be a first-order approximation for frozen soil and of at least the same reliability as using a constant temperature value. The modeled ground heat flux shows a spin-up-like period of about 10 days to 15 days, in which in extreme cases the values can reach the range of +/- (20 W/m^2 and 30 W/m^2). During the winter and spring the simulated ground heat flux has values between -5 W/m^2 and 8 W/m^2, in single cases up to +/- 15 W/m^2. The absolute values again increase to 20 to 40 W/m^2 just before the melting period in April to May. Since the assumptions of both the soil temperature and the default value of the thermal conductivity of 0.3 W/(mK) are relatively arbitrary, we have analyzed the sensitivity to the lower boundary. These test simulations were performed with different fixed values for the soil temperature (-2°C, -1°C, -0.5° C, 0° C, 0.5° C, 1° C). The results show that the different parameterizations lead to changes in the energy budget of the snow cover of +/- 1 W/m^2 to 7 W/m^2, the snow surface temperature of +/- 0.1 K to 2 K, and the end of ablation of up to 4 days.

Despite a small deficit, the snow depth is accurately simulated for both Arctic stations (Figure 9 to Figure 12). The shape of the snow cover curve is reproduced in detail throughout the whole

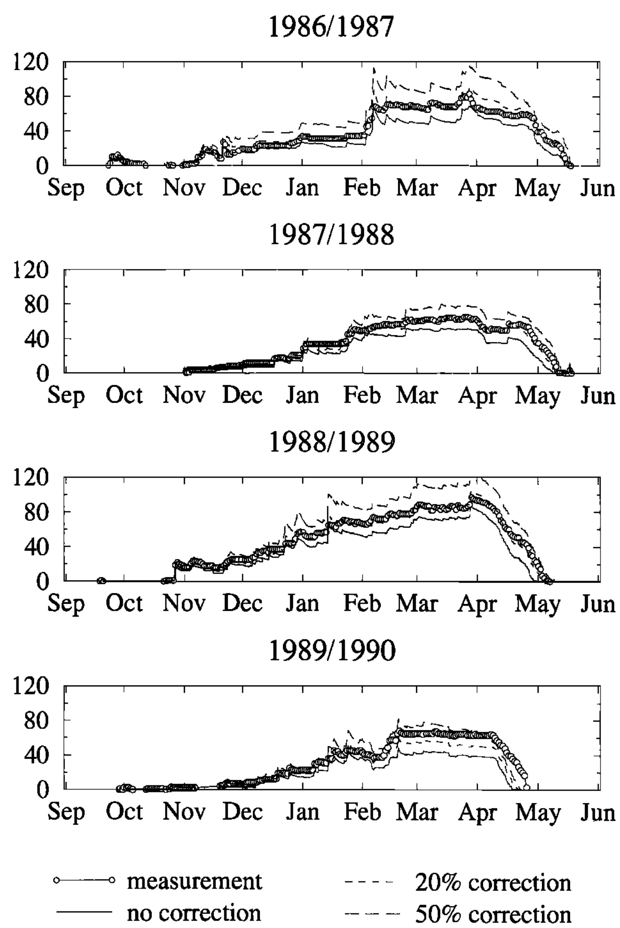


Figure 10. Snow depth (in centimeters) in Sodankylä for 1986-1990. Shown are observations, simulations with no correction of the precipitation measurements, and modeled results assuming a 20% and 50% measuring error for precipitation.

