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Modeling the snow cover in climate studies
2. The sensitivity to internal snow parameters and interface processes

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Abstract. In order to find an optimal complexity for snow-cover models in climate studies, the influence of single snow processes on both the snow mass balance and the energy fluxes between snow surface and atmosphere has been investigated. Using a sophisticated model, experiments were performed under several different atmospheric and regional conditions (Arctic, midlatitudes, alpine regions). A high simulation quality can be achieved with a multi layered snow-cover model resolving the internal snow processes (cf. part 1,[*Loth and Graf,* this issue]). Otherwise, large errors can occur, mostly in zones which are of paramount importance for the entire climate dynamics. Owing 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 W/m^2 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 sublimation at the Earth's surface, are important links of global and regional feedback mechanisms [*Walsh,* 1993]. These parameters are drastically changed by the occurrence and aging of snow cover. For this reason, snow anomalies enormously influence the climatic processes on timescales 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 from autumn to early summer is highly dependent upon extremely variable snow conditions at middle and high latitudes.

Despite a high sensitivity of atmospheric processes to snow anomalies, in atmospheric models the snow physics is only roughly parameterized. The internal snow processes such as temperature diffusion, phase changes of water, and aging are frequently neglected. However, these processes significantly alter the energy and mass fluxes at the lower boundary of the atmosphere, and 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. Unfortunately, the results of recent 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. This begs the question, Do atmospheric models that use crude snow physics sufficiently reproduce the important nonlinear processes regarding snow?

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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 toward 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 also concentrate on the hydrological snow parameters like the snow area or the snow water equivalent. A snow-cover model which is designed for implementation into an atmospheric model has to solve two problems: accurate simulation of the snow-cover duration and precise estimation of' the exchange fluxes between snow cover, atmosphere, and ground during the snow period. In addition, the submodel has to be efficient with respect to computer time.

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 aging) 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 aging period and consequently the amount and duration of the meltwater occurring during the ablation period.

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, midlatitudes, alpine regions). The tool, a multi layered model resolving the internal snow processes [*Loth et al.,* 1993], was carefully tested [*Loth and Graf,* this issue]. It succeeded in describing both the physical properties of the snow cover and the coupling of atmosphere and snow. A brief description is provided for both the multi layered snow-cover model and the observed data sets which are used for the sensitivity experiments. Results are shown testing the dependence of the snow-cover simulation, the procedure resolving the vertical snow structure, the parameterization of the snow albedo, the treatment of the liquid-water processes, and the parameterization of the internal heat conduction and heat exchange with the ground.

Please note that the multi layered snow model has to be considered as a second generation snow model in general circulation models (GCMs). In present versions of atmospheric models the snowpack is simulated with an energy and mass balance model and both the spatial variability and the vegetation masking effect are described using critical values of snow water equivalent [e.g., *Roeckner et al.,* 1992]. Thus, although the spatial variability and the influence of different vegetation types on both the snow properties and the energy exchange between atmosphere and surface are of the same importance for climate simulations and their parameterization should probably also be improved, this paper focusses on the snow physics only. The goal of this study is to investigate the physical long-term behavior of the snow cover, which is a complex three-phase system and is characterized by nonlinear physics.

2. Model and Data Sets

Snow-Cover Model

The simulations considered in the following as the "basic version" were performed with a one-dimensional state-of-the-art snow-cover model. The physics and verification of this multi layered model are described in detail by *Loth et al.* [1993] and in part 1 (*Loth and Graf,* this issue). The model resolves the internal snow processes of heat conduction, mass redistribution due to water-vapor 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-vapor concentration represents a diagnostic variable. The fluxes are positive if they are directed into the snow cover. The number of layers depends on the vertical structure of the snow cover. In general it ranges from two to five layers. The time step is chosen according to the frequency of the observations. A snow cover is initialized when snowfall leads to a new snow depth of at least 0.3 cm. The minimum snow depth the model is able to handle is 0.1 cm. This value is chosen for numerical reasons. When the snow-cover model is used within a GCM, this parameter is of less importance, since the grid-cell area is assumed to be patchy snow-covered if the snow water equivalent is less than 10 mm and 15 mm for surface albedo and the turbulent fluxes respectively (cf. part 1). In the case of thin snow covers, both the surface albedo and the thermal properties approach the properties of the bare soil with decreasing snow mass.

2.1. 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, 78.3°N, 15.5°E, 30 m) and Sodankylä (Finland, 67.4°N, 26.7°E, 180 m). 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/yr), higher precipitation rates are observed in Sodankylä (500 mm/yr), where snowfall events occur on 2 out of every 3 days in winter.

