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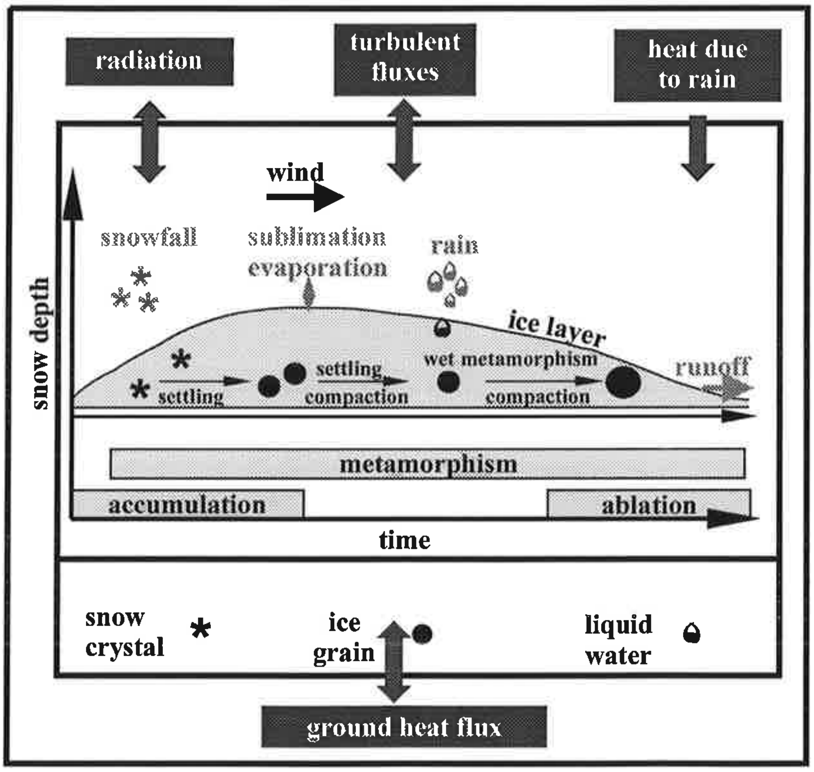

MODELLING THE SNOW COVER FOR CLIMATE STUDIES

PART I: Long-Term Integrations under Different Climatic Conditions Using a Multi-Layered Snow-Cover Model

PART II: The Sensitivity to Internal Snow Parameters and Interface Processes

by

BETTINA LOTH • HANS F. GRAF

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Bettina Loth and Hans F. Graf

Max-Planck-Institut für Meteorologie Hamburg

Bundesstraße 55
D-20146 Hamburg
phone: +49-40-41 173 378/246
fax: +49-40-411 73 298
email: loth@dkrz.de, graf@dkrz.de

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Modelling the Snow Cover in Climate Studies
Part I: Long-Term Integrations under Different Climatic Conditions Using a Multi-Layered Snow-Cover Model

Bettina Loth and Hans F. Graf
Max-Planck-Institut für Meteorologie, Hamburg

Abstract

A physically based snow cover model has been designed and optimized for climate studies. Using this multi-layered model, it is guaranteed to simulate all climatically relevant interactions between snow cover and the atmosphere. Simpler models, despite their obvious speed advantages, can lead to large simulation errors at mid-latitudes (shown in part II). These inaccuracies in the calculated energy fluxes at the snow surface and the snow mass directly influence the simulation quality of the climatic processes within GCM[1]'s.
The multi-layered model, which can be used for stand-alone experiments as well as implemented within atmospheric models, has been tested under different climatic conditions (Arctic, mid-latitudes, alpine regions). The key parameters of the model are

(i) snow albedo
(ii) description of liquid water (storage capacity, transmission rate).

The simulation results further depend on

(iii) the turbulent fluxes at the snow surface
(iv) the parameterization of the heat conduction
(v) density changes due to ageing
(vi) the choice of snow layers.

One version of the model is formulated for the Hamburg climate model ECHAM.

1 Introduction

Since the 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, including climatic studies. From the hydrological point of view, for example, the appearance of snow cover leads to a temporal shift in the runoff from autumn to spring. Unusually deep snow covers, or a shift in the snow melt period towards a period with intensive rainfall can lead to heavy floods. Models are important tools for predicting such situations. Furthermore, snow cover models are useful for interpreting satellite pictures, traditionally problematic due to the similar radiation properties of clouds and snow cover, or for reconstructing historical snow data. 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 (tourism).

[1] General Circulation Models

Considering the snow cover as part of the climate system, we have focused on the interaction between the snow cover and the atmosphere. Due to the drastic alteration of albedo and the modification of the turbulent heat fluxes at the earth surface, snow cover 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; Mercalli and Paludi, 1995; 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 such as measuring uncertainties.

Despite the obvious importance of snow cover for climatic processes and feedback mechanisms, even in sophisticated AGCMs the snow processes are only crudely described. The majority of atmospheric models do not resolve the internal snow cover processes which determine the snow surface properties and thus control the interface fluxes between snow cover and atmosphere. The snow albedo, for example, the parameter for describing the ageing processes of snow, is often parameterized as a time-independent function of temperature. The consequence of this is that the snow is unrealistically renewed when the temperature drops. This can lead to inaccuracies in the albedo ranging between 0.4 to 0.5 in spring, where the albedo values determine the ablation behaviour and the end of the snow cover period. Unfortunately, feedback mechanisms estimated by models show a wide range of responses to any given snow anomaly (Randall et al., 1994). Numerical studies mainly depend upon the complexity of the snow model itself.

This paper describes results performed with a multi-layered snow cover model which has been developed and optimized 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) each representing a special type of snow cover. The model was verified by comparing a set of observed and simulated snow parameters, specifically the snow mass balance (snow mass, runoff) and the snow parameters controlling the interactions between the atmosphere and the snow cover (snow albedo, surface temperature). In the case of Col de Porte, the results were compared with 2 other state-of-the art-models, which are used for snow cover simulation in climate studies (Brun et al., 1992; Douville et al., 1995).

2 Description of the Model

The snow-cover model (Loth et al.,1993), the numerical and physical structure of which is schematically given in Fig. 1 and Fig. 2 respectively, 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-vapour diffusion, transmission and storage of liquid water, heat conduction, melting, freezing, and ageing). The snow mass balance is determined by the intensity of snowfall, 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 rain water. 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 ansatz 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-vapour pressure is set to the saturation value over ice (for dry snow) or water (for snow with liquid water), the water-vapour concentration is a diagnostic variable and depends on temperature only. The prognostic variables of the model are the depth of snow layers, the snow albedo, the snow temperature, the density of dry snow and the liquid-water content. The number of layers depends on the vertical structure of the snow cover. In general it ranges from 2 to 5 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. Due to the ageing processes, snow can be transformed into firn snow and ultimately into layers of compact ice. Aged snow, firn snow and layers of compact ice differ in both snow density and the shape and the size of the snow gains. The three different types of snow, which are considered in the model's physics, are defined as follows;

(i) cold snow without liquid water (penetrating water immediately freezes)
(ii) wet snow with a fixed temperature of 0^{o} C
(iii) 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 and from 3% to 5% for aged 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 density changes (Anderson, 1976). In wet snow conditions the ageing processes of the snow density are accelerated and to account for this the ageing 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 sun the snow albedo is parameterized by a linear function of time. The ageing coefficient is set to -0.006/d for dry snow and -0.071/d for melting snow covers with a depth below 25 cm respectively (Gray and Landine, 1987). For deep snow covers an exponential approach is used (Verseghy, 1991)

$$\alpha(t) = 0.5 + (\alpha(t-1) - 0.5) \cdot \exp(-(0.24 \cdot \Delta\tau)) \tag{1}$$

The modification of the snow albedo by clouds and/or the sun height is parameterized following Siemer (1988):

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

where N_{cl} is the amount of clouds, 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 snow albedo is 0.85.

At the end of each time step, neighbouring snow layers are merged if the depth of one layer is less than 3 mm. In order to avoid a high number of layers, neighbouring layers are also combined according to the following criteria;

(i) if their temperature difference is less than 3 K
(ii) if they consist of the same snow type
(iii) if the density difference is less than 150 kg/m^3.

N.B. The snow layers at the interfaces with the atmosphere and with the ground are excluded from this procedure.

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 Tab. 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 Tab. 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:

$$Q_G = Q_{G0} \cdot \left(1 - f_{cl} \cdot N_{cl} \right) \tag{3}$$

Q_{G0} and N_{cl} are the incoming short-wave radiation at the snow surface for clear-sky conditions and the total amount of clouds 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 downwards 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-vapour pressure and the air temperature (Idso, 1981). The radiation emitted by clouds depends on several factors;

(i) the temperature at the cloud base.
(ii) the emissivity of the clouds.
(iii) the transmissivity of the atmosphere in the water-vapour window 8 μm to 14 μm.
(iv) the amount of clouds (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).

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). 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/(m·K) for dry moor and 3.35 W/(m·K) 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/(m·K), 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/(m·K).

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 the falling precipitation 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.

3 Description of the Data Basis

In order to verify the accuracy of any snow-cover model which is to be used as part of a climate study, simulated results should be compared with observations of both the parameters describing snow mass and duration (water equivalent, snow depth, runoff) and the parameters controlling the energy fluxes at the snow surface (snow albedo, snow surface temperature).

In this instance we used a comprehensive data set taken at **Col de Porte** (45 °N, 6 °E, 1320 m) in the French Alps by the Centre d'Etude de la Neige in Grenoble (Brun et al., 1992). 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. The simulation period for Col de Porte covers the period 17 December 1988 to 8 May 1989. 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. The differences between the measurements were caused by the following;

(i) compaction of the grass vegetation
(ii) locally different accumulation rates due to a horizontal redistribution of already aged deposited snow
(iii) the fact that the ultrasonic sensor may have been more exposed to the sun than the measurements taken with the snow pit (Brun et al., 1992).

In addition to the input data necessary for model runs (Tab. 2), the incoming short and long wave radiation was measured. Since the amount of snowfall and rain is determined separately, assessing a snow-rain criterion becomes unnecessary.

The accuracy of the snow cover model was further tested under conditions different to those of Col de Porte. In order to do this we analysed data from four synoptic stations, each representing special types of snow or climatic conditions, as briefly outlined below.

Svalbard Lufthavn (Spitsbergen): Winter conditions determined by polar night and low temperatures. Frequently high winds cause snow drifting and wind compaction of the deposited snow. Despite its relatively northern geographical position, the air pollution is similar to that of the mid-latitudes.

Sodankylä (Finland): As at Svalbard Lufthavn, winter conditions determined by polar night and low temperatures. Unlike Svalbard Lufthavn, the air is clean, horizontal snow redistribution is of minor importance, as too are the effects of wind compaction.

Gander (Newfoundland): A typical lowland station at the mid-latitudes. Since it is situated near the sea, maritime conditions lead to a relatively deep snow cover compared with other mid-latitude locations.

Zugspitze (German Alps): Extreme conditions due to the station's high altitude. Compared to Col de Porte, the mean wind speed is significantly higher, the mean temperatures are lower and the snow cover lasts longer.

The geographical position of the stations is shown in Fig. 3.

The data sets analysed for the synoptic stations cover the period 1 September 1986 to 1 June 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 clouds) are taken at 3-hour-intervals. Precipitation is recorded every 6 hours. The snow depth is daily measured.

4 Results

4.1 Model Consistence (Col de Porte)

The model simulates the snow cover properties fairly accurately (Fig. 4 to Fig. 7). As is demanded for climate studies, the snow mass and the parameters describing the energy exchange with the atmosphere are simulated to a very high standard. The comparison of observed and simulated snow depth shows that the duration of the snow cover is well described. The simulated water equivalent, which expresses the accumulated snow mass in water column length, documents a relatively precise estimation of the snow mass. Both snow parameters controlling the energy exchange between the atmosphere and the snow cover (snow albedo and snow surface temperature) are reproduced in detail. These results lead to the assumption that the model is physically consistent.

Throughout the whole period, the observed and simulated values of snow depth correspond very well (Fig. 4a). From January to Mid-May the simulated curve is nearly identical to the measured curve of the ultrasonic sensor. Differences between these curves occur from the beginning of March to Mid-April and from the end of April to Mid-May, where the simulated values are higher in comparison with those taken by the ultrasonic sensor. These differences, ranging between 10 cm and 20 cm, are caused by both measurement and simulation uncertainties. The main measuring error is the compaction of the vegetation below the snow cover. This compaction leads to a negative bias and a value which is systematically too small. The main simulation error is an inaccuracy in the liquid water processes which results in an accelerated runoff (see below). During the whole period the modelled results are between the two observed values. The values taken by the snow pit are mainly above both the ultrasonic sensor measurements and the simulated values. The simulations exceed the pit measurements only at the beginning of March, when the simulated curve exceeds both measurements by about 20 cm. This overestimation of snow depth results from an inaccurate determination of new snow depth and might be caused by the use of inaccurate precipitation data, an error in the simulated new snow density and the occurrence of a horizontal redistribution of snow. In order to correct the systematic measuring error, the precipitation data were increased by 15% (Brun, 1994), which is, of course, only an approximation. In reality the measuring error depends on the type of precipitation and the wind speed. In the case of low wind speeds, the measuring error can be smaller, which in turn leads to an overestimation of the new snow depth. The dependence of the new snow density on the wind speed and the possibility of horizontal snow redistribution are not considered in the model. The absence of both parameters partly contributes to the uncertainties.

As with the snow depth, the simulated water equivalent (Fig. 4b) is also seen to be systematically smaller than observations, when compared with the actual pit measurements. The differences between the simulated and observed values are -10 mm to -20 mm from December to mid April. They increase to about -50 mm during the ablation period. An analysis of the runoff data (Fig. 5) shows that these inaccuracies (like 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. Put simply, the snow cover loses liquid water too quickly. This error can be caused by two different factors;

(i) by uncertainties in the storage capacity for liquid water
(ii) by an erroneous simulation of the heat conduction at the snow base, which leads to errors in the rate of phase transformation.

An improvement in the simulation occurs if the storage capacity for liquid water is assessed independently for densities greater than 200 kg/m^3. We suppose a capacity of 10% for snow densities greater than 200 kg/m^3, 20% for snow densities above 300 kg/m^3 and 30% for densities above 400 kg/m^3.

The snow-surface temperature (Fig. 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 II.

The snow albedo, which controls short-wave absorption and strongly influences the whole snow cover development, corresponds very well with observations (Fig. 7). The changes in albedo due to both dry and wet snow metamorphism are also well simulated. The assumption of linear ageing 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. This could be attained by using an exponential function for the albedo ageing of cold snow. 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 ageing processes of the following periods (settling, compaction etc.). Since the albedo is already well simulated by the model, this further improvement is not implemented.

4.2 Arctic Conditions

The climatic conditions at the Arctic stations of Svalbard Lufthavn and Sodankylä are typical for land areas north of 60^{o} N. Due to the low temperatures the whole winter precipitation falls as snow.

Despite a small deficit, the snow depth is accurately simulated for Arctic conditions. The shape of the snow cover curve is reproduced in detail for 1986-90. 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, the ageing processes are limited to the occurrence of dry metamorphism, which can be well simulated, since the processes are reduced to phase changes of vapour and ice instead of the wet snow processes including processes and transformations of all three phases. The third water phase (liquid water) only influences the snow cover in spring since the whole precipitation in winter falls as snow and the occurrence of stronger melting processes is limited to the period Mid-April to Mid-May. The end of the ablation period is mainly determined by the snow mass and the albedo. Due 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.

Fig. 8 shows the simulation results for Sodankylä. The above mentioned differences between simulations and observations increase with the duration of the snow cover. They reach their maximum between February and the end of the ablation period (-10 cm to -20 cm). The main cause for this inaccuracy seems to be a measuring error in the snowfall mass. In order to demonstrate the influence of this error on the simulation of the snow cover under Arctic conditions, results are shown assuming a correction value for precipitation of 20% (Fig. 9). As can be seen, the maximum differences between simulations and observations are reduced to small values. For 1987/88 and 1988/89 these differences range between 2 cm and 5 cm.

For 1986/87 the correction of the precipitation data also leads to an excellent correlation between estimations and measurements. The simulated reduction in the snow depth at the end of March, however, was not observed. This leads to the possibility that further sources of error exist. In this particular case, a misinterpretation of the type of precipitation occurs on 30 March 1986, when snow fell at an air temperature of 1 °C, a temperature at which the model assumes a mixture of about 60% snow and 40% rain. Additionally, an overestimation in the values of both the new snow density and the evaporation can partly cause these differences.

Using corrected data for the precipitation measurements also decreases the deficit for 1989/90. But the precipitation measuring error only explains half the story. A further candidate for the deviations from the observations is an overestimation of the evaporation (the only mass lost under Arctic conditions in winter). This can be explained by assuming an overestimation of the water vapour flux at the snow surface which can occur under atmospheric conditions of extreme stability.

In general, the observed snow cover at the Svalbard Lufthavn station (Fig. 10) 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 very thin snow cover (snow depth below 25 cm) the model shows encouraging simulation results. The systematic simulation deficit of snow depth is mainly caused by three factors;

(i) the use of uncorrected data of measured precipitation
(ii) the neglect of horizontal redistribution of snow
(iii) an overestimation of the turbulent fluxes under Arctic conditions due to snow drift extreme atmospheric stability.

Snowfall measurements have an error of more than 50% under conditions of high wind speeds such as exist in Svalbard Lufthavn. If the measured amount of precipitation is corrected by 25% the differences between the observed and simulated snow cover are drastically reduced. The deficit nearly disappears for 1986/87 and 1987/88. In these years differences between simulations and observations of between 10 cm and 20 cm occur during the secondary accumulation period (March 87 and February 88) due to the use of uncorrected precipitation data.