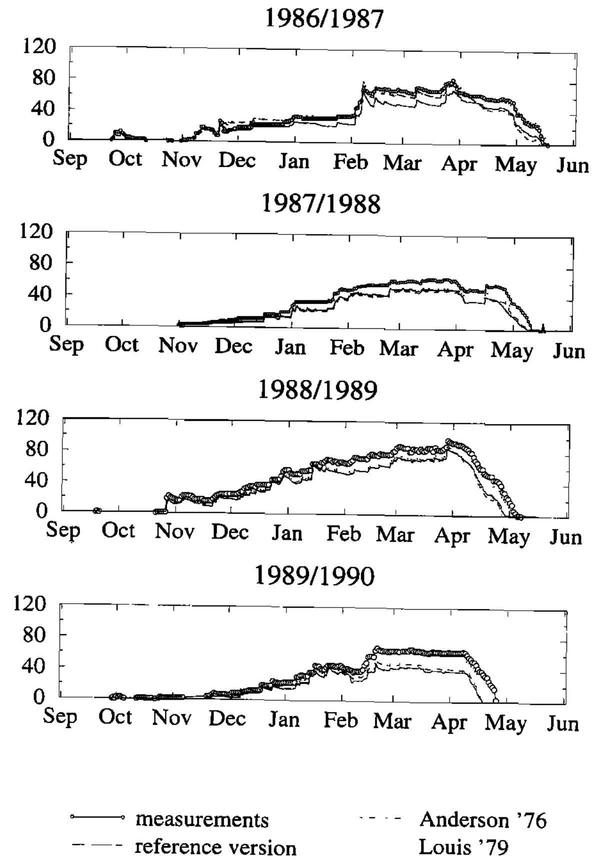


Figure 11. Snow depth (in centimeters) in Sodankylä for 1986-1990. Shown are observations and simulations using different turbulent schemes. The reference version and the scheme introduced by *Louis* [1979] are based on the similarity theory, whereas the approach proposed by *Anderson* [1976] is a bulk formula.

period of 1986-1990. This is mainly due to the fact that the snow-cover development is characterized by a strict series of accumulation, metamorphism, and ablation periods. During autumn and winter, no melting occurs, and the aging processes are limited to dry metamorphism. Dry snow can be well simulated, since the processes are reduced to the simple case of phase changes of vapor and ice. Any change in the energy balance leads to a change in the snow temperature. The liquid water influences the snow cover in spring, when precipitation partly falls as rain and warm-air advection as well as absorbed short-wave radiation induce stronger melting processes. The end of the ablation period, which occurs from mid-April to mid-May, is mainly determined by the snow mass and the albedo. Owing to a slight underestimation of the snow depth, the simulated end of the ablation period can be shifted in comparison with the observed date by some 5 to 10 days.

The simulation results for Sodankylä (Figure 9) show that at this location the above mentioned differences between modeled and observed values systematically increase with the duration of the snow cover. These deviations reach their maximum (-10 cm to -20 cm) between February and the end of the ablation period.

One main cause for this inaccuracy seems to be a measuring error in the measured snowfall. In order to demonstrate the influence of this error on the simulation of the snow cover under Arctic

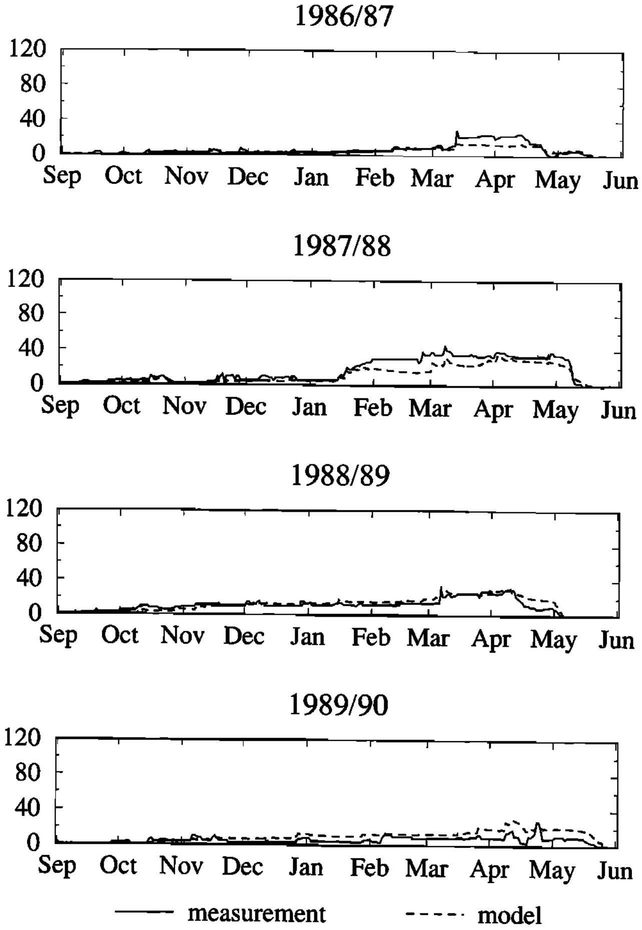


Figure 12. Observed and simulated snow depth (in centimeters) at Svalbard Lufthavn (Spitsbergen) for 1986-1990.

conditions, we compared model simulations using different correction values (1%, 5%, 10%, 20%, 50%, 75%, 100%, 150%, 200%). Results are shown for which we assumed a correction value for precipitation of 20% and 50%, respectively, the values between which the modeled curves fit best the observed snow depth without correcting any potential model inaccuracy (Figure 10) As can be seen, the maximum differences between simulations and observations are reduced to small values. A 20% correction of the precipitation data leads to an excellent correlation between estimations and measurement for 1986-1987 to 1988-1989, whereas a 50% correction results in an overestimation of the modeled snow depth for these years. In particular for 1987-1988 and 1988-1989 a 20% correction reduces the difference to small values (2 cm to 5 cm). Using corrected data for the precipitation measurements also decreases the deficit for 1989-1990, but simulated and observed snow depth values correspond well for a 50% correction whereas a 20% correction only explains half the story.