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

In comparison to Gander (49.0°N, 54.6°W, 150 m) and Col de Porte (45oN, 6oE, 1320 m), Zugspitze (German Alps, 47.4°N,11.0°E, 2962 m) 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/yr). 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 and sublimation), causing horizontal redistribution of the snow and wind compaction as an additional metamorphism mechanism.

The data sets cover the time periods September 1, 1986 to June 1, 1990 (synoptic stations), and December 17, 1988 to May 8, 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, and the model simulations, the results of which are shown in this paper, were performed with a +15% correction of the measured precipitation. No data correction has been applied for the simulations for the synoptic stations. According to the measuring frequency the model time step is set to 1 hour and 3 hours for the snow monitoring station Col de Porte and the standard synoptic stations, respectively. In the simulations with the data sets from the synoptic stations, we prescribe the soil temperature at 5 cm depth, whereas we coupled the snow-cover model with the soil model of the Hamburg gloabl climate model ECHAM [Roeckner et al., 1992] for the Col de Porte calculations.

The model performs well in simulating the snow cover (cf. part 1). The snow depth is slightly underestimated for Arctic conditions. The maximum differences between the simulated and observed values range between -10 cm and -20 cm. For Gander the maximum deviations between measurements and model results of between 10 cm and 40 cm occur in December, January, March, and April. At Zugspitze the snow depth curve is generally described by the model. The values are overestimated by 70 cm to 150 cm, in August 1987 and from December to the end of ablation of 1987-1988 and 1988-1989. For 1989-1990 the snow depth is underestimated by nearly the same amount from February to June. The alpine snow cover at Col de Porte, which has been verified with observed values of snow depth, water equivalent, surface temperature and albedo, is reproduced in detail.

3. Results

3.1. Model Vertical Discretization

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 aging 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 two layers was built up. Results of this study are shown for two stations: Sodankylä and Gander.

In contrast to this simplified two-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, neighboring 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 neighbor if its depth does not exceed a thickness of 0.3 cm at the end of the time step.

In both versions the initial snow cover is built up of two layers of the same depth when snowfall results in a new snow depth of at least 0.3 cm. Further snowfall leads to an opening of a layer of new snow, provided the layer depth exceeds 0.3 cm. 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 (Figure 1 to Figure 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

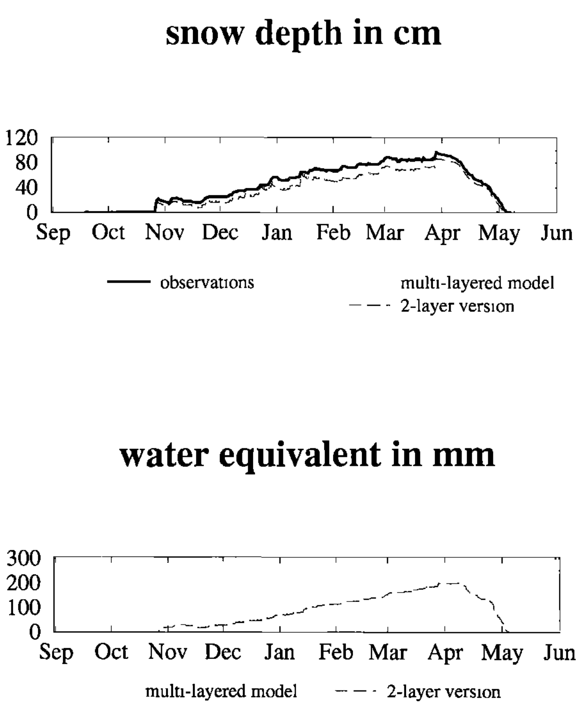

Figure 1. Snow depth and water equivalent at Sodankylä for 1988-1989 using models with a different vertical resolution of the snow-cover processes. Shown are observed values of the snow depth and simulation results of snow depth and water equivalent using a multi layered model and a two-layer version.