In Svalbard Lufthavn snow depth and local accumulation rates are determined by both the precipitation rate and strong snow-drifting processes. The redistribution of accumulated snow, which frequently occurs under high wind speed conditions, partly contributes to the general simulation deficit and explains the qualitative difference in the simulation results of 1989/90. In contrast to 1986-89, 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. 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 can not be simulated with a 1 D snow-cover model.

In comparison with the mid-latitudes, the turbulent fluxes in Arctic regions have two unique characteristics. Firstly, mass transports caused by drifting snow reduce the turbulent motion of the atmosphere. Warmser and Lykossov (1995) posited that these processes can be considered as effectively an increase in the air density. Secondly, 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).

In April 1989 a further difference occurs when the reduction of the snow depth is simulated at the right time, but by 5 cm too little. Since observed data are not available for Svalbard Lufthavn in 1986-90, the thermal conductivity of the soil is set to the default value. This seems an underestimation of the actual value. The snow depth is determined exactly during the ablation period, when the thermal conductivity of the soil is changed from 0.3 W/(m·K) to 0.6 W/(m·K). Sensitivity studies concerning this problem are also shown in part II.

The high wind speed causes still further metamorphic processes which are not considered by the model. Within the air, the fine structures of the snow crystals are ground down by the wind, resulting in a more compact accumulation. A further type of snow metamorphism, wind compaction, increases the snow density on the ground by exerting extra pressure on the snow.

4.3 Lowland station at Mid-Latitudes

The mid-latitudes 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 mid-latitudes, Newfoundland has a relatively high precipitation rate due to frequently occurring maritime air masses.

Under these conditions the simulation quality is mainly determined by the snow-rain-criterion (Loth et al., 1993). 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 rain water.

For Gander (Fig. 11), differences which resulted from this problem occurred in November 1986, January and March 1988, March 1989 and in the months February to April 1990. Difficulties in determining the type of precipitation occur in the spring periods of all simulations. The date of snow-cover disappearance is therefore only approximately determined. Differences between the observed and simulated end of the ablation period can be up to10 days.

The snow-cover properties are further strongly influenced by the following parameters;

(i) the value of the snow albedo,
(ii) the storage of liquid water,
(iii) the internal heat conduction,
(iv) the rate of snow metamorphism
(v) the turbulent fluxes at the snow surface.

In order to guarantee a successful simulation of the relatively thin snow cover at mid-latitudes, all these processes must be described with a high degree of accuracy (cf. part II).

The snow cover reacts very sensitively to the value of the snow albedo. The snow albedo, which depends on both the ageing processes of snow and the pollution rate at the snow surface, is one of the main parameters determining the snow mass and the duration of the snow cover. Since the incoming short-wave radiation represents the largest energy source in all seasons, small changes in the albedo can lead to drastic changes in the amount of energy absorbed.

The occurrence of liquid water in the snow leads to a rapid decrease in the snow albedo, which remains at relatively low values until new snowfall. When liquid water stored within the snow starts to freeze at temperatures below 0^{o} C the release of latent heat prevents an effective cooling of the snow cover. Thus, an accurate determination of the liquid-water content influences the estimation of the snow mass, the snow albedo and the snow-surface temperature. Furthermore, the liquid water within the snow cover leads to an intensification of the ageing processes. Wet snow metamorphism and liquid-water transmission are roughly parameterized (Jordan, 1991; Loth, 1995), causing uncertainties in the estimated snow cover.

The internal heat conduction is characterized by the combined effect of temperature diffusion, water-vapour diffusion and absorption of short-wave radiation.These three processes have all to be considered in a model, since they are based on different physical mechanisms. Despite the small climatological value of 2 W/m^2 to 5 W/m^2, the ground heat flux might strongly influence the snow cover since it determines the time and intensity of the melting processes at the snow base. In the mid-latitudes the amount of ground heat flux depends above all on the state of the underlying soil (frozen / non frozen) before the accumulation period.

4.4 Zugspitze

Due to altitude, the climate at Zugspitze (47.4° N,11.0° E, 2962 m) 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 freezing point can occur throughout the whole year. The mean rate of precipitation for 1951/80 is 2000 mm/a. Precipitation from October to mid May appears only in the solid form.Wind speed is higher at Zugspitze in comparison with the mid-latitude lowland stations, usually showing mean values between 7 m/s and 12 m/s. The extreme values range between 25 m/s 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.

Under extreme conditions, such as those which occur at Zugspitze, the model describes the snow cover relatively well (Fig. 12). The large differences which occur in August 1987 between the simulated and observed snow depth mainly result from a horizontal redistribution of the accumulated snow. Due to the exposed location of the station, snow drifting occurs frequently and with high intensity. In the summer of 1987 this process even caused an inaccuracy of between 1 m and 1.5 m.

During the simulation period of 1987/88 the snow depth is overestimated from January through March and from April through September by between 70 cm and 1 m. The simulation also exaggerates the observed snow cover in 1988/89. In January and February, as well as from mid June to October, 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 1987/88 and 1988/89.

One reason for this 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 influences 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 ageing rates for higher wind speeds.

The simulated uncertainties can be further caused by the following parameters;

(i) snow albedo
(ii) ground heat flux
(iii) the turbulent fluxes at the snow surface
(iv) intensity of the internal snow processes in particular the liquid-water processes, the ageing rates and the internal heat conduction.

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 ageing coefficient of the albedo implicitly contains this influence for moderate conditions.

The ground heat flux, the turbulent fluxes at the snow surface (including the evaporation) and the intensity of the internal snow processes (liquid-water processes, ageing rates, internal heat conduction) determine the snow-surface temperature, the mass lost during cold periods due to evaporation and the occurrence of melting processes at the snow base. Thus they are important parameters for estimating the snow mass, the energy exchange at the snow surface and the end of the ablation period (cf. part II)

A misinterpretation of the type of precipitation and an underestimation of new snow density in summer can also not be discounted. A sensitivity to the choice of the snow-rain-criterion, which only exists at Zugspitze in the transitional seasons, can nevertheless strongly influence the snow cover. In autumn the type of precipitation determines the beginning of the accumulation period. Shortly before the end of the ablation period, misinterpretation of the aggregational state of precipitation can lead to a temporal shift in the end of the annual snow-cover cycle compared to observations. Misinterpretations in spring and early summer are visible as an integrated effect in the ablation period.

Negative differences between the simulated and the observed values also occur from March until mid June 1990 when unrealistic melting processes are modelled 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.

5 Comparison with other state-of-the-art models

The results for Col de Porte can be compared with simulations performed with the multi-layered model CROCUS (Brun et al., 1992) and the snow parameterization of the soil scheme ISBA1 (Douville et al., 1995). CROCUS is used operationally in avalanche forecasting and concentrates on a precise simulation of the internal snow processes. Unlike the MPIfM model, CROCUS explicitly calculates the process of liquid-water transmission. The ageing processes are estimated from the microphysical point of view. The snow albedo is parameterized as a function of the grain size and the structure of the snow crystals. The soil scheme ISBA1 is implemented in the French weather service's (METEO-FRANCE) climate model, ARPEGE. It uses a relatively simple snow parameterization. The snow cover is considered as one layer. The processes which depend on the vertical structure of the snow cover (heat conduction, internal mass redistribution, ageing processes) are parameterized. Liquid water and processes connected with phase transformation solid / liquid are neglected.

The snow depths estimated by the different models all correspond well. All three models calculate values between the curves of the measurements of the snow pit and the ultrasonic sensor.

The snow albedo is only shown in Douville et al. (1995). The differences in albedo between ISBA1 and the MPIfM snow-cover models are negligible. The snow-surface temperature shows a cold bias in both CROCUS and ISBA1. The magnitude of this systematic error is -2 K to -5 K in the daily means and is inversely proportional to temperature. The MPIfM model reproduces low snow-surface temperatures with a high degree of accuracy. Differences from the observations only occur for the MPIfM snow-cover model in spring (Fig. 6) when the snow surface is too wet. Results of the water equivalent are given neither in Brun et al. (1992) nor in Douville et al. (1995).

Numerical experiments (Loth and Graf, 1996) show that all climatically relevant snow processes and parameters have to be well simulated in order to guarantee a good simulation of the interface fluxes between the snow cover, the atmosphere and the underlying soil.

6 Summary

A multi-layered snow cover model (Loth et al., 1993) working on basic physical principles, was verified under different conditions. This 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 ageing 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.

The snow cover is well reproduced by the model. The seasonal cycle as well as single events during the accumulation, ageing and melting periods are well described. In particular, snow depth and snow albedo are successfully simulated for both cold and wet snow. Simulation inaccuracies, which are generally of less importance, 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 also be achieved by introducing a snow albedo modification caused by pollutants and by simulating the wind-driven ageing processes.

The model performs well in simulating the snow cover under Arctic conditions. The snow depth is slightly underestimated. Differences between simulations and observations for these land areas are largely due to

(i) drifting snow,
(ii) precipitation measuring error and
(iii) uncertainties in the estimation of the turbulent fluxes at the snow surface.

The turbulent fluxes are overestimated under snow drift conditions and warm-air advection following a cold period. When snow drifting occurs in the Arctic regions, the suspended snow increases the air density and thus enhances the atmospheric stability. In cases of warm-air advection following a period of very stable atmospheric conditions, a trace of cold air remains directly over the snow surface and prevents an effective energy exchange.

Thin snow cover at mid-latitudes is well simulated by the multi-layered model provided that

(i) the precipitation measuring error is corrected,
(ii) the type of precipitation is known and
(iii) the turbulent fluxes at the snow surface are well described.

The end of the ablation period is mainly determined by

(iv) the changes in albedo and
(v) the liquid-water storage.

The simulated snow depth at the Gander station is underestimated when using uncorrected precipitation rates or applying a criterion for the distinction between snowfall and rain. The maximum deviations between observations and simulations then range between 10 cm and 40 cm. A misinterpretation of the type of precipitation in spring can lead to a shift in the end of the snow-cover period of between 4 days to 8 days.

The internal temperature diffusion and the rate of metamorphism influence the time of the occurrence of melting processes in the snow cover. The ground heat flux can cause melting processes at the bottom of the snow cover and thus runoff. If this is not estimated, a systematic error occurs in the snow depth, which affects the simulations up to the end of the snow period.

Deep alpine snow cover is reproduced in detail. Difficulties occur when high wind speeds lead to snow drift and wind compaction. During the ageing period the properties of the snow cover are determined by

(i) the snow albedo,
(ii) the turbulent fluxes,
(iii) the storage and transmission of liquid water,
(iv) the internal heat conduction and
(v) the processes of metamorphism.

Due to the relatively long snow period, evaporation leads directly to an effective ablation of the snow cover. The other parameters influence the time of the occurrence of melting processes, the energy balance and the snow-surface temperature. Despite the relative low temperatures at the snow surface, melting processes can appear at the snow base throughout the whole snow-cover period. A realistic simulation of these processes is very important for a correct estimation of the snow cover. The climatically relevant snow parameters of the ablation period are the snow albedo, the storage capacity for liquid water and the rates of settling and compaction. A misinterpretation of the type of precipitation in spring or early summer might lead to a shift in the end of the ablation period.

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. The next step in the model development is to implement a parameterization for these changes.

One version of the multi-layered snow-cover model is now formulated as part of the soil scheme SURF and will be implemented in the AGCM ECHAM4. This version uses a maximum of five snow layers and a partial snow coverage of the grid cell for low water equivalents.

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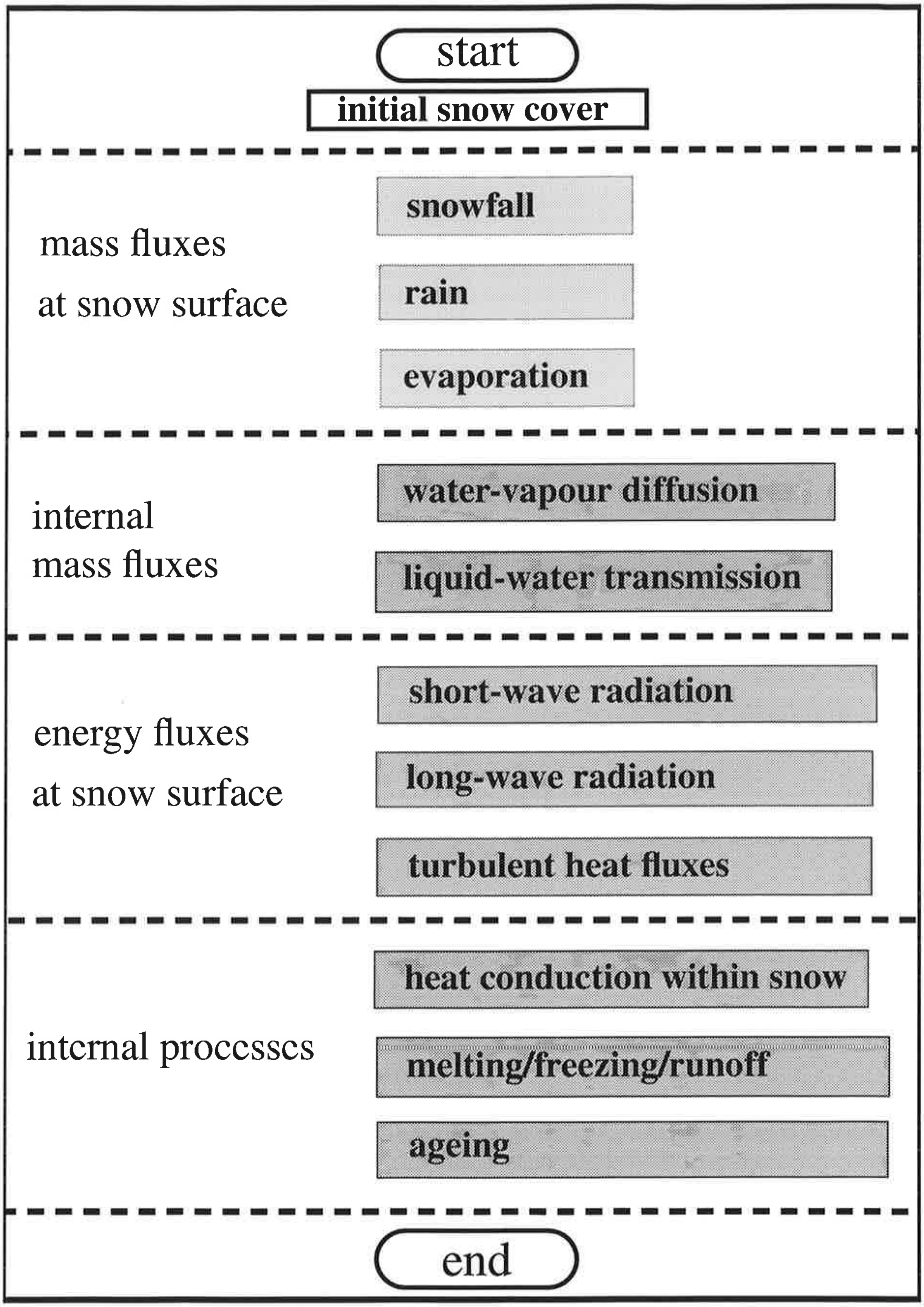

Fig. 1: General structure of the multi-layered snow cover model (Loth et al., 1993).

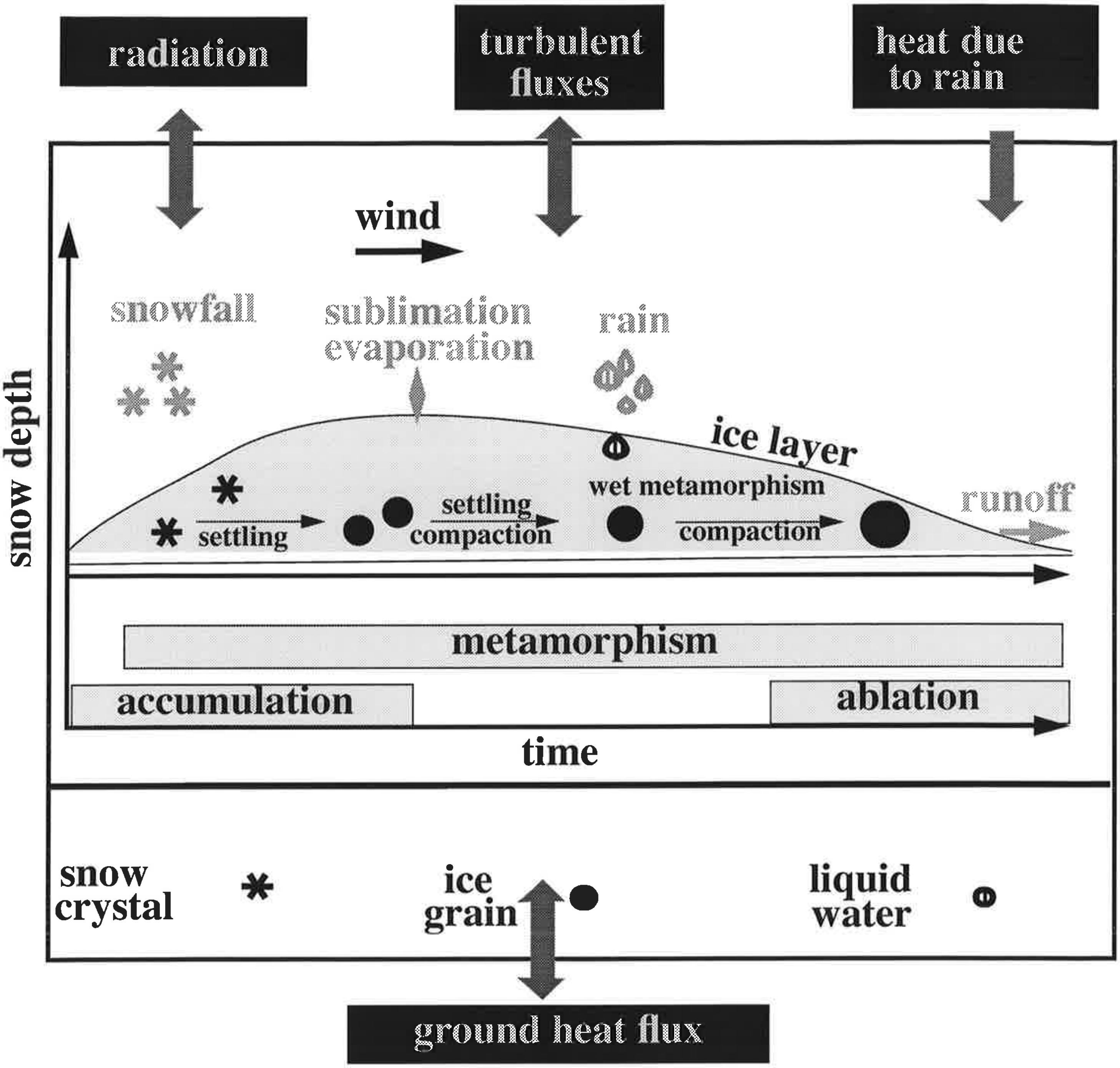

Fig. 2: Schematic representation of the internal snow processes and the exchange fluxes between snow cover, atmosphere and ground which have to be considered in climate studies.