A further main candidate for the deviations of simulations from the observations is the parameterization of the turbulent fluxes at the snow surface. Figure 11 shows the influence of the turbulent scheme on the snow-cover simulations. These simulations were performed for Sodankylä using uncorrected precipitation data and, alternatively, three different turbulent schemes: the reference version (compare section 2), the scheme used in the ECHAM model [*Louis*, 1979], and a bulk-formulation proposed by *Anderson* [1976], which has been developed for snow conditions. The reference scheme and the parameterization from *Louis* [1979], which are both based on the Monin-Obukhov similarity theory, differ in the formulation of the stability functions. The results show that compared with the reference scheme implementing the Louis scheme leads to systematically higher snow depths (+2 cm to +18 cm) and a better correspondence between the simulated and observed values. Please note that these uncertainties are of the same magnitude as the probable measuring error of the precipitation data. For 1986-1987 and 1989-1990, higher values are simulated from December to the end of ablation in April/May. For 1987-1988 and 1988-1989, relevant changes only occur from April to May. Using the Anderson scheme also leads to higher snow depth, but only for December 1986 to February 1987 by about 10 cm to 20 cm and for February to mid-March 1990 by about 2 cm to 5 cm. The simulated curve is nearly identical to the results of the reference version for April to June 1987, 1987-1988, 1988-1989, and autumn and spring of 1989-1990. The differences in the simulated snow depth correspond with systematic differences between the model versions in the snow mass and the turbulent fluxes. In comparison to the reference version, the simulated water equivalent increases using the Louis scheme by only some millimeters in the accumulation period of 1987-1988 and 1988-1989, but 20 to 40 mm in December 1986 to May 1987 and the spring seasons of 1987-1990. The modeled differences result from changes in both the water-vapor flux, which directly influenced the snow mass balance, and the sensible heat flux, which influences the energy budget and the mass balance if melting occurs. Compared to the stability-dependent approaches, the turbulent fluxes estimated with the bulk formula [*Anderson*, 1976] have the highest absolute amounts. The other two parameterizations mainly differ in the case of wet snow conditions. The turbulent heat fluxes differ between the three different model versions by 5 W/m^2 to 40 W/m^2 (running average over 24 values).

In general, the observed snow cover at the Svalbard Lufthavn station (Figure 12) is also well reproduced in the model. The end of the ablation period is determined with a high degree of accuracy. Thus, even for a relatively thin snow cover of about 20 cm to 30 cm, the model shows encouraging simulation results. The systematic simulation deficit of snow depth is mainly caused by the following three factors: (1) the use of uncorrected data of measured precipitation, (2) the neglect of horizontal redistribution of snow, and (3) an overestimation of the turbulent fluxes under Arctic conditions due to snow drift extreme atmospheric stability. In contrast to 1986-1989, when the snow depth was underestimated, the snow depth was overestimated by 5 cm to 10 cm from mid-November 1989 until the end of the ablation period in spring 1990.

Like for Sodankylä, sensitivity studies were performed assuming different correction values for precipitation. The best fit without considering any other potential error candidate was found for a correction value of 25%. If the measured amount of precipitation is corrected by this value, the differences between the observed and simulated snow cover are drastically reduced. The deficit nearly disappears for 1986-1987 and 1987-1988. In these years, differences between simulations using uncorrected precipitation data and observations ranging between 10 cm and 20 cm occur during the secondary accumulation period (March 1987 and February 1988).

In Svalbard Lufthavn snow depth and local accumulation rates are determined by both the precipitation rate and a frequent redistribution of snow. Snow-drifting and redistribution processes, which are not considered in the model, can lead to an underestimation or an exaggeration of the simulated snow depth. This uncertainty partly contributes to the general simulation deficit, but it might also explain the qualitative difference in the simulation results of 1989-1990. The observed increase in the snow depth in the second third of April 1990 cannot be accounted for in the precipitation data. This leads to the assumption that either an error occurred in the data transfer, or that the growth of the snow depth was caused by snow drifts, which is not considered in the snow-cover model.

In comparison with the midlatitudes, the turbulent fluxes in Arctic regions have two unique characteristics. First, mass transports caused by drifting snow reduce the turbulent motion of the atmosphere. *Wamser and Lykossov* [1995] posited that these processes can be considered as effectively an increase in the air density. Second, the turbulent fluxes of warm-air advection are overestimated following a period of stable conditions.This is due to a cold insulation layer remaining over the snow cover, which prevents a mass and energy exchange between the snow cover and the atmosphere [*Morris,* 1989]. Like for Sodankylä, we found systematic deviation when using different parameterizations for the turbulent fluxes. However, the simulated snow depth curves only differ from April to the end of the ablation period by about 2 cm to 5 cm.