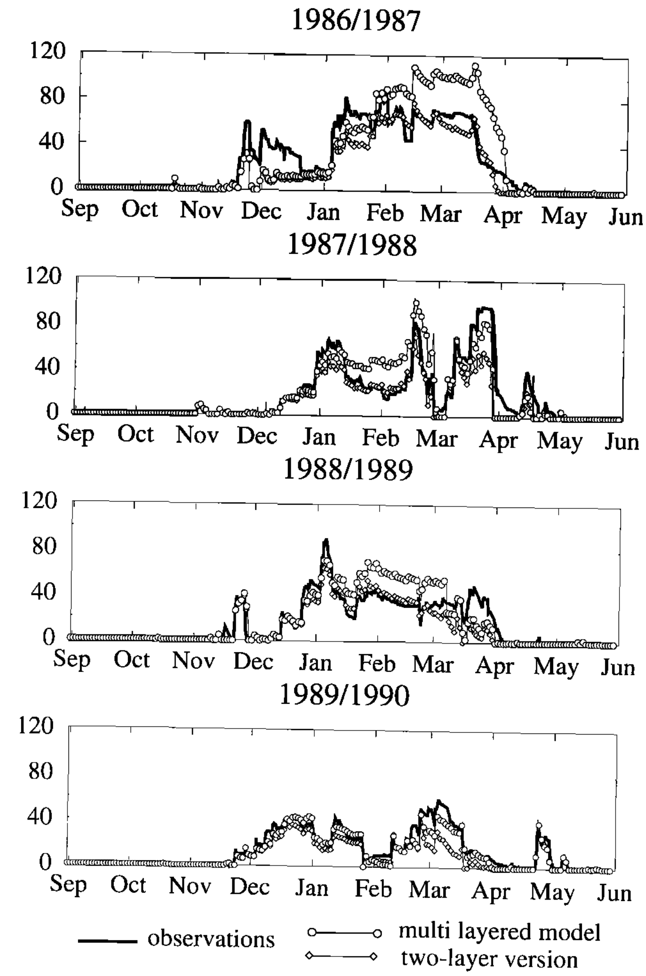

Figure 2a. Observed and simulated snow depth (in centimeters) at Gander using models with a different vertical model resolution for 1986-1990. Simulation results are shown using a multi layered model [*Loth,* 1995] and a two-layer version.

midlatitude 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 Table 1 for 1986-1987) confirm that a qualitatively different simulation accuracy of the two-layer version for Arctic and midlatitude conditions. Whereas for Sodankylä the differences between the multi layered model and the two-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 two-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.

Sodankylä. At Sodankylä differences between the two model versions occur in both snow depth and water equivalent during the

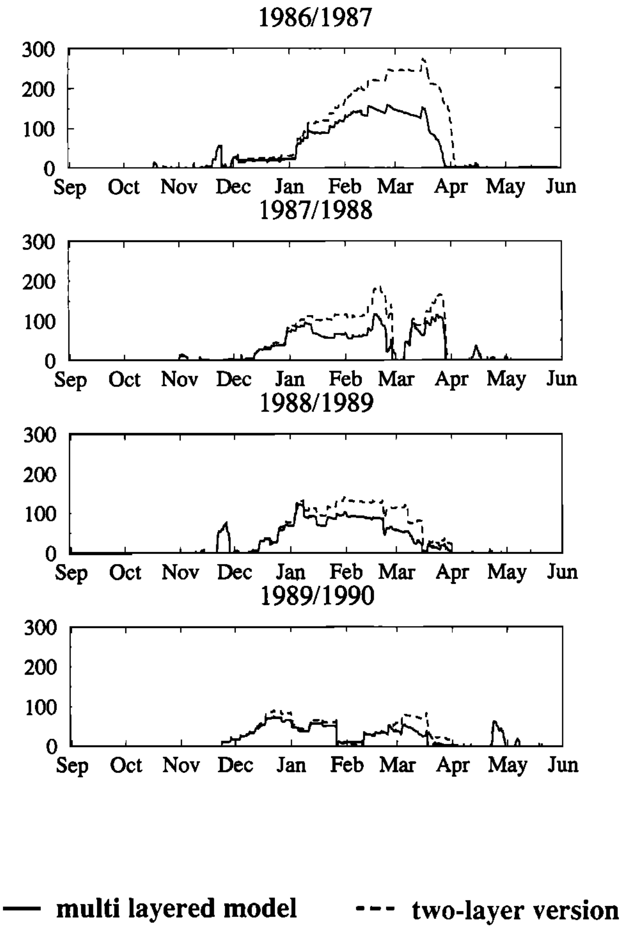

Figure 2b. Same as Figure 2a but for simulated water equivalent (in millimeters).