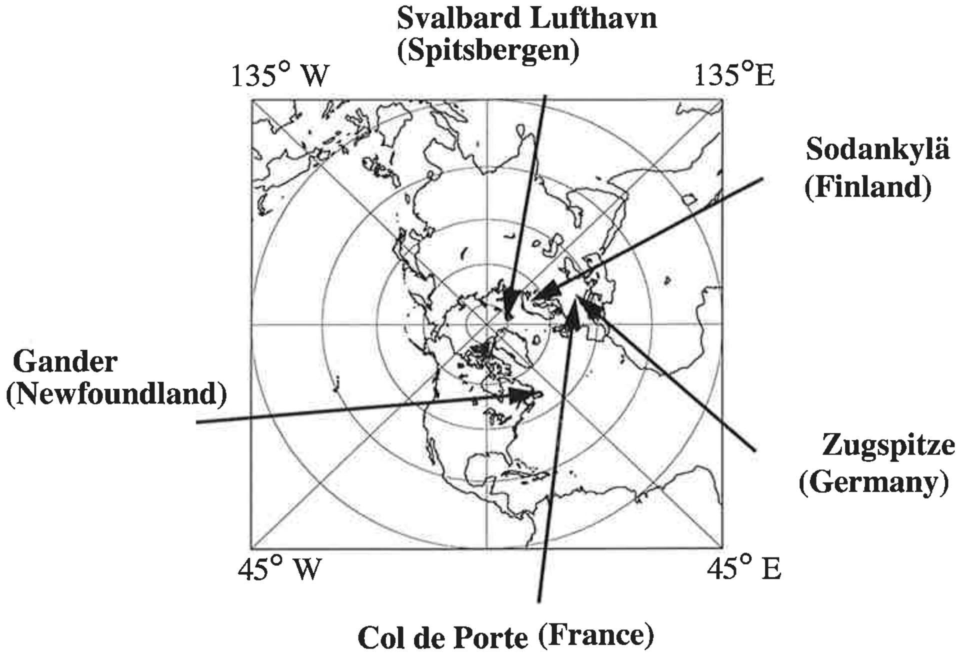

Fig. 3: Measuring stations used for the verification of the multi-layer snow-cover model.

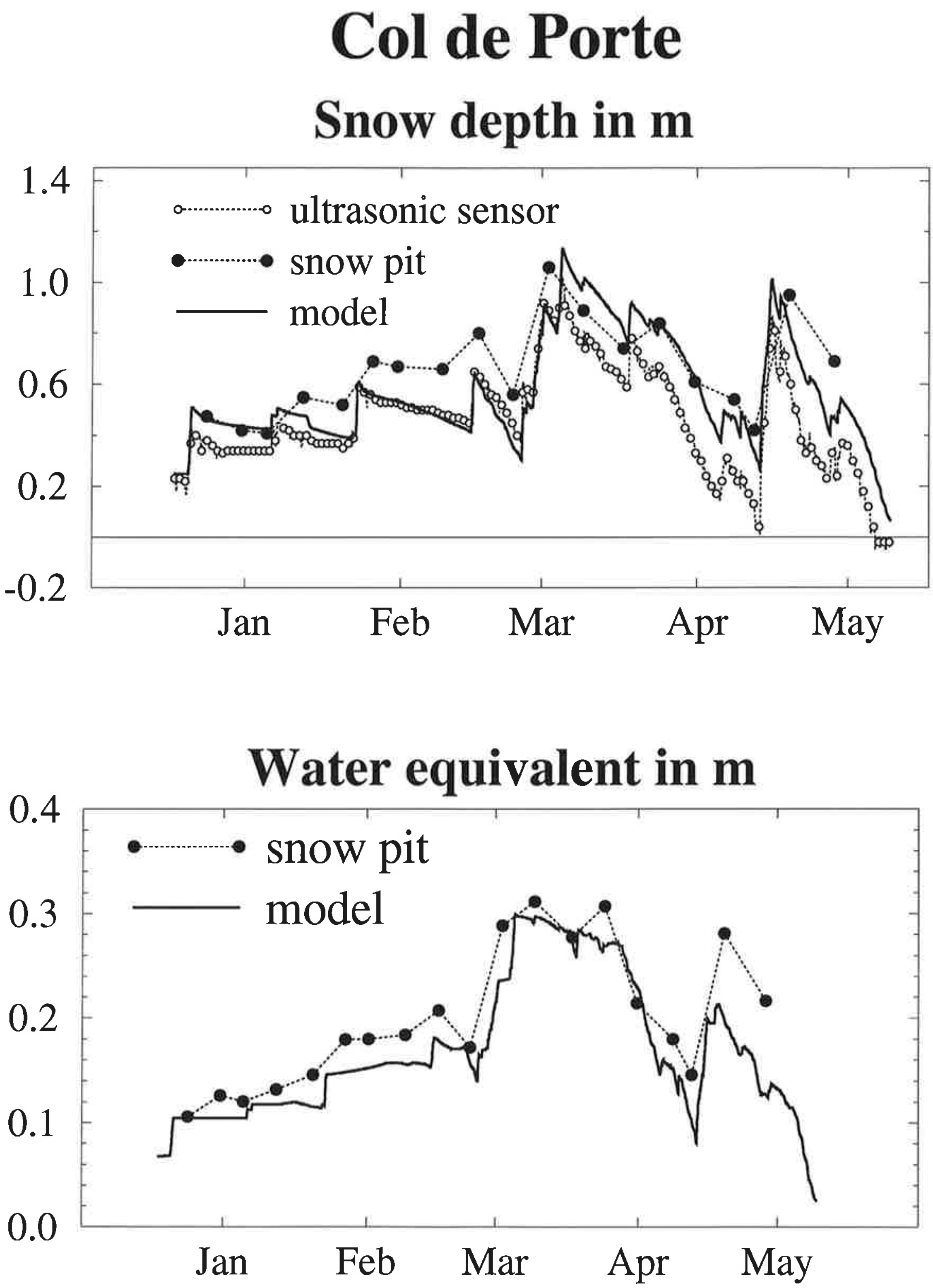

Fig. 4: Measured and simulated values of snow depth and water equivalent at Col de Porte for the time period 17 December 1988 to 8 May 1989.

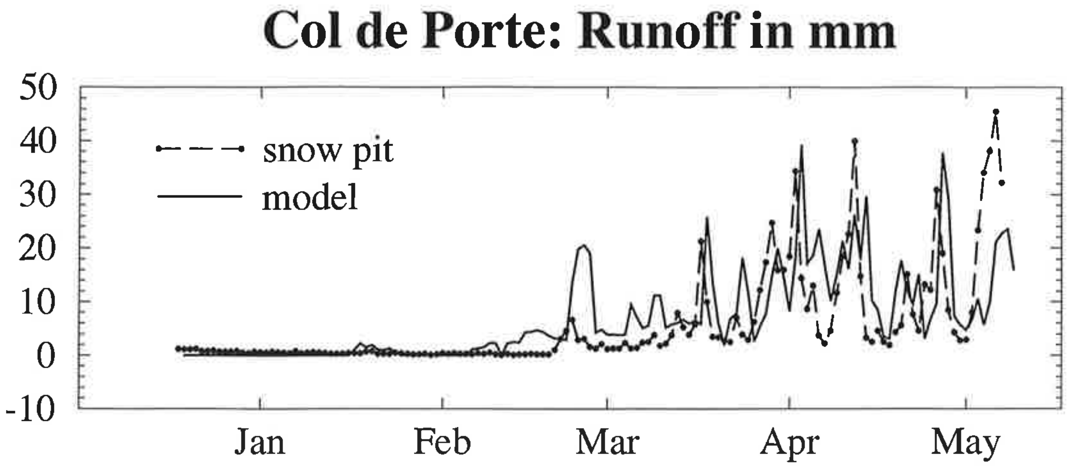

Fig. 5: Measured and simulated runoff (daily sums) at Col de Porte for 17 December 1988 to 8 May 1989.

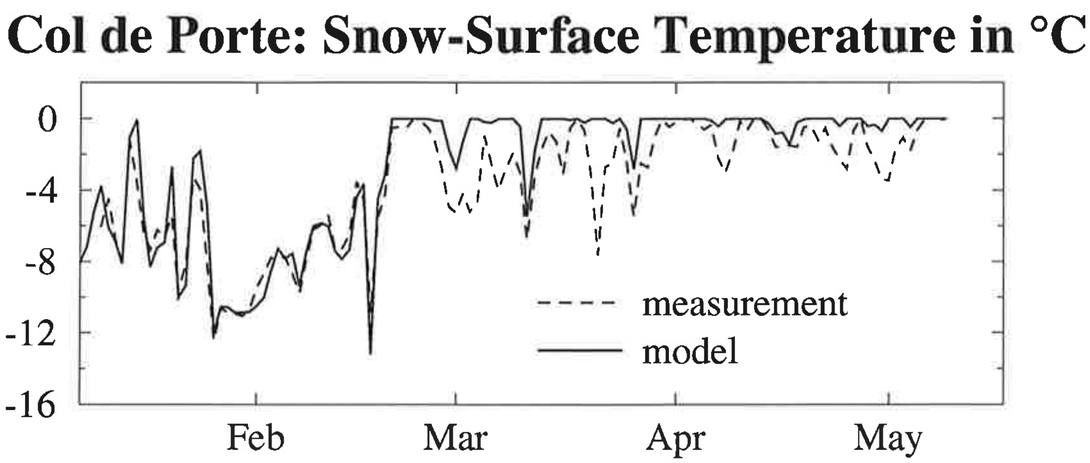

Fig. 6: Daily mean values of measured and simulated snow-surface temperature at Col de Porte for 8 January 1989 to 8 May 1989.

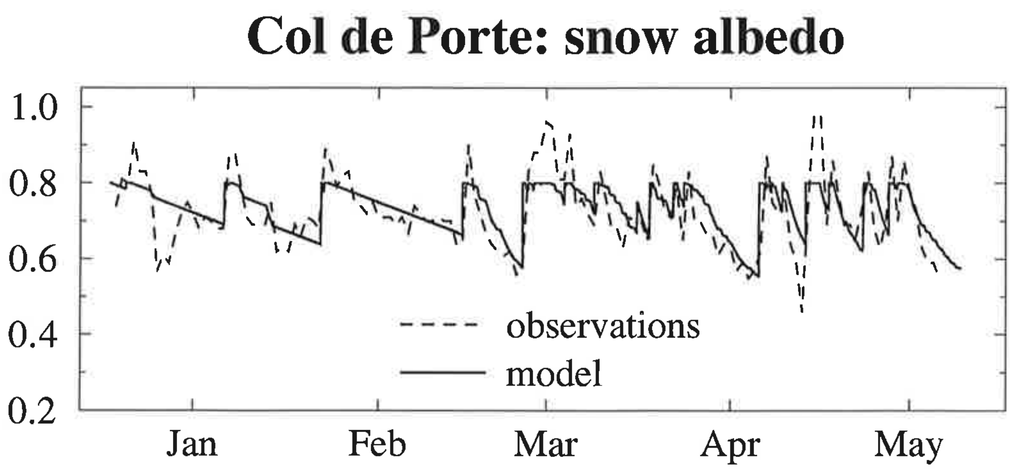

Fig. 7: Observed and simulated values of snow albedo at Col de Porte for the time period 17 December 1988 to 8 May 1989.

Sodankylä: snow depth in m

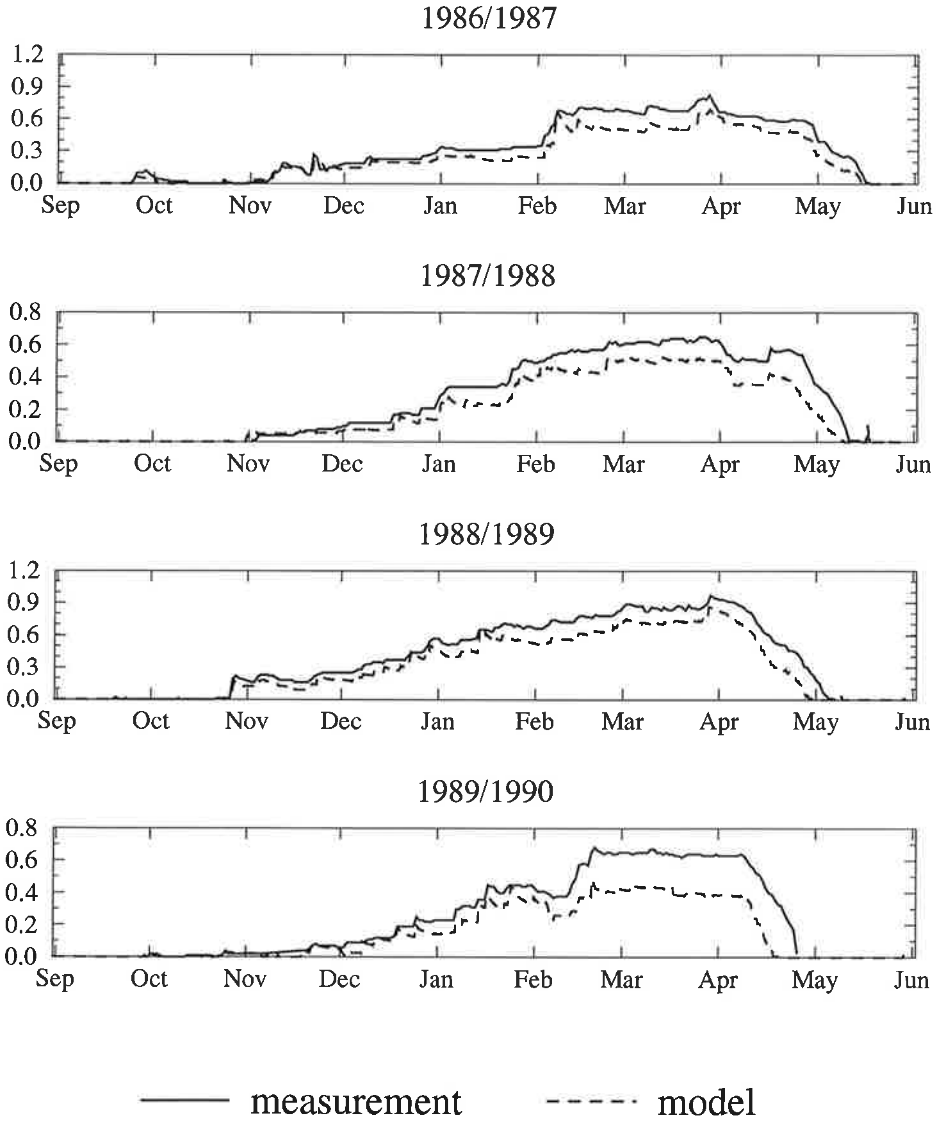

Fig. 8: Observed and simulated snow depth at the Finish station Sodankylä for 1986-90.