Strong winds lead to a modification of the new snow density and a higher rate of metamorphism. Both processes, which are not considered in the model, result in a higher snow density. Within the air, the fine structures of the snow crystals are ground down by the wind, resulting in a more compact accumulation. Strong blowing wind exerts extra pressure on the snow, which accelerates the increase in the snow density and thus the aging process.

The difference in April 1989, when the reduction of the snow depth is simulated at the right time, but by 5 cm too little, might be strongly influenced by the lower boundary of the model. The assumption that the soil temperature is equal to the snow base temperature of the previous time step seems an underestimation of the actual value when using the default value for the thermal conductivity of soil. The snow depth is determined exactly during the ablation period, when the simulated ground heat flux is increased. Sensitivity studies concerning this problem are also shown in part 2 [*Loth and Graf,* this issue] (changes in assumed soil temperature and thermal conductivity of the soil have a compensating effect on simulated snow depth).

4.3. Maritime Station at Midlatitudes

The midlatitudes are characterized by the activity of cyclones and by the influence of high-pressure areas. From October to April the type of precipitation can either be snowfall or rain. Compared with other lowland conditions at midlatitudes, Newfoundland has a high precipitation rate (about 1000 mm/yr) due to frequently occurring maritime air masses.

At Gander (Figure 13), differences between the simulated snow depth and the observations of 8 cm to 35 cm occur at the end of November to mid-December 1986, in December 1989, and in all Januaries. Further difficulties in determining the snow depth accurately which are of the same magnitude occur in the spring periods of all periods (mainly March 1988-1990). The date of snow-cover disappearance in mid-April is only approximately determined. Differences between the observed and simulated end of the ablation period can be up to10 days.

The snow cover at Gander is determined by the entire three-phase physics at the snowpack and the interface fluxes at surface and snow base. Using synoptic data sets only made it necessary to perform a wide range of sensitivity studies to identify the important parameters and main simulation uncertainties. We do not describe all sensitivity studies in this section but focus on the most interesting results.

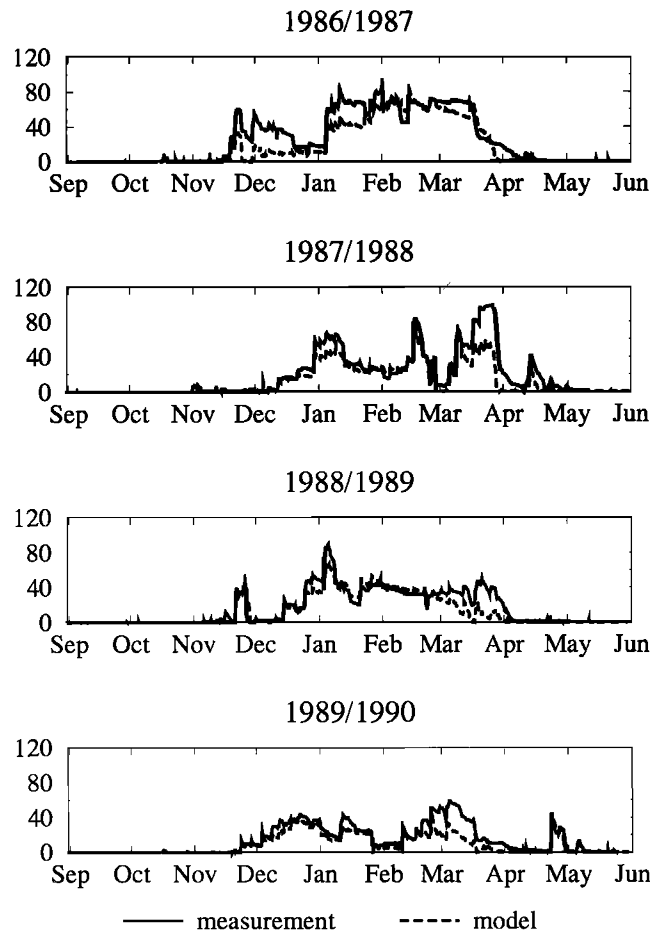
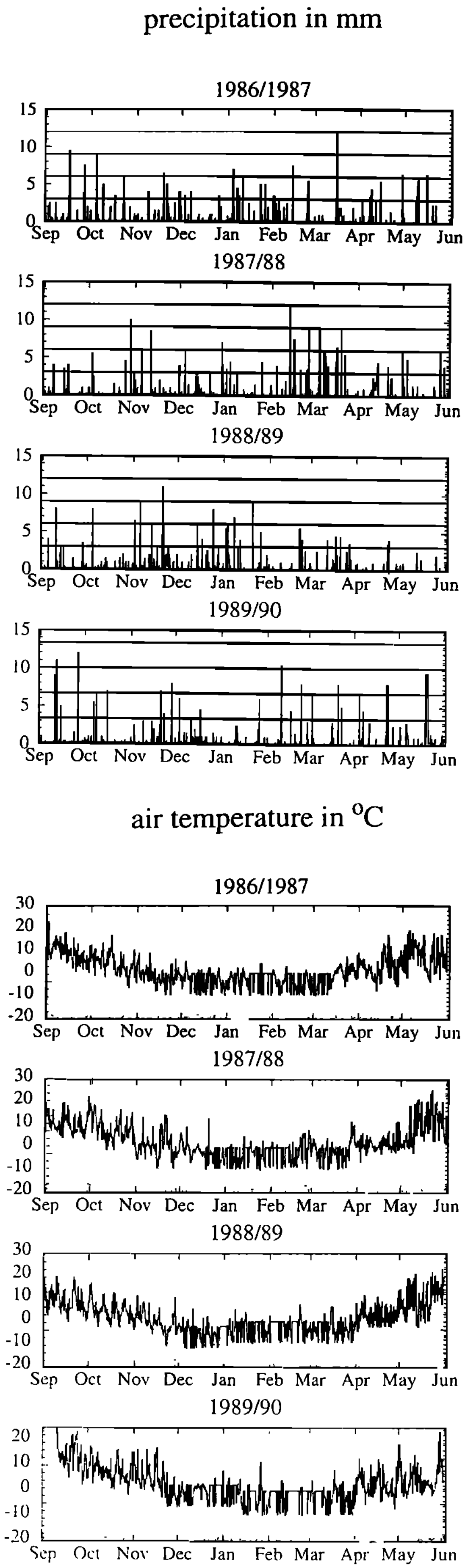


Figure 13. Observed and simulated snow depth (in centimeters) at Gander (Newfoundland) for 1986-1990.

The uncertainty of the snow-cover simulation to the boundary conditions (the incoming short- and long-wave radiation, the turbulent heat scheme, the lower boundary condition, and the snow-rain criterion) is relatively high. Reasonable changes occur in all these parameters, namely, using liquid-water contents in the radiation scheme instead of the climatological approach for the clouds as described in section 2, implementing a different turbulence scheme [*Louis,* 1979, *Anderson,* 1976], assuming different soil temperatures (see below), assuming different correction values for the precipitation, and changing the snow-rain criterion to an air temperature of 0°C or a wet-bulb temperature of 1°C as proposed by *Anderson* [1976]. This changes result in uncertainties of the simulated snow depth with the same order of magnitude and lead to differences in the simulated snow depth of less than 1 cm to about 35 cm to 40 cm for the period 1986-1990.

The assumption of a positive or negative the soil temperature can directly influence the snow mass due to simulated or nonsimulated melting processes at the snow base. We tested different assumptions for the soil temperature; constant values (-1.5°C, -1°C, -0.5°C, 0°C, 0.5°C, 1.5°C, 2°C) and values which depend on the simulated snow base temperature of the precious time step. To this snow base temperature we added +0.5 K, +1 K, and +1.5 K. The simulation results which are shown in this section are performed with a soil temperature of 1.5°C. The differences in the snow depth resulting from the modification of the lower boundary as described above range between 4 cm and 12 cm in autumn and winter and can even be the double of these values in spring.



Under the conditions at Gander the simulation quality strongly depends on the snow-rain criterion. Since the distinction between snowfall and rain determines the accumulation rate of the snow cover, a misinterpretation of the type of precipitation can lead to three effects: a temporal shift in the snow-cover period, errors in the estimated snow mass or depth, and an inaccuracy in the sometimes enormous heat gain of the snow cover due to freezing of rainwater. The differences between the simulated and observed snow depths, which occur in November 1986, January and March 1988, March 1989, and February to April 1990 probably result from this problem (cf. Figure 14, which shows the data of precipitation and air temperature for 1986-1990). Using a different, but also reasonable, snow-rain criterion for single events can reduce the deviations between simulated and observed snow depth to very small values (1 cm to 2 cm). Using the snow-rain criterion proposed by *Anderson* [1976] leads to differences between the two model versions, which are just before the ablation period 16 cm to 27 cm and 40 mm to 70 mm in the snow depth and the snow water equivalent, respectively.