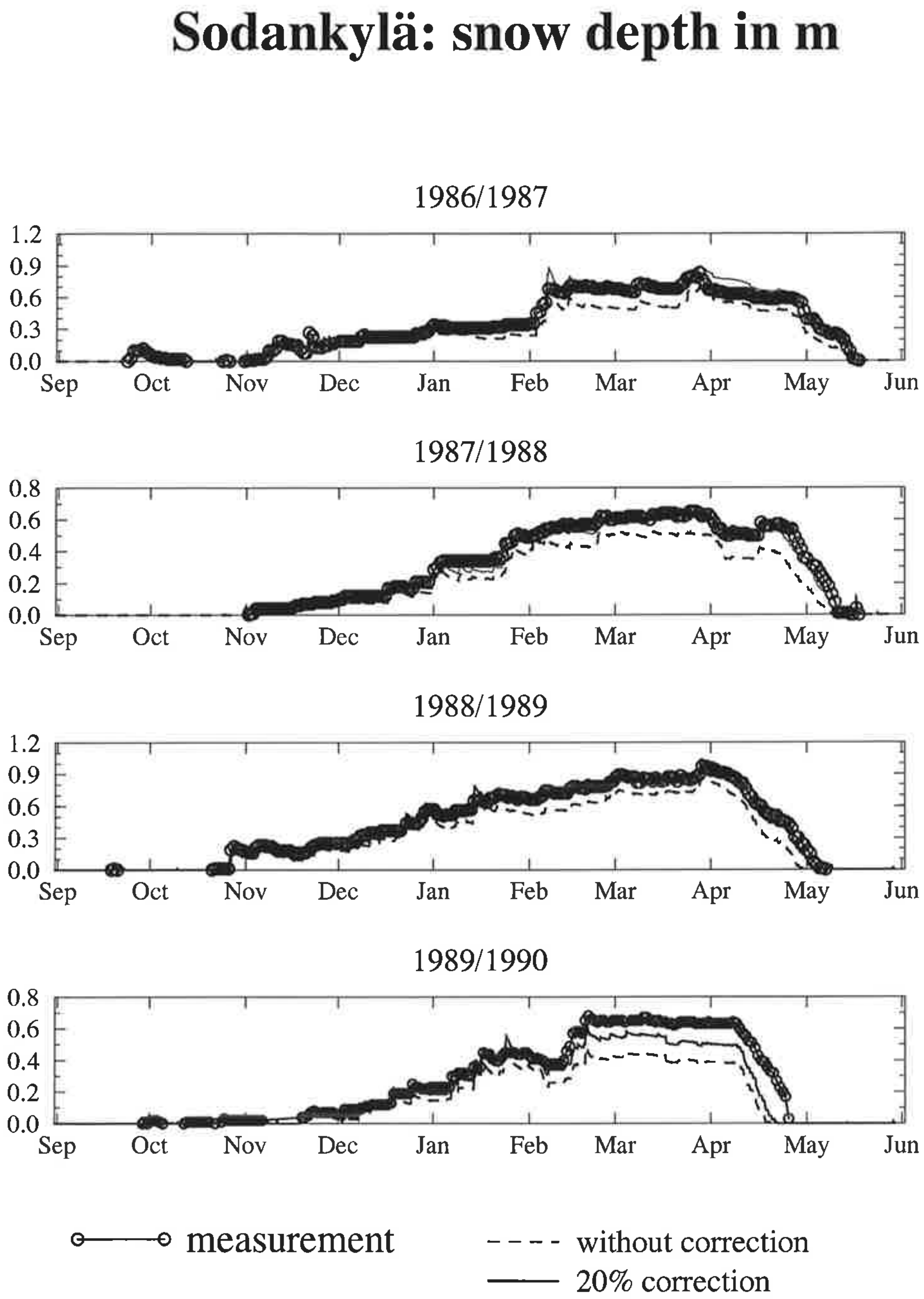

Fig. 9: Snow depth in Sodankylä for 1986-90. Shown are observations, simulations without any correction of the precipitation measurements and modelled results assuming a 20% measuring error for precipitation.

Svalbard Lufthavn: snow depth in m

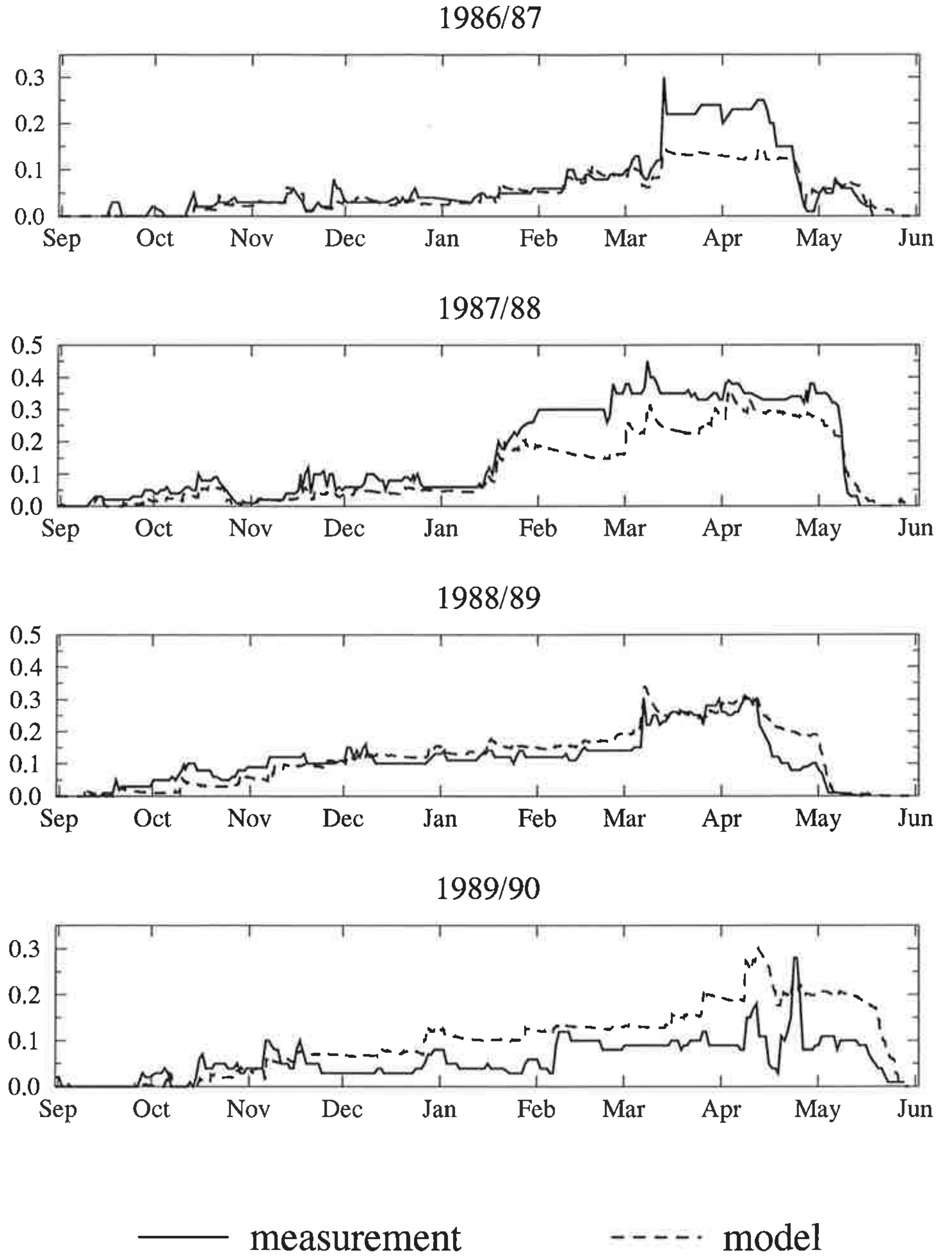

Fig. 10: Observed and simulated snow depth at Svalbard Lufthavn (Spitsbergen) for 1986-90.

Gander: snow depth in m

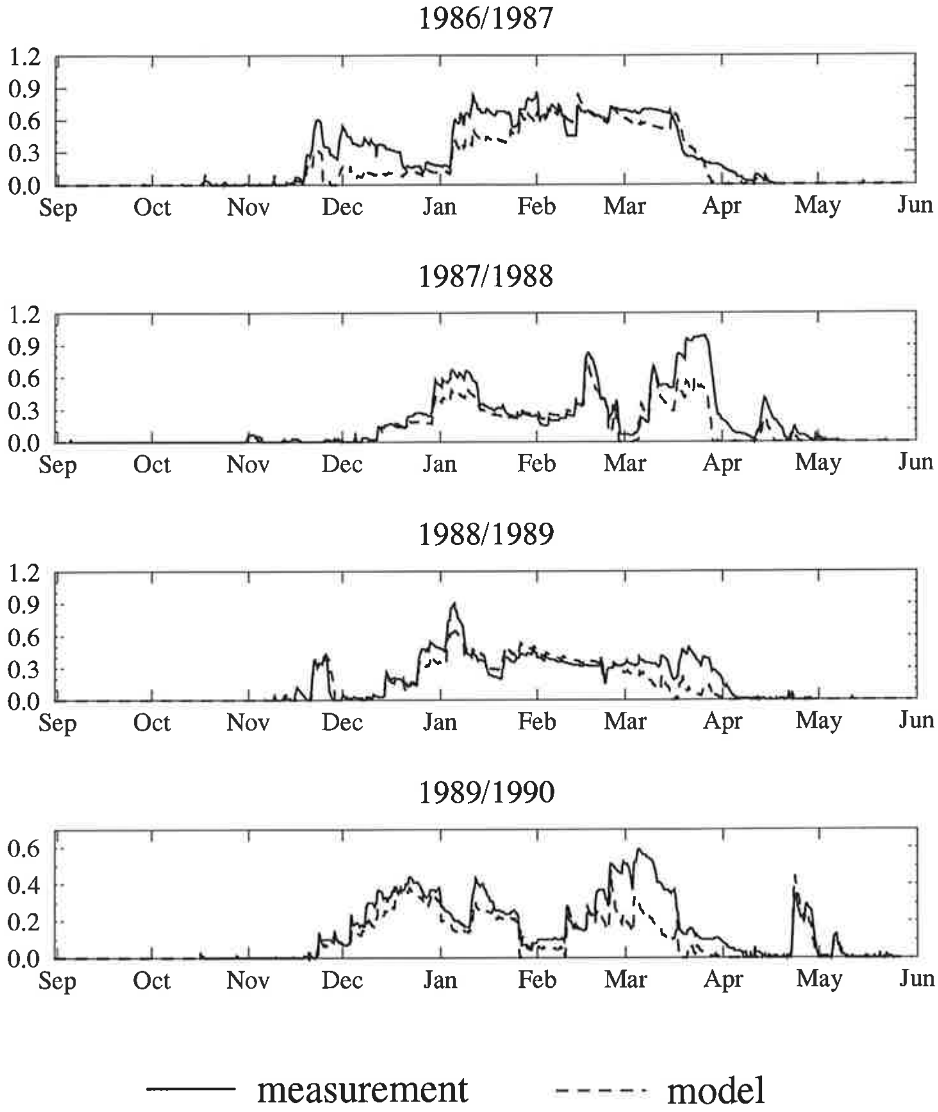

Fig. 11: Observed and simulated snow depth at Gander (Newfoundland) for 1986-90.

Zugspitze: snow depth in m

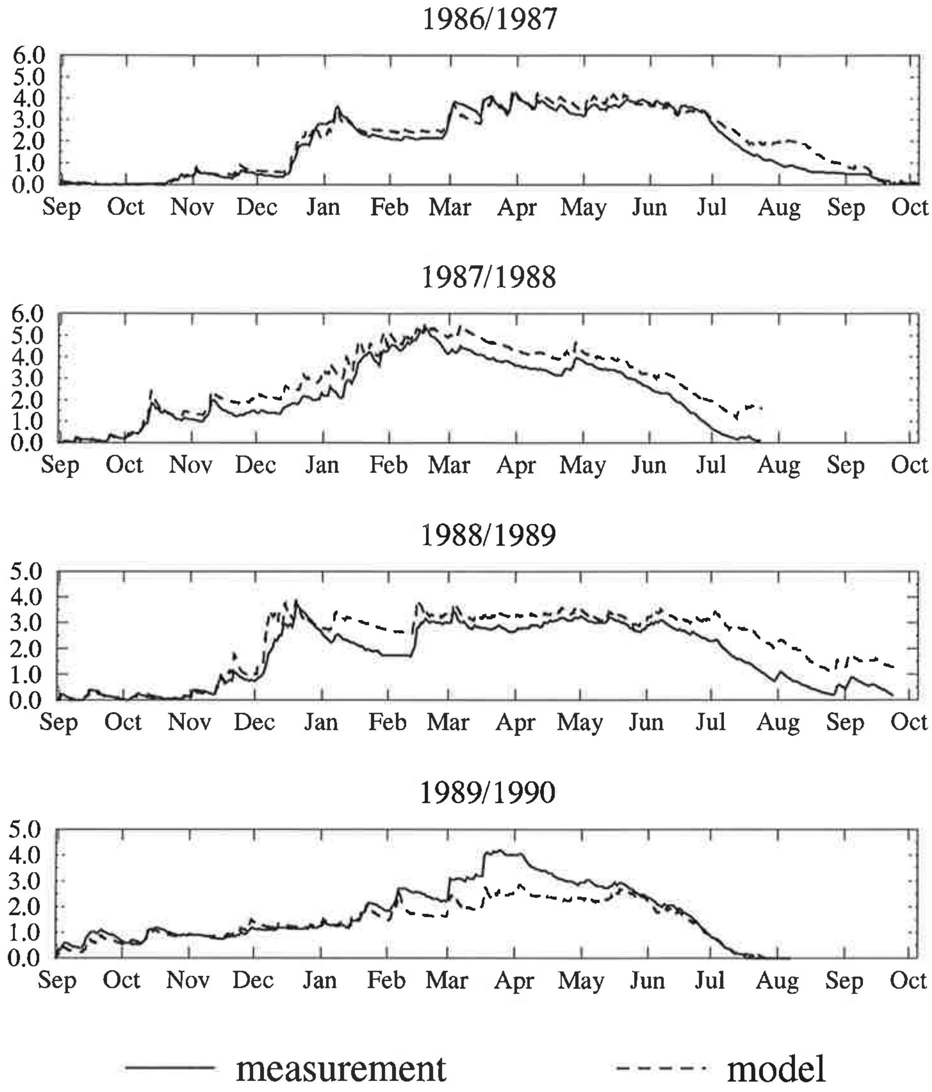

Fig. 12: Observed and simulated snow depth at Zugspitze for 1986-90.

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-vapour 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-vapour flux and latent heat flux with respect to snow surface temperature

Tab. 1: Input parameters of the snow-cover model for cases coupled with atmospheric models

symbol	input parameter
T_A	air temperature at 2 m
T_{D},h_{rel}	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

Tab. 2: Input parameters of the snow-cover model in stand-alone simulations

Modelling the Snow Cover for Climate Studies

Part II: The Sensitivity to Internal Snow Parameters and Interface Processes

Bettina Loth and Hans F. Graf

Max-Planck-Institut für Meteorologie Hamburg

Abstract

In order to find an optimal complexity for snow cover models within GCM[1]s, the climatic relevance of single snow processes on both the snow mass balance and the energy fluxes between snow cover and atmosphere has been investigated. Using a sophisticated model, experiments were performed under several different atmospheric and regional conditions (Arctic, mid-latitudes, alpine regions). A successful simulation of snow covers in GCMs is only guaranteed with a multi-layered snow cover model resolving the internal snow processes. Otherwise, large errors can occur, mostly in zones which are of paramount importance for the entire climate dynamics. Due to simplifications of such a model, the mean energy balance of the snow cover, the turbulent heat fluxes and the long-wave radiation at the snow surface may alter by between 1 and 8 W/m^2. The snow-surface temperatures can be systematically changed by about 10 K.

1 Introduction

Atmospheric processes are strongly affected by the fluxes at the lower boundary. The surface parameters, in particular the surface albedo and the evaporation from the earth's surface, are important links of global and regional feedback mechanisms (Walsh, 1993). These parameters are drastically changed by the occurrence and ageing of snow cover. For this reason, snow anomalies enormously influence the climatic processes on time scales ranging from days to thousands of years (Walsh, 1993). In spring in particular, snow cover anomaly patterns are closely related with changes in air temperature and with parameters describing the atmospheric circulation (Groisman et al., 1994a,b; Gutzler and Rosen, 1992). The climate variability of mid and high latitudes from autumn to early summer is highly dependant upon extremely variable snow conditions.

Unfortunately, the results of numerical sensitivity studies mainly depend upon the models and the parameterizations used. Randall et al. (1994) showed that 14 atmospheric models responded with 14 different snow-feedback mechanisms to a definite snow anomaly. The results differed in both sign and magnitude of the anomalies. In atmospheric models the snow physics is often only roughly parameterized. Internal processes such as temperature diffusion, phase changes of water and ageing are frequently neglected. But, these processes significantly alter the energy

[1] General Circulation Models

and mass fluxes at the lower boundary of the atmosphere and thus control the interactions between atmosphere and snow cover. Thus, it has to be supposed that the surface parameters such as surface temperature as well as the energy and mass fluxes at the lower boundary of the atmosphere (radiation balance, turbulent heat fluxes, evaporation) might be erroneous. Differences in the end of the ablation period are very important for the simulation of the whole climate system, since the disappearance of the snow cover drastically changes the surface albedo, the thermal conductivity of ground and the roughness length. In atmospheric circulation models these enormous changes either induce or suppress complex feedback mechanisms. Differences in the water-vapour flux at the snow surface, which might be caused only by the use of different snow parameterizations, change the simulated concentration and transport of water vapour in the atmosphere. A shift in the proportion of sensible to latent heat, even if the energy balance of the snow cover remains the same, results in changes in the atmospheric circulation. This begs the question: Do atmospheric models that use crude snow physics sufficiently reproduce the important nonlinear processes regarding snow?

To answer this question we look for an optimized parameterization of the snow cover. This parameterization has to reproduce all climatically relevant feedback mechanisms between the snow cover and the atmosphere as well as being efficient in terms of computing capacities. Hydrological models are orientated towards a precise simulation of the runoff. In general, they are not able to determine the energetic exchange between the snow cover and the atmosphere. Simple snow parameterizations in atmospheric models concentrate on the snow mass parameters like the snow area or the water equivalent. The exchange fluxes of snow cover, atmosphere and ground, however, depend upon a variety of snow parameters. They are determined by the internal structure of the snow (snow type, vertical gradients of temperature and density, degree of ageing) as well as by the interface snow parameters (albedo, emissivity, roughness length, thermal conductivity between snow and soil). The fluxes at the snow surface control the energy and mass exchange between snow cover and atmosphere during the snow period and thus directly affect the atmospheric processes. Furthermore these fluxes have an integrated effect on the atmospheric processes by determining the sum of evaporation and runoff during the ageing period and consequently the amount and duration of the melt water occurring during the ablation period.

A snow-cover model which is designed for implementation into an atmospheric model has to solve the following problems:

(i) an accurate simulation of the snow cover duration, since the appearance and disappearance of the snow cover leads to drastic changes in the properties of the earth surface.

(ii) a good estimation of the exchange fluxes between snow cover, atmosphere and ground.

(iii) In addition the sub-model has to be efficient with respect to computer time.

This paper investigates the possibility of simplifying sophisticated snow cover models while at the same time keeping the simulation quality necessary for climate studies. In order to handle this problem, numerical sensitivity experiments were performed under different climatic conditions (Arctic, mid-latitudes, alpine regions). The tool, a multi-layered model resolving the internal snow processes (Loth et al., 1993), was carefully tested (Loth and Graf, 1995). It succeeded in describing both the physical properties of the snow cover and the coupling of atmosphere and snow.

The paper is structured as follows: a brief description is provided for both the multi-layered snow-cover model (Chapter 2) and the observed data sets (Chapter 3) which are used for the sensitivity experiments. Results are shown (Chapter 4) testing the dependence of the snow-cover simulation on the following:

(i) the procedure resolving the vertical snow structure.
(ii) the parameterization of the snow albedo.
(iii) the treatment of the liquid-water processes.
(iv) the parameterization of the internal heat conduction and heat exchange with the ground.

In Chapter 5 the results of the paper are summarized and some final conclusions are given.

2 Snow-Cover Model

The simulations considered in the following as the "basic version" were performed with an one-dimensional state-of-the-art snow cover model. The physics and verification of this multi-layered model are described in detail in Loth et al. (1993) and Loth and Graf (1996). The model resolves the internal snow processes of heat conduction, mass redistribution due to water-vapour diffusion, extinction of short-wave radiation, the storage and transmission of liquid water, the processes of dry and wet metamorphism as well as the passing of melting-freezing cycles. The prognostic variables of the model are the snow albedo, the depth of the layers, the snow temperature, the snow density and the liquid-water content. Assuming saturation, the water-vapour concentration represents a diagnostic variable. Working with a time step of 1 hour the minimum snow depth is 1 mm.

3 Data Sets

The test simulations were performed by using input data from a snow monitoring station in the **French Alps** (Col de Porte) and four standard synoptic stations, each representing a different type of climatic condition.

The Arctic climate determines the snow cover in Svalbard Lufthavn (Spitsbergen) and Sodankylä (Finland). At both stations, solar radiation and rain are absent from November to spring and the entire winter precipitation falls as snow. Whereas the conditions in Svalbard Lufthavn are extremely dry (180 mm/a), higher precipitation rates are observed in Sodankylä (500 mm/a), where snowfall events occur on 2 out of every 3 days in winter.

At the mid-latitudes, temperatures near to freezing point, melting water within the snowpack and rainfall accompany the snow-cover development from autumn to spring. Gander (Newfoundland) represents conditions at a flat land mid-latitude station. The data set from Col de Porte (French Alps) represents deep snow covers.

In comparison to Gander and Col de Porte, Zugspitze (German Alps) has extreme conditions due to its position. Because of its altitude, air and snow temperatures are low, snowfall occurs during the whole year and the precipitation rate is high (2000 mm/a). Fairly unpolluted air results in a high amount of global radiation for clear-sky conditions and a high snow albedo. Furthermore, strong winds influence the turbulent fluxes (including the evaporation from snow), causing horizontal redistribution of the snow and wind compaction as an additional metamorphism mechanism.

The data sets cover the time periods 1 September 1986 to 1 June 1990 (synoptic stations) and 17 December 1988 to 8 May 1989 (Col de Porte). The measuring intervals at the synoptic stations are 1 hour (air temperature, dew-point temperature, surface pressure, wind speed) and 3 hours (amount and type of clouds). The precipitation is recorded at 6-hour-intervals. In Col de Porte all measurements are taken hourly.

4 Results

4.1 Model Strategy

The snow-cover models which are implemented in atmospheric models often neglect the vertical structure of the snowpack (Verseghy, 1991; Douville et al., 1995). This seems to be insufficient, since melting-freezing cycles, which occur in a thin surface layer, strongly affect the surface temperature in spring and should be resolved. Another reason for having a vertical resolution of the snow cover is to enable a more detailed consideration of the ageing processes. Only a multi-layered model allows us to consider the properties of new snow (small amounts of thermal conductivity, high albedo). In simple models simulation inaccuracies occur since these properties are "lost" due to a mass-weighted merging of the fresh snow with the aged snow at the ground.

In order to study the influence of the simulations on the vertical resolution of the snowpack, a sensitivity experiment was performed keeping the complexity of the model's physics but changing the choice of the snow layers. Unlike the multi-layered version in the sensitivity experiment, all snow layers were merged at the end of each time step and a snow cover of 2 layers was built up. Results of this study are shown for two stations: Sodankylä and Gander.

In contrast to this simplified 2-layer version, the multi-layered model merges all snow layers only if the total snow depth is less than 2 cm. In order to avoid a large number of layers, neighbouring layers are combined if the difference in their snow density is less than 150 kg/m^3, the layers belong to the same snow type (wet snow, cold snow, compact ice) and the difference between the snow temperatures is less than 3 K. This procedure is not applied to the snow cover interface layers. A snow layer is merged with its neighbour if its depth does not exceed a thickness of 3 mm at the end of the time step.

In both versions the initial snow cover is built up of 2 layers of the same depth when snowfall results in a new snow depth of at least 3 mm. Further snowfall leads to an opening of a layer of new snow provided the layer depth exceeds 3 mm. In the case of light snowfall, the new snow is added to the uppermost snow layer and the new snow parameters of the surface layer are mass-weight estimated.

The results (Fig. 1 to Fig. 4) show that both the snow mass and the single components of the energy exchange at the snow surface depend on the vertical structure of the snow cover. The sensitivity is, however, different for Arctic (Sodankylä) and mid-latitude regions (Gander). In Sodankylä the vertical structure of the snowpack only affects the simulation during the ablation period and determines the variability but not the mean values of the snow-surface temperature. In Gander, variability and mean values of all snow parameters, as well as of the fluxes between the snow cover and the atmosphere, are affected by the choice of the snow layers. In order to estimate the magnitude of the changes, the annual mean of the daily deviations between the two model versions is calculated. The results (shown in Tab.1 for 1986/87) confirm that qualitative differences occur in the simulation accuracy of the 2-layer version for Arctic and mid-latitude conditions. Whereas for Sodankylä the differences between the multi-layered model and the 2-layer version are negligible (below 0.5 W/m^2), considerable changes occur for Gander. In comparison with the multi-layered model the mean daily energy balance is systematically reduced in the 2-layer version by 5 W/m^2. The mean daily turbulent heat fluxes decrease by 5 W/m^2 and 8 W/m^2 for sensible and latent heat, respectively. The mean daily flux of ground heat is even changed by 12 W/m^2. Explanations of these results will be given by considering the differences of the single snow parameters in detail for both conditions.

At **Sodankylä,** differences between the two model versions occur in both snow depth and water equivalent during the ablation period. As shown for 1988/89 (Fig. 1), using a 2-layer version results in a deeper snow cover, since melting does not occur until the total snow cover is heated to 0 ^{o}C. The delayed melting processes increase snow depth and the water equivalent in the 2-

layer version by between 1 cm and 12 cm and between 4 cm and 8 cm, respectively. Since the multi-layered version somewhat underestimates the snow depth, the 2-layer version seems to better correspond with the observed snow depth. However, the differences between the model versions are smaller than the systematic error of the model and the reason for the underestimation of the snow depth is the systematic error of the snowfall measurements (Loth and Graf, 1996). Thus the 2-layer version is "better" for the wrong reason. Simulation errors caused by the uncertainties of the measurements, in particular the snowfall rate, and simulation inaccuracies are compensating each other.

As was mentioned earlier, at Sodankylä the mean value of the snow-surface temperature is not influenced by the choice of the snow layers. However, the variability is reduced in comparison with the multi-layered model. This characteristic occurs in all four simulation periods. It results from an increase in the thickness of the uppermost snow layer and its thermal inertia. As a consequence, the turbulent heat fluxes as well as the energy balance of the snow cover are changed. The maximum of these changes ranges between 10 W/m^2 and 25 W/m^2. In comparison to the multi-layered model, the deviations in the 2-layer version's energy budget are always positive. This is due to the fact that the increased thermal inertia of the surface layer intensifies the energy exchange between snow cover and atmosphere.

The simulations for **Gander** start to differ at the end of the accumulation period in mid December to the beginning of January. The differences between the two model versions, which last until the end of the ablation period, reach their maximum at the end of March with 30 cm to 60 cm in snow depth and up to a 20 cm in water equivalent (Fig. 2). Compared with the observed data the snow accumulated on the ground is overestimated by the 2-layer version.

Drastic changes between the two model versions also occur in the energy fluxes at the interfaces. The mean value of the snow-surface temperature is changed by approximately 10 K. This is shown for the winter of 1986/87 (Fig. 3) and results from a deeper surface layer in the 2-layer version. Due to the small thermal conductivity of snow, the surface cools down in the multi-layered version. This is mainly caused by the relatively high long-wave emission. In the sensitivity experiment, a larger heat reservoir exists in the uppermost snow layer which prevents an effective cooling of this layer.

Further differences of non negligible magnitude occur in the energy balance of the snow cover and in the turbulent fluxes (Fig. 4). Using the 2-layer version, the energy balance is reduced by approximately 10 W/m^2 although the surface temperature is increased. This results from a higher evaporation and a larger latent heat flux than in the multi-layered version. The changes range between 20 and 50 W/m^2. Significantly higher values are also possible (e.g. 1989/90). The sensible heat flux is decreased in the 2-layer version by -10 to -20 W/m^2.

The choice of the snow layers also influences the runoff characteristics of the model. The multi-layered version shows a frequent runoff with values below 0.5 mm, whereas a runoff occurs only seldom in the 2-layer version, but then with higher intensity. This different model behaviour is again caused by the deeper layers of the 2-layer version. A larger mass has to be heated compared with the thin layers of the multi-layered model. If the snow temperature in the 2-layer version is at the freezing point, an energy surplus leads to a larger melt water mass.

4.2 Albedo

The determination of the snow albedo, which is usually the key snow parameter in contemporary climate models, can be divided in four sub areas:

(1) assessment of new snow albedo
(2) the increase in snow albedo in the case of slight snowfall
(3) changes in albedo during dry ageing and melting periods
(4) the influence of the underlying soil in the case of a thin snow cover.

The sensitivity to the snow albedo is analysed with the data set taken from the **Col de Porte** station (1988/89). The snow albedo is parameterized alternatively with a time-dependent and with a temperature-dependent function. The latter was taken from the Hamburg climate model ECHAM (Roeckner et al., 1992). The albedo parameterizations are given in Tab.2.

Observed and simulated data for the snow albedo (Fig. 5) correlate well in the basic version of the snow-cover model (Loth, 1995). The changes due to dry and wet metamorphism are described with a high degree of accuracy. In contrast, implementing the ECHAM-parameterization leads to an unrealistic high variability of the snow albedo during cold periods and even to erroneous mean values in the case of air temperatures near the melting pint. An albedo overestimation of +0.2 in December and about +0.1 from the end of January to February is followed by a drastic reduction in the snow albedo at the beginning of the melting period. The ECHAM-parameterization simulates markedly smaller values than observed for temperatures near to freezing point. These errors exert a strong influence on the estimation of the snow depth and the water equivalent, which decline too rapidly in March (Fig. 6). The exaggerated absorption results in an unrealistic period of snow-free conditions in April. The final end of the snow-cover period occurs about 8 days too early.

The altered snow albedo leads to differences in both temperature and fluxes at the snow surface. The daily mean values of the snow-surface temperature differ between the model versions by 0.1 K to 4.2 K (Fig. 7). From the end of January to mid February the measured temperatures are reproduced in detail by the basic version, whereas the use of a temperature-dependent albedo leads to a general reduction of the daily mean temperature by 1.0 Kto 1.5 K. In spring (March to May) the simulated values are systematically smaller compared to those of the basic version.

The energy balance changes systematically due to the modified parameterization of the albedo (Fig. 8).The ECHAM snow albedo leads to an increased energy balance from the end of January until the removal of the snow cover. The differences range between 2 W/m^2 and 25 W/m^2 (running average over 24 values).

The turbulent heat fluxes also show remarkable changes between the two model versions. Whereas the sensible heat flux is again decreased using the ECHAM-temperature function for the albedo (by -2 W/m^2 to -5 W/m^2 in the running average of 24 values), the latent heat flux is increased by 5 W/m^2 to 7 W/m^2.

4.3 Storage and Transmission of Liquid Water

The penetration of rain water into cold snow, which immediately causes freezing processes, represents an enormous heat source for the snow cover. Due to these phase changes of water, the response of the surface temperature to changes in the air temperature is delayed and an effective cooling is prevented. The crossing of the melt-freeze-cycle at the snow surface can lead to a constant surface temperature over a period of days to weeks. The growth of grain size and the increase in snow density are faster in wet snow than in dry snow. This results in an accelerated increase in thermal conductivity and in an exponential reduction of the snow albedo with time.

Sensitivity experiments are shown for two stations (**Gander and Col de Porte**). In Gander, the exchange fluxes between atmosphere and snow cover are strongly determined by the frequency and mass of liquid water in the snow, which occur during the whole winter. Melting processes drastically reduce the snow cover from mid December to March (by 15 cm to 30 cm). An adequate description of the liquid-water processes for the time following the liquid water events also important.

In contrast to Gander, at Col de Porte melting processes which directly influence the atmosphere only occur between March and May. High amounts of melt-water runoff are generated from mid April to mid May. Until the middle of April, the liquid water is mainly stored in the snow cover and melting is connected with multiple crossing of melt- freeze-cycles. Since the snow base temperature is the near freezing point during the entire snow-cover period, melting processes at the base occur only from December to March. They lead to a runoff of 1 mm/d to 8 mm/d.

Neglecting the liquid-water storage leads to systematically thinner snow covers. In Gander the differences between the basic version (Loth, 1995) and the simplified model range between -0.5 cm to -15 cm in snow depth and give mass differences between -1% and -10%. The end of ablation is shifted by about 5 days. These changes are shown in Fig. 9 for 1986/87.

Changes due to the neglecting of liquid water, which additionally occur in the energy fluxes at the snow surface, the snow-surface temperature and the energy balance of the snow cover, last up to two weeks. The turbulent fluxes differ between the model versions by 4 to 20 W/m^2 (running average of 24 values). The sign of the differences depends on the observed weather situation. The energy balance is changed by 4 W/m^2 to 20 W/m^2. Thus, for the mass of the snow cover as well as for the energy exchange between snow cover and atmosphere, the liquid water is of significant relevance. Corresponding to the changes in the energy fluxes, the snow-surface temperature is modified by 0.2 K to 0.8 K.

In order to analyse the dependence of the snow cover on the transmission of liquid water, two numerical experiments were performed for Col de Porte. The first experiment was calculated without any liquid-water transmission. In the second experiment the liquid-water transmission was neglected only in the surface layer of the snow cover (OS). Although a total neglect of liquid-water transmission leads to a high correlation between the modelled value of the snow depth and the measurements taken by the ultrasonic sensor (Fig.10), it leads to a large error in the simulation of the water equivalent. The magnitude of these simulation errors is -5 cm to -10 cm from January until the middle of April and even increases to -15 cm at the end of the ablation period. The basic version (Loth, 1995) and the total neglect of liquid-water transmission overestimate the snow depth from March on by approximately 20 cm, but determine the water equivalent with a relatively high degree of accuracy. Due to the gravity force neglecting liquid-water transmission in the uppermost snow layer improves the simulation results.

Different treatment of liquid-water transmission results in different snow-surface temperatures. Compared with the observations the neglect of liquid-water transmission results in smaller daily mean values (-1.5 K to -2 K) from the end of January until March (Fig. 11). This is not dependent on whether the process is excluded from the uppermost layer or the entire snow cover. The basic version (Loth, 1995) corresponds well with the measurements for this period. From March to May the simulation quality is improved in the sensitivity experiments. This leads to the assumption that the process of liquid-water transmission is slightly overestimated by the basic version (Loth, 1995).

The energy balance differs between the model versions in the range of 2 to 20 W/m^2 (Fig. 12). Compared with the basic version (Loth, 1995) the neglect of liquid-water transmission in the uppermost snow layer reduces the energy balance by -20 W/m^2 from the end of January to the beginning of February. A further decrease in the energy balance by another - 6 W/m^2 occurs if liquid-water transmission is excluded at all.

A total neglect of liquid-water transmission results in a too early runoff (Fig. 13). The reason for this model's behaviour is that melting processes in winter, which occur at the snow cover base, immediately cause a runoff. The exception of the uppermost snow layer from liquid-water transmission does not change the runoff characteristics of the snow-cover model.

4.4 Ground Heat Flux

Snow cover sensitivity to the ground heat flux is tested at Svalbard Lufthavn and Zugspitze. At Spitsbergen the snow-cover properties are influenced only by temperature-dependent processes (temperature diffusion, turbulent fluxes, long-wave radiation), since solar radiation is absent during the winter season. In addition to that, the snow cover is very thin at Spitsbergen. At Zugspitze, with maximum snow depths between 3 m and 5 m, the ground properties can be important too. Due to the small thermal conductivity of snow, melting processes at the base can occur during the whole snow cover period despite the low air temperatures. These melting processes directly change the snow depth and water equivalent, since water cannot be stored by underlying snow layers, and lead to modifications of the fluxes between snow cover and atmosphere.

In order to analyse the influence of the soil state (frozen or non frozen) under Arctic conditions the thermal conductivity of the soil was modified at Svalbard Lufthavn. The basic version (Loth, 1995) uses the default value of the model, a thermal conductivity of the soil of 0.3 W/(m·K). This value is typical for dry sand and loam (Linke and Baur, 1970). In the sensitivity experiments the thermal conductivity of the soil is set to 0.6 W/(m·K) and 0.9 W/(m·K).

The higher thermal conductivity of the soil results in a more intense heat exchange between the snow cover and the underlying soil. The changes in the energy fluxes at the snow surface are 2 W/m^2 and 4 W/m^2 for an increase in the thermal conductivity of the soil to 0.6 W/(m·K) and to 0.9 W/(m·K), respectively. The daily mean of the snow-surface temperature changes by 1 K to 2 K. A response is also simulated in the snow depth, the water equivalent and the energy balance of the snow cover. Fig. 14 shows these results of the period 1988/89. The increase in the thermal conductivity of the soil from 0.3 W/(m·K) to 0.6 W/(m·K) results in the earlier occurrence of the melting processes and an earlier end to the snow-cover period. The energy balance is increased from the middle of April until to the disappearance of the snow cover. A further enhancement of the thermal conductivity of the soil does not markedly change the simulation.