Sensitivity studies have shown that the simulation quality for the snow cover at Gander is further influenced by the following parameters: snow albedo, storage of liquid water, internal heat conduction, rate of snow metamorphism, and turbulent fluxes at the snow surface. In order to have the chance of a successful snow-cover simulation at midlatitudes, all these processes must be resolved and described with a high degree of accuracy (cf. part 2).

Because of the high frequency of snowfall and the snow-aging processes, the (simulated) snow albedo ranges between 0.9 and 0.75 during longer times of accumulation and metamorphism periods. Only the occurrence of liquid water at the snow surface leads to a higher decrease in the snow albedo, which then ranges between 0.70 and 0.60. In the ablation period the snow albedo decreases up to about 0.50. Changes in the dry snow aging coefficient of albedo leads to small differences in the simulated snow depth of 2 cm to 6 cm. If the albedo decrease is set to -0.1/d during the melting period, the snow depth differs by -10 cm to -24 cm between the two model versions and the ablation period is shifted by about 4 days to 7 days.

The differences between the observed and simulated snow depth in November to January and March to April can also be influenced by the simulation of the internal processes. An inaccurate determination of the liquid-water content, the heat conduction, and the snow-surface temperature. Wet snow metamorphism and liquid-water transmission are roughly parameterized *[Jordan,* 1991; *Loth,* 1995], causing uncertainties in the estimated snow cover. The mean differences in the estimations due to a change in the parameterization are between 2 cm and 12 cm in snow depth and 10 mm to 20 mm in water equivalent.

The internal heat conduction is characterized by the combined effect of temperature diffusion, water-vapor diffusion, and absorption of short-wave radiation.These three processes all have to be considered in a model, since they are based on different physical mechanisms. The ground heat flux, which has small climatological values of 2 W/m^2 to 5 W/m^2, depends above all on the state of the soil (frozen/nonfrozen) before and during the snow period and determines the time and intensity of the melting processes at the snow base.

Figure 14. Measured precipitation (in millimeters) and air temperature (in degrees Celsius) at Gander for 1986-1990, shown by the black curves and gray curves, respectively.

4.4. Zugspitze

As was shown for Col de Porte, the multi layered model is able to reproduce deep alpine snow cover in detail. Using the data set from Zugspitze (47.4°N,11.0°E, 2962 m), we aimed to test the model behavior under extreme conditions. At Zugspitze the conditions are extreme concerning the depth of the snow cover and regarding the typical wind speed at this site. Because of the altitude, the climate at Zugspitze is characterized by low air temperatures and intensive snowfall. The daily maximum temperatures in winter fall significantly below 0 °C, they increase to slight positive values during the transition seasons, and range between 10 °C and 15 °C in summer. Snowfall and minimum temperatures below the freezing point can occur throughout the whole year. The mean rate of precipitation for 1951-1980 is 2000 mm/yr. Precipitation from October to mid-May appears only in solid form.Wind speed is higher at Zugspitze in comparison with the midlatitude lowland stations, usually showing mean values between 7 m/s and 12 m/s. The extreme values range between 25 and 30 m/s. Closed snow cover normally exists at Zugspitze from October through July, although in some years it is a year-round phenomenon. Similar to the Arctic stations, the soil temperature is set to the snow base temperature of the previous time step.

Under extreme conditions, such as those at Zugspitze, the model describes the snow cover development relatively well (Figure 15). The general shape of the snow depth curves is estimated for all years 1986-1990. For the period 1986-1987 the snow depth is simulated with a high degree of accuracy until mid-July. A larger difference between the simulated and observed values (120 cm to 150 cm) only occurs in August 1987. During the simulation periods of 1987-1988 and 1988-1989 the snow depth is generally overestimated. From January 1988 through March 1988 and from April 1988 through September 1988 the simulations exceed the observations by between 70 cm and 100 cm. The simulation also exaggerates the observed snow cover in January and February, as well as from mid-June to October 1989, where snow depth is overestimated by the same order of magnitude. As a consequence of these uncertainties, the date of the snow disappearance is not exactly determined for the simulation periods 1987-1988 and 1988-1989. Negative differences between the simulated and the observed values of 100 cm to 140 cm occur from March until mid-June 1990, whereas the period from November 1989 to February 1990 and from July to August 1990 is well described by the model.

Similar to the results for Svalbard Lufthavn, difficulties in the estimation of the accumulated snow result from the activity of the strong winds, which lead to high measuring errors of precipitation, snow drift, and redistribution of snow. Owing to the exposed location of the station, snow drifting and redistribution occur frequently and with high intensity. The differences between the simulated and the observed values in December 1987, December 1988, January 1989, July 1989, and March to mid-June 1990 are, at least partly, caused by these wind-dependent factors. At the end of December 1987 the wind speeds range between 10 m/s and 24 m/s for a period of about 5 days. Assuming that the "missing" snow was blowing away and reducing the simulated snow depth curve by about 90 cm leads to a good agreement of simulations and observations from January to March 1988. We receive the same degree of correspondence if we assume a high turbulence around the measuring device at high wind speeds and a consequent high measuring error for precipitation. At the beginning of December 1988 the observed wind speed was even higher than the year before and had a maximum of 32 m/s. Relatively high wind

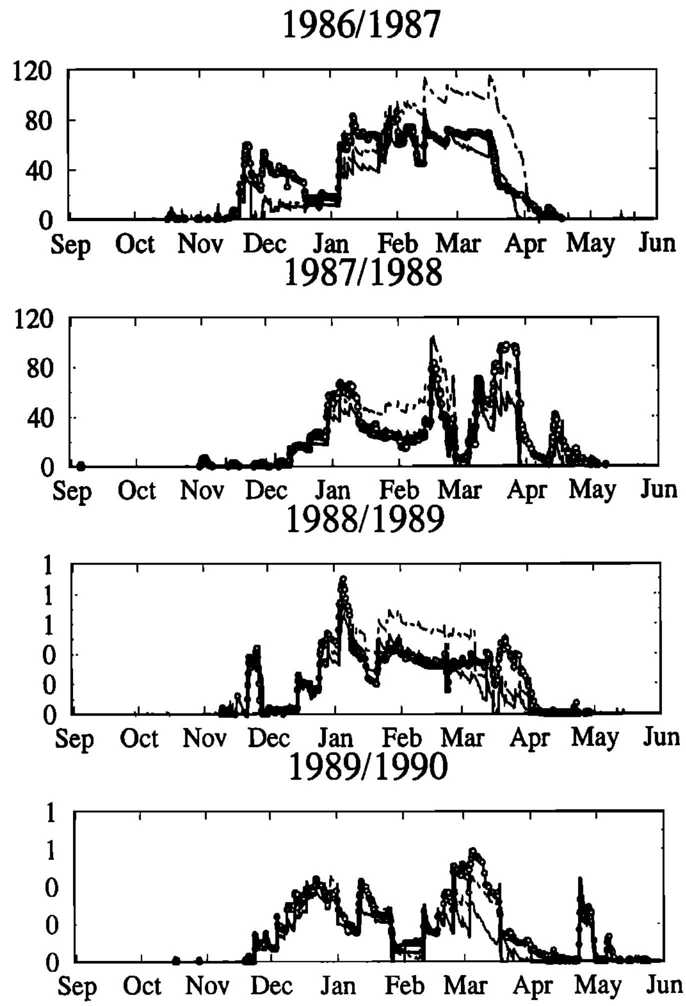


Figure 15. Observed and simulated snow depth (in centimeters) at Zugspitze for 1986-1990.

speeds of 8 m/s to 20 m/s also occur at the beginning of March 1990 and July 1989.

Owing to the midlatitude position, the snow cover at Zugspitze is further determined by the following parameters: snow albedo, evaporation and turbulent fluxes, storage and transmission of liquid water, ground heat flux and internal heat conduction, and processes of metamorphism. Not only the net radiation, but also the turbulent heat fluxes at the snow surface, evaporation, ground heat flux, and intensity of the internal snow processes (liquid-water processes, aging rates, internal heat conduction) are climatically relevant for the snow cover. All these fluxes and processes are important for the estimation of the snow, since they exert a strong impact on the determination of the snow-surface temperature, the mass lost during cold periods and the occurrence of melting processes at the snow base and thus for the calculation of snow mass, energy exchange at the snow surface and the end of the ablation period (cf. part 2).

Because of the high snowfall rate, the (simulated) snow albedo has mainly values between 0.9 and 0.7 during the accumulation and aging period. The snow albedo decreases to 0.6 to 0.4 if melting processes occur at the snow surface. In addition to the influence of liquid water, the snow albedo can drop drastically (by -0.2 to -0.5) when Sahara dust or other pollutants are advected into the alpine area and deposited on the snow surface. Since these processes depend on the regional circulation system and since aerosol concentrations are not measured, changes due to pollution are not considered in the model. The aging coefficient of the albedo implicitly contains this influence for moderate conditions.

The negative differences between the simulated and the observed values which occur from March to mid-June 1990 can partly result from the (unrealistic) melting processes which are modeled in the lowest snow layer. The problem of an overestimation of the melting at the snow base is also detected for the simulations at Col de Porte.

One reason for a general overestimation of snow depth is the fact that the wind-driven metamorphic processes are not considered in the model. Under high wind conditions, the combination of a rounding of new snow crystals in the air, wind compaction of the deposited snow, and an increase in grain size due to the accumulation of drifting snow instead of fresh snow strongly influence the overall properties of the snow cover. The effect of these processes can therefore not be neglected, and indeed a parameterization can be implemented in the model by using increased aging rates for higher wind speeds.