At Zugspitze a higher soil temperature leads to systematic changes in the snow depth and in the water equivalent (Fig. 15). An increase in the soil temperature by 0.5 K causes a reduction of the snow depth by - 30 cm to -40 cm and a decrease in the water equivalent by -20 cm to -30 cm. The end of the snow-cover period is shifted by between two weeks and one month. The mean ground heat flux increases from 3 W/m^2 to 4 W/m^2 to 6 W/m^2 for the enhancement of the soil temperature by 0.5 K, and to 5 W/m^{2}to 9 W/m^2 for 1 K, respectively (running average of 24 values). In the basic version (Loth, 1995) the soil temperature at 5 cm is estimated by adding 0.5 K to the snow temperature at the base of the snow cover.

The results of these studies (Fig. 14 and Fig. 15) show that an increase in the ground heat flux during the snow-cover period does not significantly change the fluxes at the snow surface, although a higher soil temperature leads to systematic changes in snow depth and water equivalent. However, the fluxes between atmosphere and snow cover are drastically changed due to the earlier end of the ablation period (at Zugspitze up to 1 month). During the snow period differences in the turbulent fluxes and the snow-surface temperature do not exceed 2 to 4 W/m^2 (running average of 24 values) and 1 to 3 K, respectively.

4.5 Internal Heat Conduction

Further studies were performed using different parameterizations for the internal thermal conductivity of snow (Tab. 3). The bizarre new snow crystal structure causes a high pore volume which leads to a small thermal conductivity of loosely accumulated fresh snow, approximately 0.03 W/(m.K). Due to the ageing processes, the pore volume decreases and the thermal conductivity increases with time. These processes are parameterized by describing the density dependence of the snow's thermal conductivity. Whereas Anderson (1976) assumed a quadratic function, the thermal conductivity from Yen (1981) is based on a potential approach. In atmospheric circulation models the changes in the thermal conductivity of snow are usually neglected (e.g. Roeckner et al., 1992).

The results of the test studies (Fig. 16, Fig. 17) show the snow cover's strong dependence on the thermal conductivity of snow at mid-latitudes (**Gander, Zugspitze**). A constant thermal conductivity leads to an underestimation of the snow depth during the accumulation period and to an overestimation of snow depth in the ageing and melting periods. The snow depth changes between 2 cm and 10 cm for thin snow covers and between 1 m and 1.5 m at alpine stations. The differences depend on the frequency of the snowfall, the rate of metamorphism and the thermal state of the ground. If a constant thermal conductivity is used then the water equivalent increases in comparison with the basic version (Loth, 1995). At Gander these differences are about +2 cm and last from January to May. At Zugspitze changes in the water equivalent occur from spring to the end of the ablation period and range between 9 cm and 12 cm. Changes in the parameterization of the thermal conductivity of snow for both thin and deep snow covers results in changes in the turbulent fluxes (5 W/m2 to 15 W/m2). These last for the duration of the entire snow-cover period. The differences in the energy balance and the turbulent fluxes are very similar at Gander and Zugspitze. As a result, the internal heat conduction significantly influences the energetic exchange at the snow surface at mid-latitudes.

At Zugspitze, snow mass differences also occur due to a change of the density-dependant parameterization of the thermal conductivity of snow. The differences in the snow depth between the parameterization from Anderson (1976) and Yen (1981) range between 20 cm and

70 cm. The magnitude of these differences depends on the snowfall frequency, the rate of metamorphism and the thermal state of the underlying ground. The water equivalent is also systematically changed due to the different parameterizations of the thermal conductivity of snow. In comparison to the basic version (Loth, 1995) the approach from Yen (1981) leads to a reduction, from April to the end of the snow-cover period, of about 20 cm.

Larger differences in the energy balance at Gander occur in January 1990. The use of a constant thermal conductivity for the snow increases the energy balance by about 20 W/m^2. This is connected with a drastic reduction of the snow depth by -20 cm, down to a value below 1 cm. Since the melting processes follow low temperatures and new snow, a constant thermal conductivity of 0.176 W/(m·K) increases the temperature diffusion into the underlying snow layers compared with the basic version (Loth, 1995), and the temperature decreases. Consequently, the sensible and latent heat fluxes are changed by about 10 W/m^2 and 20 W/m^2 respectively (running average of 24 values). In the basic version (Loth, 1995), the snow cover's surface layer reaches melting temperature at an earlier point.

5 Summary and Conclusions

The sensitivity studies discussed above strongly suggest that general circulation models need multi-layered snow-cover models in order to simulate all relevant interactions between the snow cover, the atmosphere and the ground (soil, glacier, sea ice). Simplifications lead to large errors, especially in the case of air temperatures near freezing point.

Snow cover in the mid-latitudes requires the resolution of the internal vertical gradients of temperature, density and liquid water. Otherwise, the surface temperature is not adequately simulated, the runoff behaviour is changed, and the end of ablation is shifted. In the transitional seasons the same is true for the high latitudes. The marginal zone of snow coverage is the zone of highest baroclinity and strongly influences the formation and tracks of cyclones. The simulated variability of the climate system and the reliability of the results of numerical experiments therefore depend strongly on the simulation quality in the mid-latitudes.

A temperature function of the snow albedo leads to accelerated melting processes. In comparison with a time-dependent parameterization the daily mean of the snow-surface temperature is modified by 0.1 to 4.2 K (Col de Porte, 1988/89). The end of the ablation period is shifted by approximately 8 days.

At mid-latitudes, the internal processes (liquid-water processes, heat conduction, ageing) and the ground heat flux strongly influence the snow cover during the whole winter season. Neglecting the liquid-water leads to a reduction of the snow mass by 1 to 10 %. Different parameterizations of the liquid-water transmission cause changes in the daily mean of the surface temperature, the water equivalent and the snow-surface energy budget. At Zugspitze, an increase in the soil temperature by 0.5 K results in a decrease of snow depth by -30 to -40 cm. The end of the ablation period is shifted by between 2 weeks and 1 month.

The snow cover at mid-latitudes reacts sensitively to the liquid-water processes. Neglecting the liquid water leads to systematically smaller snow depth and water equivalent values. The temperature and the energy fluxes at the snow surface are also changed. For thin snow cover the storage capacity for liquid water is the single control parameter, since the phase transformations solid/liquid often affect the entire snow cover. The simulation of deep snow cover requires a parameterization of both the storage and the transmission of liquid water. An overestimation of the liquid-water processes results in an underestimation of the variability of the snow-surface temperature in spring.

In the Arctic region the ground heat flux only influences the snow cover the ablation period. In spring and summer, changes in the ground heat flux lead to modifications in the fluxes at the snow surface (radiation balance, turbulent fluxes) and a temporal shift in the end of the ablation period. Information on the soil type, the soil state before the first snowfall and the temperature of the ground are necessary in order to estimate the date of snow disappearance. In the case of deep snow cover at mid-latitudes the ground heat flux influences the snow-cover properties (snow mass, end of ablation), but does not affect the fluxes at the snow surface. A changed ground heat flux influences the atmospheric processes as an integral effect at the end of the snow-cover period. However, an accurate determination of the end of ablation requires a precise estimation of the ground heat flux during the whole snow period.

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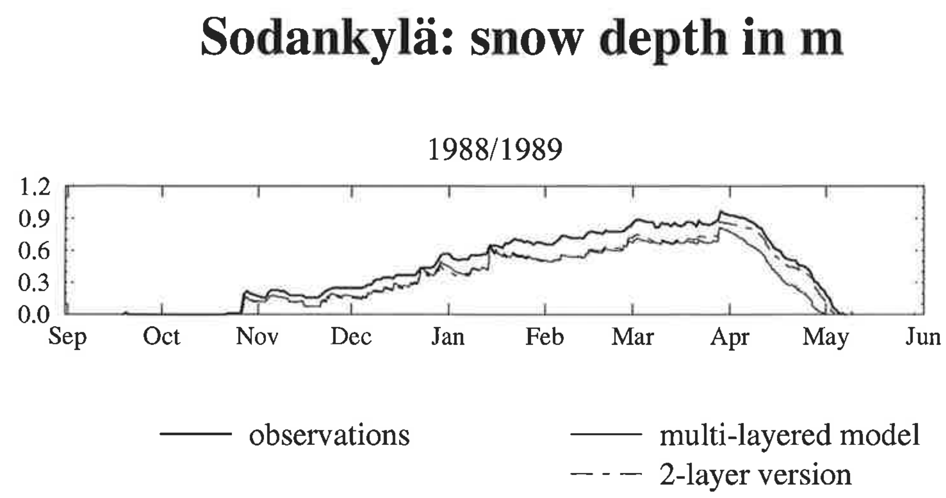


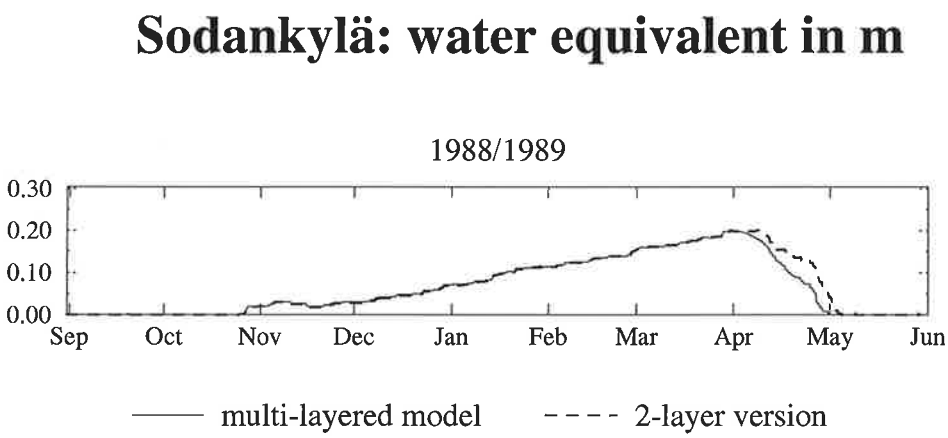


Fig. 1: Snow depth and water equivalent at Sodankylä for 1988/89 in dependence of the vertical resolution of the snow-cover processes. Shown are observed values of the snow depth and simulation results of snow depth and water equivalent using a multi-layered model and a 2-layer-version.

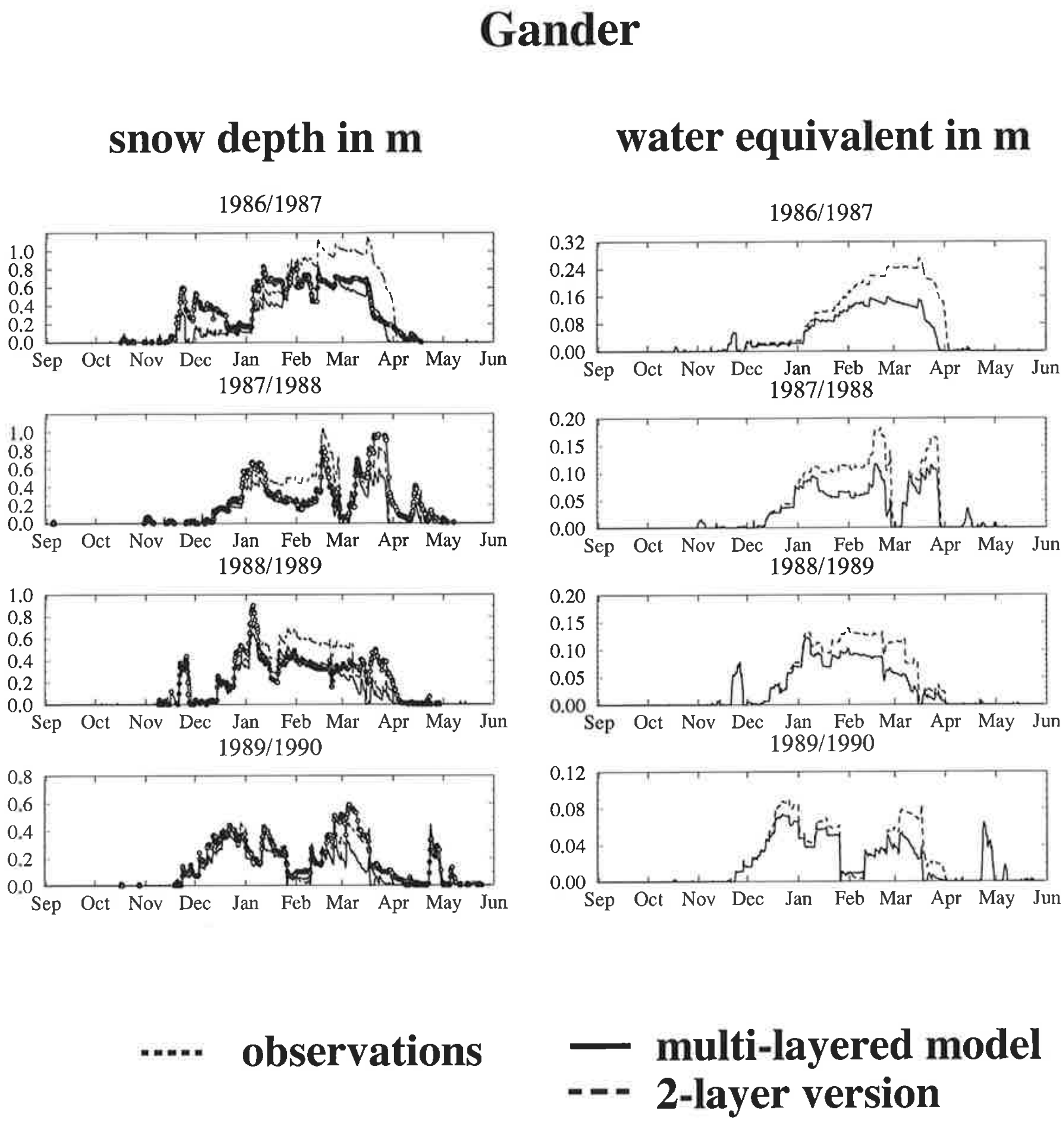


Fig. 2: Simulated snow depth and water equivalent at Gander in dependence of the vertical resolution of the snow model for 1986-90. Shown are observed values of snow depth and simulation results for snow depth and water equivalent using a multi-layered model (Loth, 1995) and a 2-layer version.

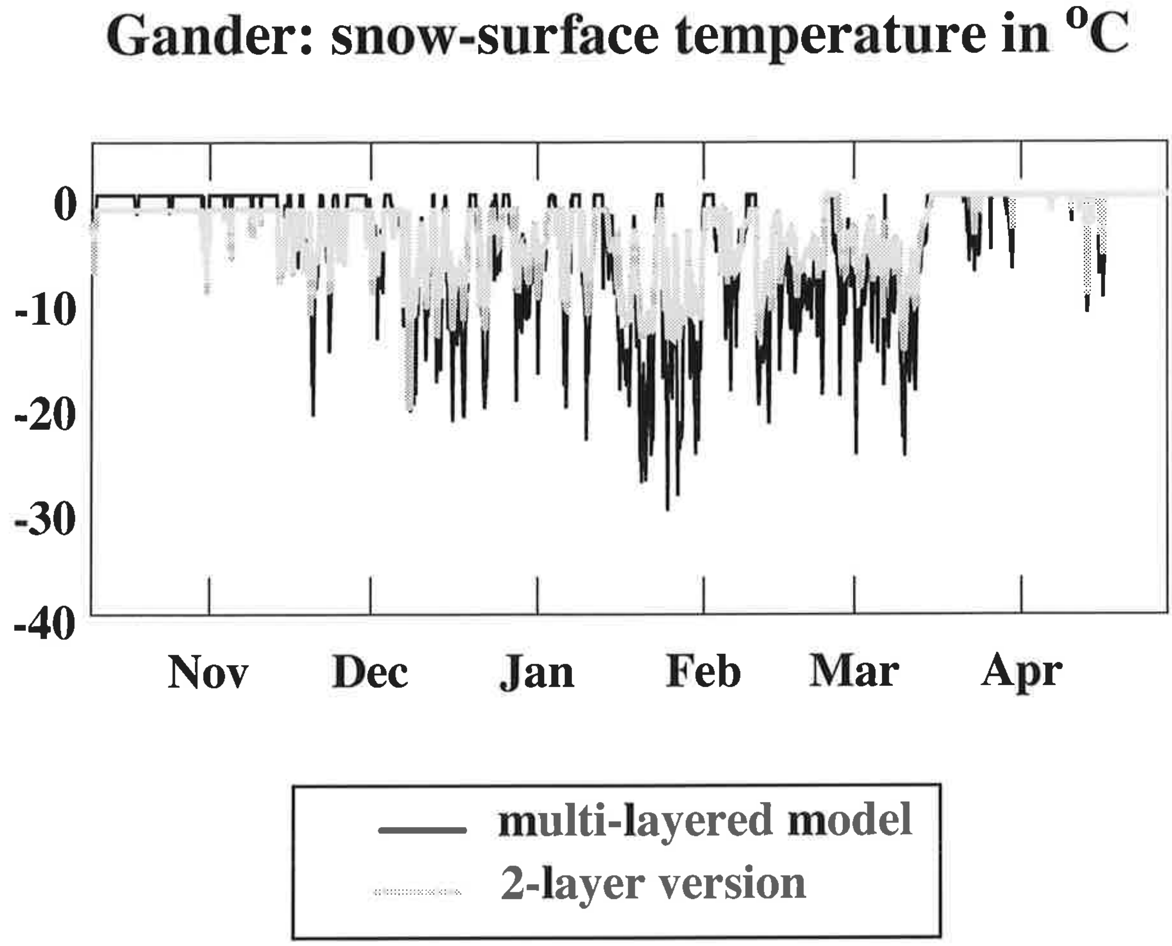
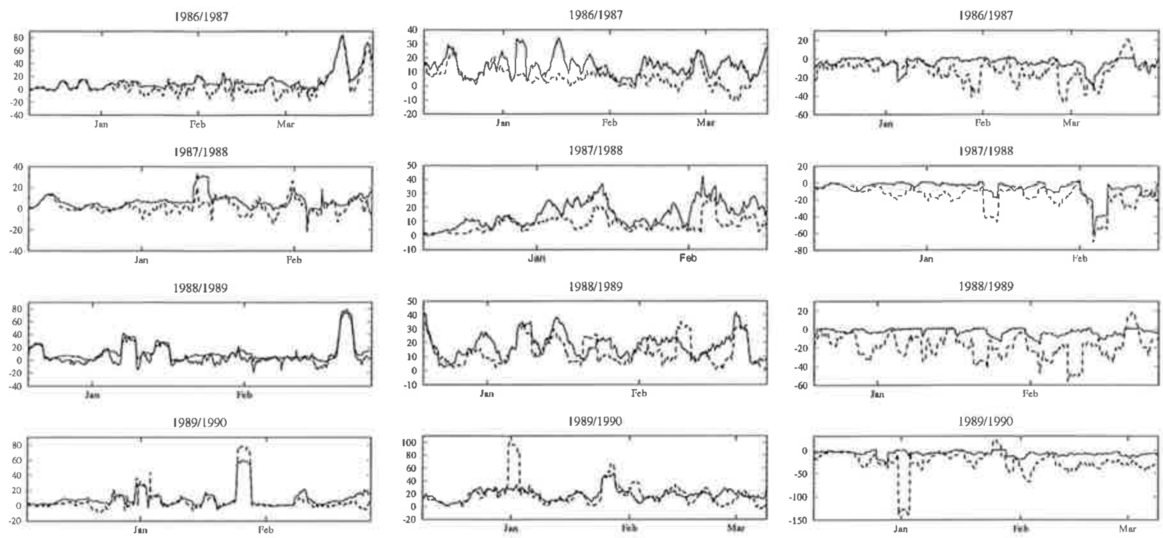


Fig. 3: Simulated snow-surface temperature at Gander for 1986/87 using a multi-layered model (Loth, 1995) and a 2-layer version (running average over 24 values).