A misinterpretation of the type of precipitation and an underestimation of new snow density in summer can also not be discounted. A high sensitivity to the choice of the snow-rain criterion exists in autumn and summer. During this time, the type of precipitation determines the beginning of the accumulation period or the end of the ablation. The measurements of snow depth, precipitation, and temperature lead to the assumption that the major part of the inaccuracies in August 1987, mid-July to September 1988, and July to October 1989 results from this problem. From mid-July to September, precipitation occurs at positive and negative air temperatures. According to the implemented snow-rain criterion, the model simulates snowfall (more than 50% of the precipitation for air temperatures below 1°C) during these months. This does not seem to be the best parameterization for the summer season. The simulation deficit which stems from the snow-rain criterion is increased due to the fact that the (unrealistic) higher albedo prevents an accurate simulation of the absorption of short-wave radiation and, as follows, the melting intensity is underestimated. The superposition of both effects, and probably further, results in the simulation deficit of +100 cm to +150 cm. The snow-rain criterion may also influence the simulation quality in winter and spring. It might have caused the inaccuracy in March 1990, for example, when the air temperature was near the freezing point for a 10-day period and the precipitation was partly assumed as rain by the model.

5. Summary

A multi layered snow-cover model [*Loth et al.*, 1993], working on basic physical principles, was verified under different conditions. The seasonal cycle as well as single events during the accumulation, aging and melting periods are well described. The model estimates all climatically relevant snow properties (snow depth, water equivalent, albedo, snow-surface temperature) with a high degree of accuracy. The key parameters of the model are the albedo, the retention capacity, the transmission rate of liquid water, the new snow density, and the aging rates. The simulation accuracy also depends on the turbulent fluxes, the internal heat conduction, the ground heat flux, and the choice of the snow layers, which is a purely numerical phenomenon.

At midlatitudes, simulation inaccuracies of deep snow covers mainly result from the rough parameterization of the liquid-water transmission, which leads to an underestimation of the snow mass at the beginning of the ablation period and occasionally to an overestimation of the snow-surface temperature due to a wetter than observed snow surface. A new parameterization of the liquid-water transmission has been introduced into the next generation of the model.

A further improvement of the simulation results can be achieved by taking into account drifting snow, simulating the wind-driven aging processes, correcting more precisely the error of the precipitation measurements, and improving the accuracy of the estimated turbulent fluxes over snow covers.

One version of the multi layered snow-cover model is now formulated as part of the soil scheme SURF and will be used in simulations with the global climate model ECHAM. This version limits the number of snow layers and considers a partial snow coverage of the grid cell for low water equivalents as well as the influence of forest on the snow albedo.

The complexity of a multi layered snow-cover model is necessary for carrying out this type of study, which aims to find out the climatically relevant snow parameters and processes. Since snow is a three-phase system, the physical behavior of which is characterized by a range of non-linear processes, the long-term sensitivity to single processes can only be analyzed by resolving the full set of snow physics. Further, the interdependence of (internal and interface) snow parameters and the exchange fluxes at the snow-cover boundaries does not allow using simplified box or linear models for such an analysis. Whether or not multi layered snow-cover models are necessary to resolve all snow-related feedback mechanisms in a climate model should be analyzed in future intercomparison projects. The answer, of course, depends on the general complexity of the atmospheric model. An adequate proportion in complexity of the different parts of the climate system and the scale of the studies has to been guaranteed. However, if an atmospheric model is used in cold regions, where at least for half a year many grid cells are snow covered, a multi layered snow-cover model is as important as a multi layered soil model, which is used in nearly all state-of-the-art climate models. Looking at the temperature profiles, which are quadratically with the highest gradient in uppermost snow layers, leads to the assumption that the internal temperature of the snow cover has to be resolved in order to accurately determine the surface temperature. On the other hand, climate models are used for long-term studies ranging from 100-year to 1000-year integrations. On these timescales, small inaccuracies can lead to drastic simulation error, in particular, if the described system is involved in a range of feedback mechanisms.

The focus of this paper is on the physically relevant complexity of a snow-cover model for climate studies. Modifications to the snow properties caused by different types of vegetation and the representation of horizontally heterogeneous conditions were excluded from this paper's investigations. The snow albedo, the heat exchanges between snow and ground, as well as the turbulent fluxes at the snow surface do, however, strongly depend on these parameters and has to be considered when the multi layered snow model is coupled to a three-dimensional model. The parameterizations of these influences used in the ECHAM version are described in this paper.

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