Gander: fluxes at snow surface in W/m^2

energy balance | **sensible heat flux** | **latent heat flux**



— multi-layered model   - - 2-layer version

Fig. 4: Energy balance of the snow cover, sensible and latent turbulent heat flux at the station Gander for 1986-90. Shown are simulation results of a multi-layered model (Loth, 1995) and a 2-layer version as running average of 24 values.

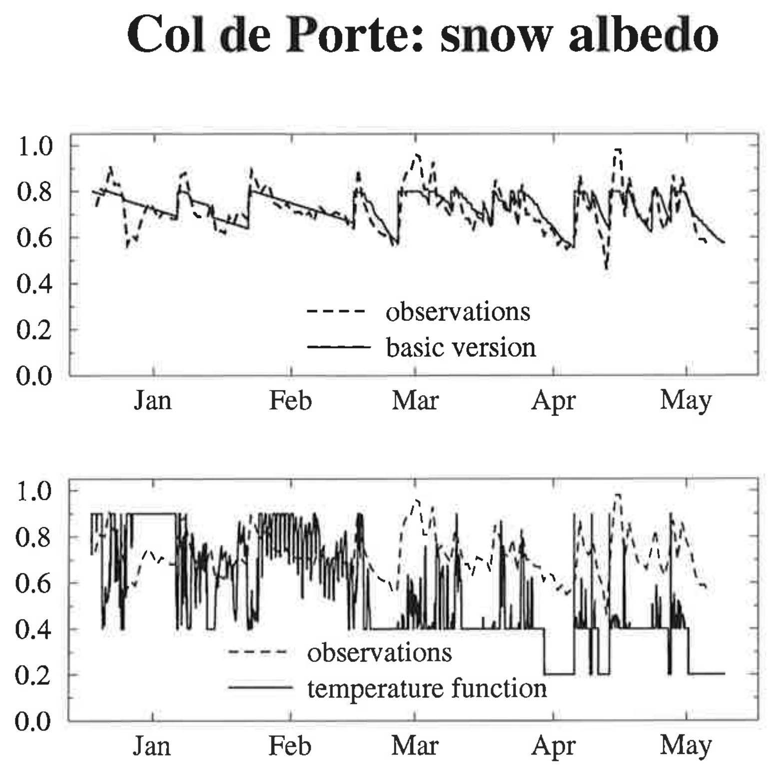


Fig. 5: Measured and simulated snow albedo for Col de Porte during the time interval 17 December 1988 to 8 May 1989. The simulations are performed with a time dependent ageing function and alternatively with a function of temperature for the albedo.

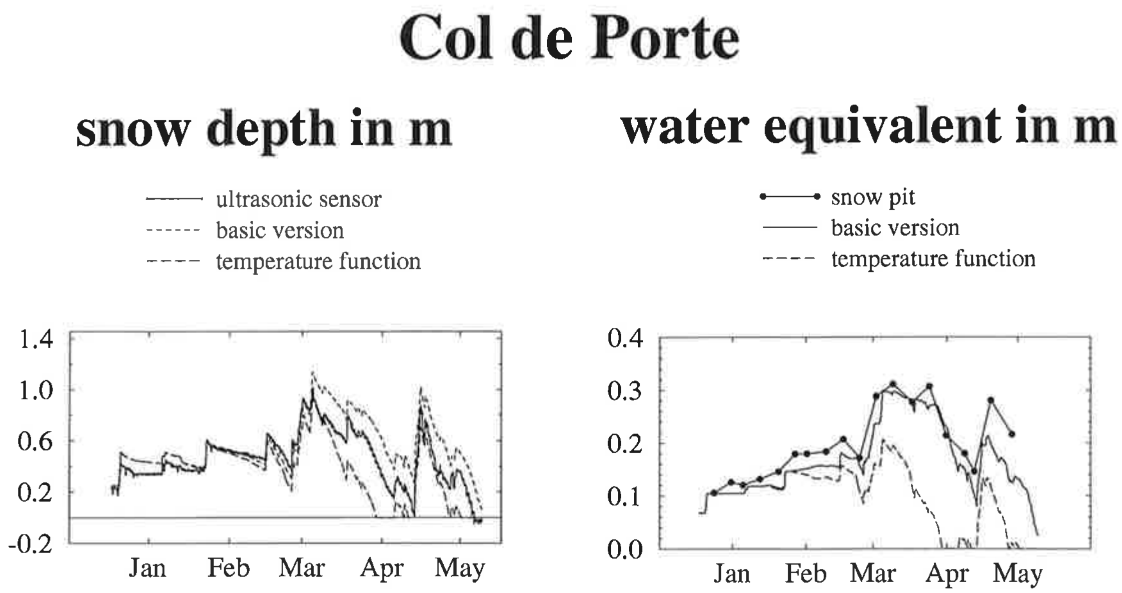


Fig. 6: Measured and simulated snow depth and water equivalent for Col de Porte during the simulation period 1988/89 using different parameterizations for the snow albedo.

Col de Porte: snow-surface temperature in °C

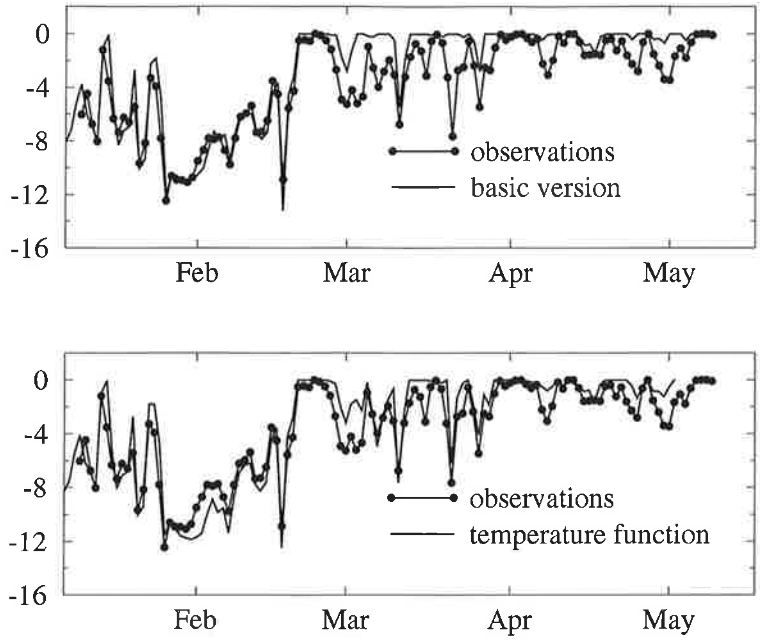


Fig. 7: Daily mean values of the snow-surface temperature in Col de Porte for the time period 8 January 1989 to 8 May 1989 from observed data and simulated results using different parameterizations for the snow albedo (see Tab. 4.1).

Col de Porte: energy balance in W/m^2

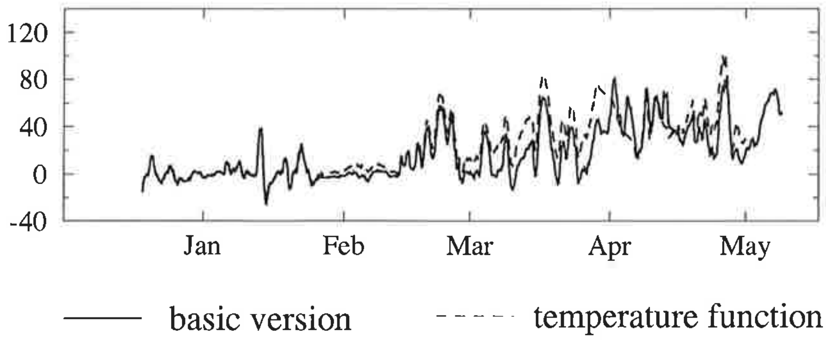


Fig. 8: Simulated values of the energy balance of the snow cover using different parameterizations for the snow albedo. Shown are the running averages over 24 values for Col de Porte during the time interval 17 December 1988 to 8 May 1989.

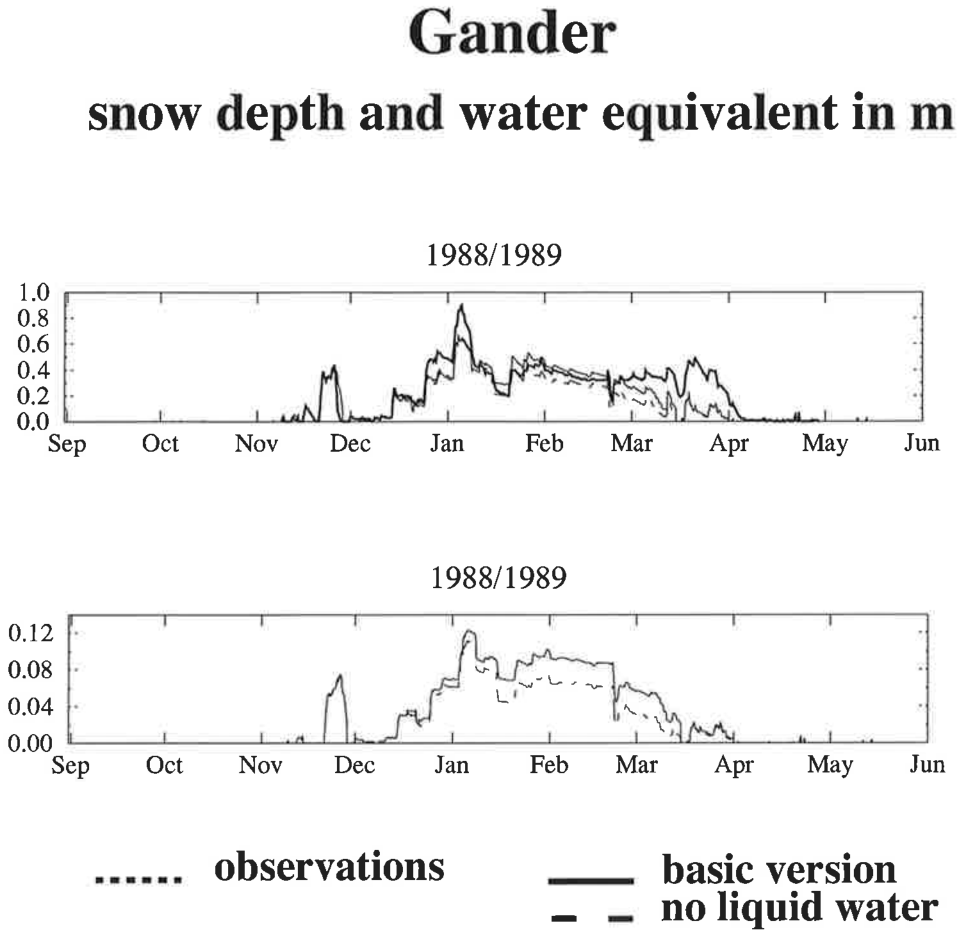


Fig. 9: Snow depth and water equivalent at Gander for the simulation period 1988/89. Shown are observed values of snow depth and simulation results using the basic version (Loth, 1995) and neglecting the storage of liquid water.

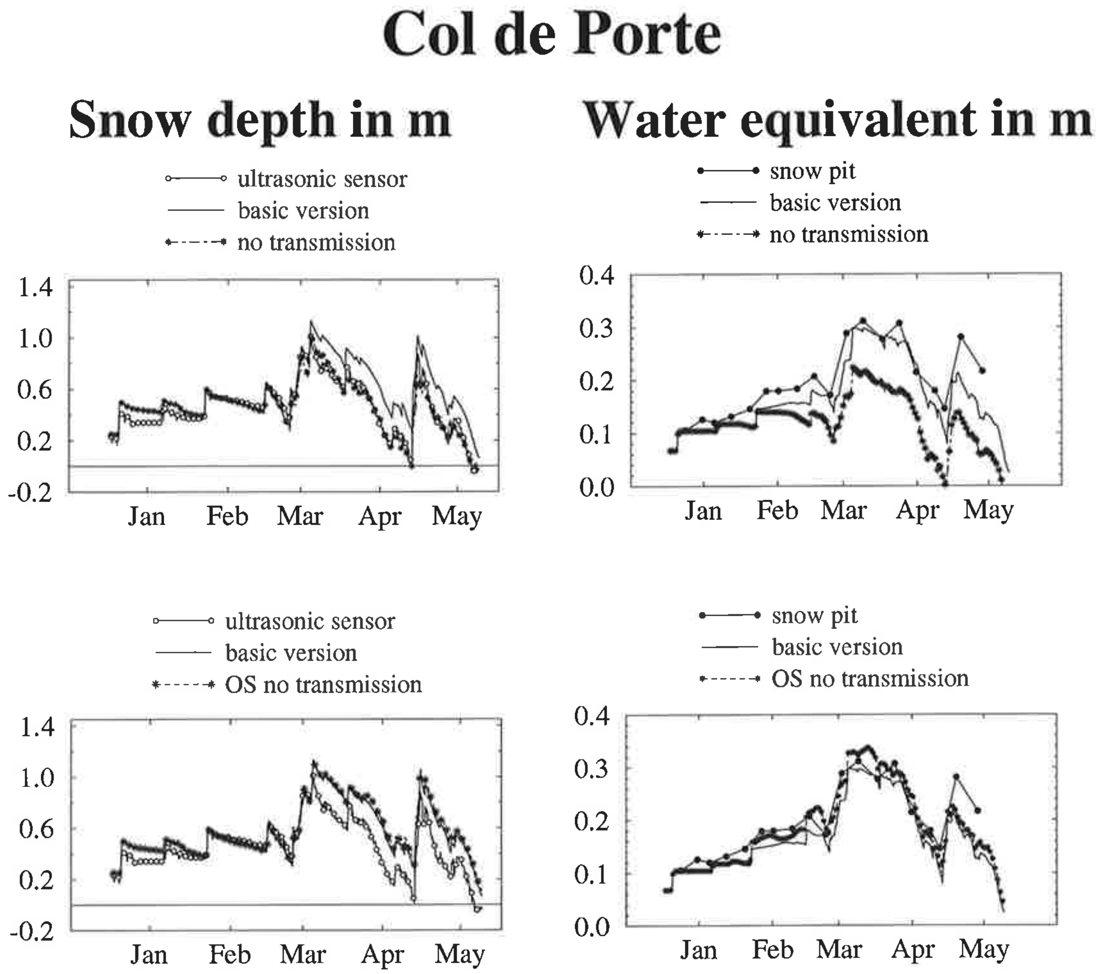


Fig. 10: Observed and simulated values of snow depth and water equivalent for Col de Porte. using different parameterizations for the transmission of liquid water within the snow cover. OS symbolise the surface layer of the snow cover.

Col de Porte: snow-surface temperature in °C

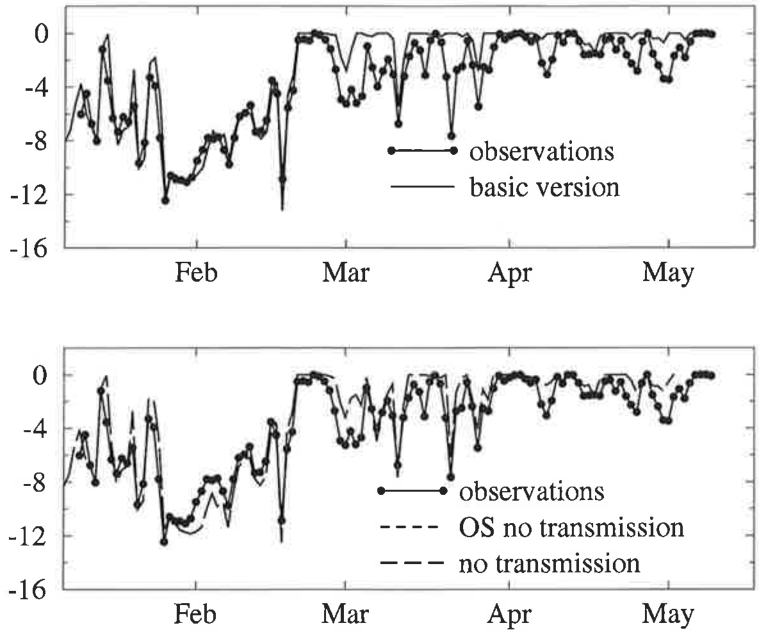


Fig. 11: Observed and simulated mean daily values of the snow-surface temperature using different parameterizations for the transmission of liquid water at Col de Porte for 8 January 1989 to 8 May1989.

Col de Porte: energy balance in W/m^2

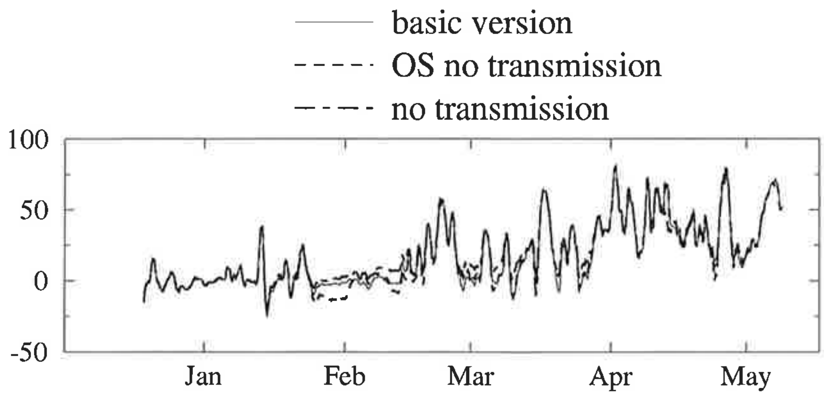


Fig. 12: Energy balance of the snow cover in dependence of the parameterization for the liquid-water transmission at Col de Porte during the time period 17 December 1988 to 8 May 1989 (running averages over 24 values).

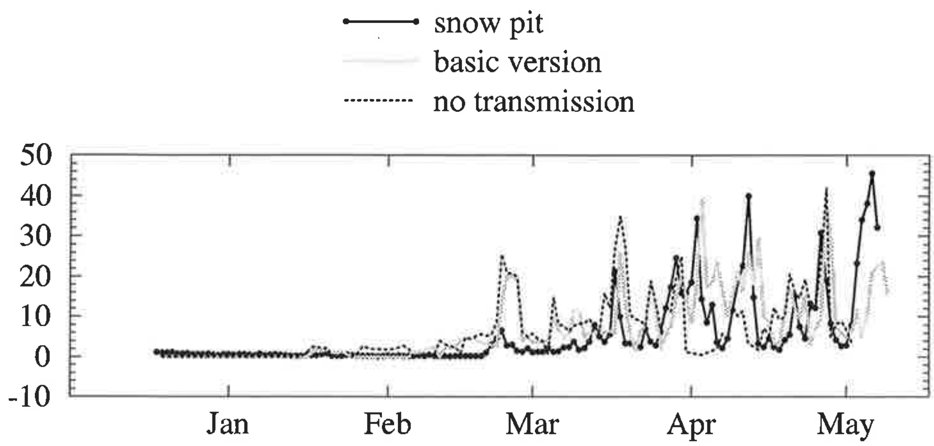
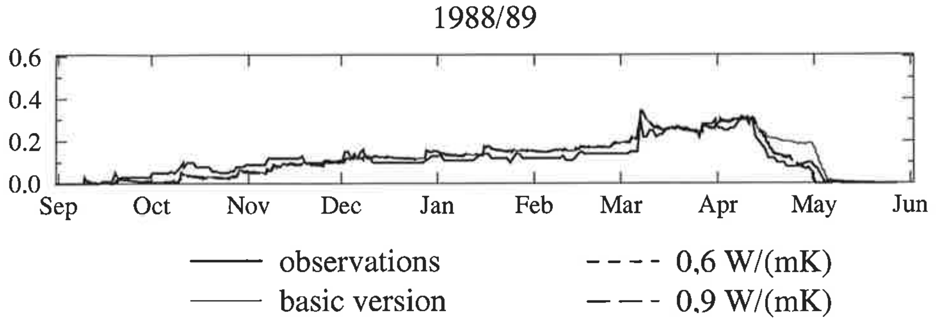


Fig. 13: Mean daily values of the observed and simulated runoff using different parameterization for the liquid-water transmission at Col de Porte during the time period 17 December 1988 to 8 May 1989.

Svalbard Lufthavn: snow depth in m



Svalbard Lufthavn: water equivalent in m

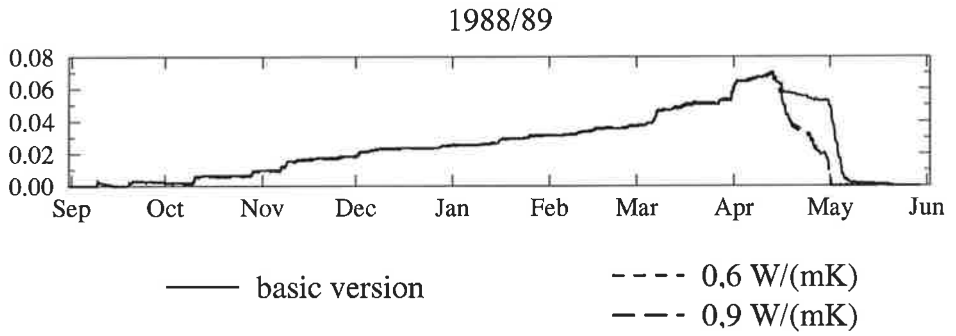
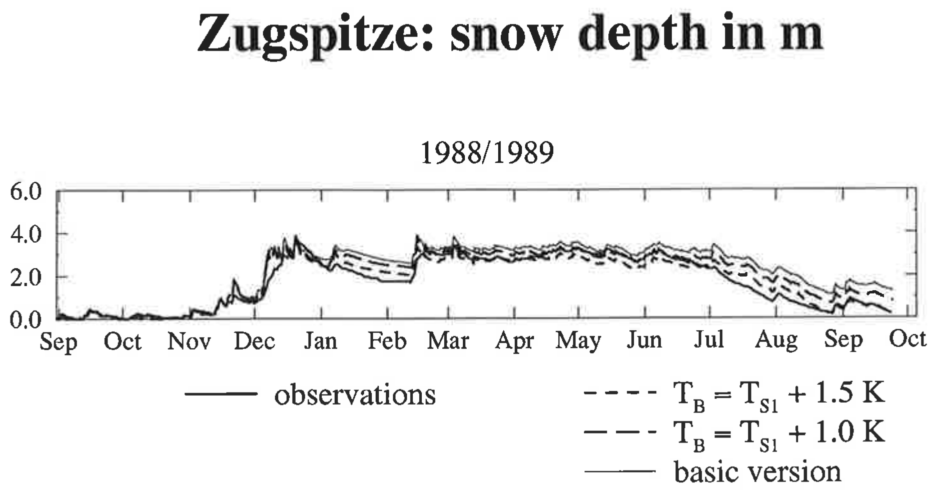


Fig. 14: Dependence of simulated values of snow depth and water equivalent on the thermal conductivity of the ground. Shown are the observed values of snow depth and simulation results using a thermal conductivity of soil of 0.3 W/(m·K), 0.6 W/(m·K) and 0.9 W/(m·K) for Svalbard Lufthavn during 1988/89.



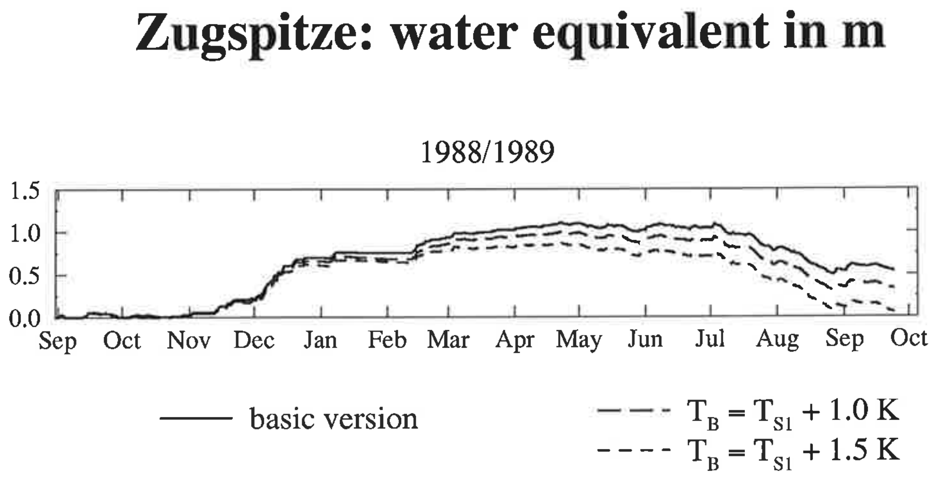
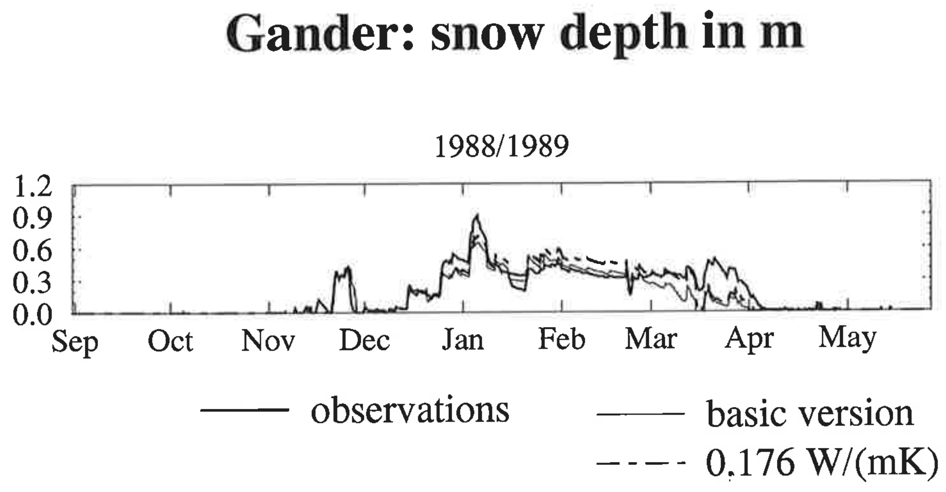


Fig. 15: Snow depth and water equivalent at Zugspitze for 1988/89 using different soil temperatures at 5 cm depth. Shown are observed snow depths and simulation of snow depth and water equivalent.



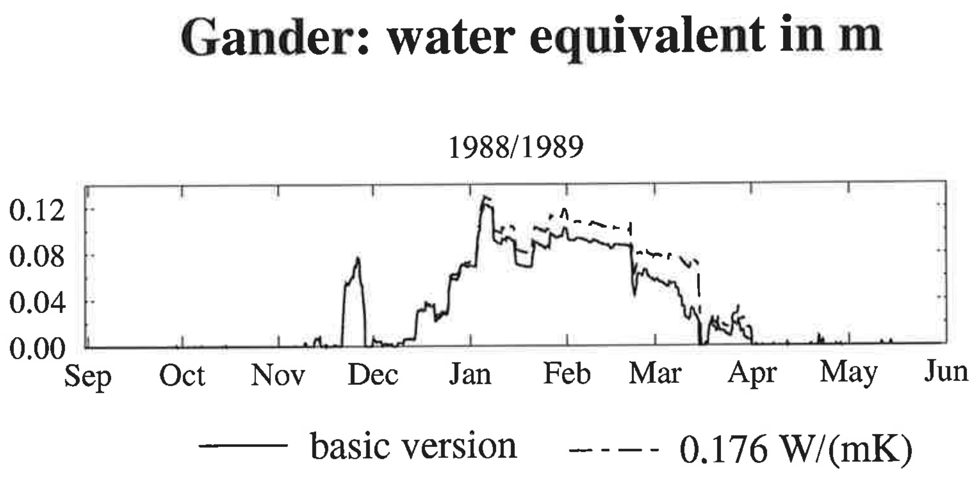


Fig. 16: Dependence of snow depth and water equivalent at the mid-latitude station Gander on the parameterization of the thermal conductivity of snow for 1988/89.

Zugspitze: snow depth in m

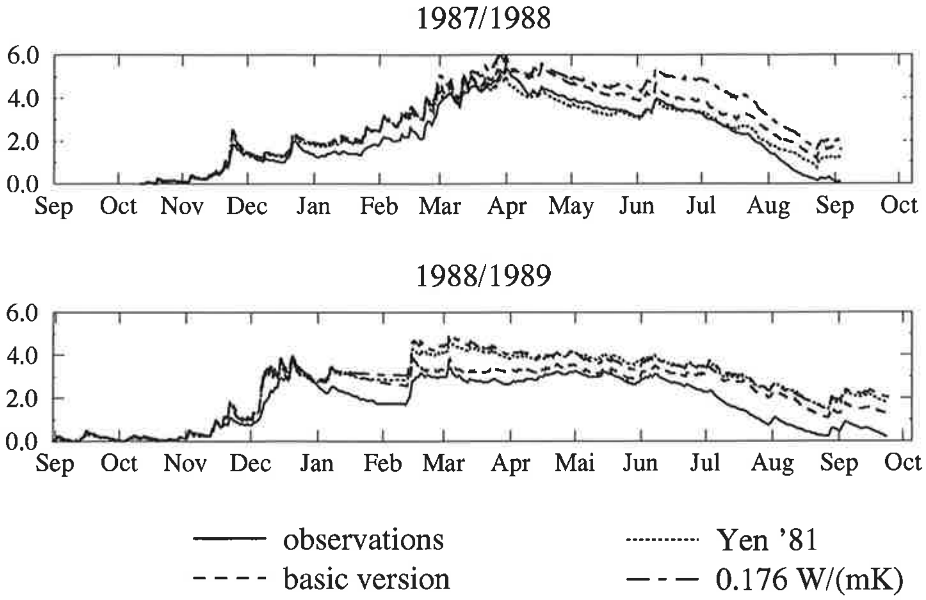


Fig. 17: Dependence of the simulated snow depth on the parameterization of the thermal conductivity of snow. Shown are the observed values snow depth at Zugspitze and simulation results using parameterizations from Anderson (1976), Yen (1981) and a constant value of 0.176 W/(m·K).

energy flux	absolute differences in W/m^2 Sodankylä	Gander
sensible heat flux at snow surface	-0.04	-5.3
latent heat flux at snow surface	0.3	-7.9
long-wave net radiation	-0.1	3.1
ground heat flux	-0.3	11.9
direct heat due to rain	~ 0.0	~ 0.0
energy balance	-0.3	-4.9

Tab. 1: Mean daily differences in energy fluxes using a 2-layer version of an snow-cover model and a multi-layer model (Loth, 1995) for the period 1986/87 at the stations Sodankylä (Finland) and Gander (Newfoundland).

	formula	coefficients	reference
basic version cold snow	$\alpha_s(t) = \alpha_s(t-1) + a_\alpha \cdot \Delta\tau$	a_α = -0.006/d	Gray and Landing (1987)
basic version melting period $h_s > 25$ cm	$\alpha_s(t) = a_\alpha + (\alpha_s(t\text{-}1) - a_{\alpha.}) \cdot \exp(-b_\alpha \Delta\tau)$	$a_\alpha = 0.5$ $b_\alpha = 0.24$	Verseghy (1991)
basic version melting period $h_s \leq 25$ cm	$\alpha_s(t) = \alpha_s(t-1) + a_\alpha \cdot \Delta\tau$	a_α = -0.071/d	Gray and Landine (1987)
ECHAM (temperature function)	$\alpha_s(t) = a_\alpha - b_\alpha \cdot (T_{sf} - T_r)/(T_M - T_r)$	$a_\alpha = 0.8$ $b_\alpha = 0.4$ T_r =263.15 K T_M=273.16 K	Roeckner et al. (1992)
linear ansatz	$\alpha_s(t) = \alpha_s(t-1) + a_{\alpha 1,2,3} \cdot \Delta\tau$ $a_{\alpha 1}$ for cold snow $a_{\alpha 2}$ for melting snow &$h_s > 25$ cm $a_{\alpha 3}$ for $h_s \leq 25$ cm	$a_{\alpha 1}$=-0.006/d $a_{\alpha 2}$=-0.015/d $a_{\alpha 3}$=-0.071/d	Gray and Landine (1987)

Tab. 2: Parameterizations of the snow albedo in the basic version (Loth, 1995) and in sensitivity experiments. $\alpha_s(t)$ and $\alpha_s(t\text{-}1)$ are the snow albedo at the time t and t-1, respectively. $\Delta\tau$ is the time step of the model in days.

reference	formula	coefficients	remarks
basic version Anderson (1976)	$\lambda_s = a_A + b_A \cdot \rho_s^2$	$a_A = 0.02$ $b_A = 2.5 \cdot 10^{-6}$	
Yen (1981)	$\lambda_s = a \cdot (\rho_s / \rho_w)^b$	$a = 2.22$ W/(m·K) $b = 1.88$ $\rho_w = 1000$ kg/m^3	
	$\lambda_s = 0.176$ W/(m·K)		constant value

Tab. 3: Parameterization of the thermal conductivity of snow λ_s in the basic version (Loth, 1995) and the sensitivity experiments. ρ_s is the snow density in kg/m^3.

Tables of Figures (paper1)

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