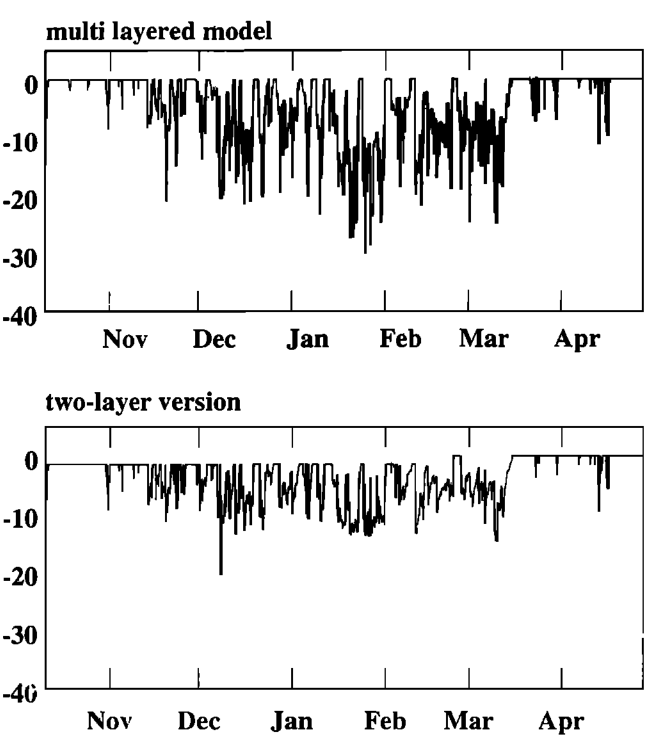

Figure 3. Simulated snow-surface temperature (in degrees Celsius) at Gander for 1986-1987 using a multi-layered model [*Loth*, 1995] and a two-layer version (running average over 24 values).

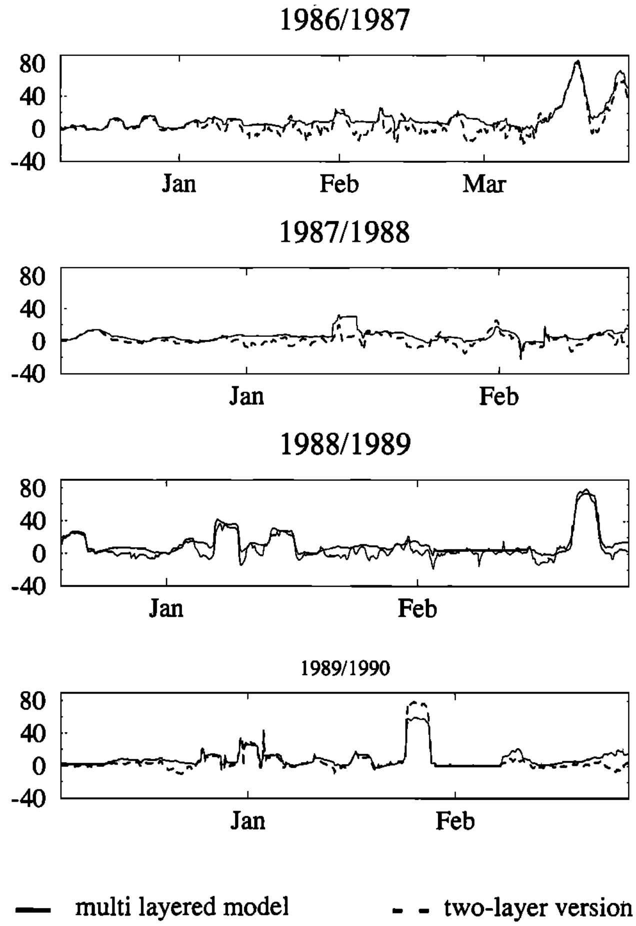

Figure 4a. Energy balance of the snow cover (in W/m^2) at the station Gander for 1986-1990. Simulation results of a multi layered model [*Loth,* 1995] and a two-layer version as running average of 24 values.

ablation period. As shown for 1988-1989 (Figure 1), using a two-layer version results in a deeper snow cover, since melting does not occur until the total snow cover is heated to 0°C. The delayed melting processes increase snow depth and the water equivalent in the two-layer version by between 1 cm and 12 cm and between 40 mm and 80 mm, respectively. Since the multi layered version somewhat underestimates the snow depth, the two-layer version seems to better correspond with the observed snow depth. However, the differences between the model versions are within the observation errors, and the reason for the underestimation of the snow depth is the systematic error of the precipitation measurements *[Loth and Graf*, this issue]. Thus the two-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 serve to compensate 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 two-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.

Gander. 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 200 mm in snow water equivalent (Figure 2). Compared with the observed data, the snow depth is overestimated by the two-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 1986-1987 (Figure 3) and results from a deeper surface layer in the two-layer version. Owing 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 (Figure 4). Using the two-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 than in the multi layered version. The changes range between 20 and 50 W/m^2. Significantly higher values are also possible (e.g., 1989-1990). The sensible heat flux is decreased in the two-layer version by -10 W/m^2 to -20 W/m^2.

The choice of the snow layers also influences the runoff characteristics of the model (drainage from the bottom of the pack at a point). The multi layered version shows a frequent runoff with values below 0.5 mm, whereas a runoff occurs only seldom in the two-layer version, but then with higher intensity. This different model behavior is again caused by the deeper layers of the two-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 two-layer version is at the freezing point, an energy surplus leads to a larger meltwater mass.

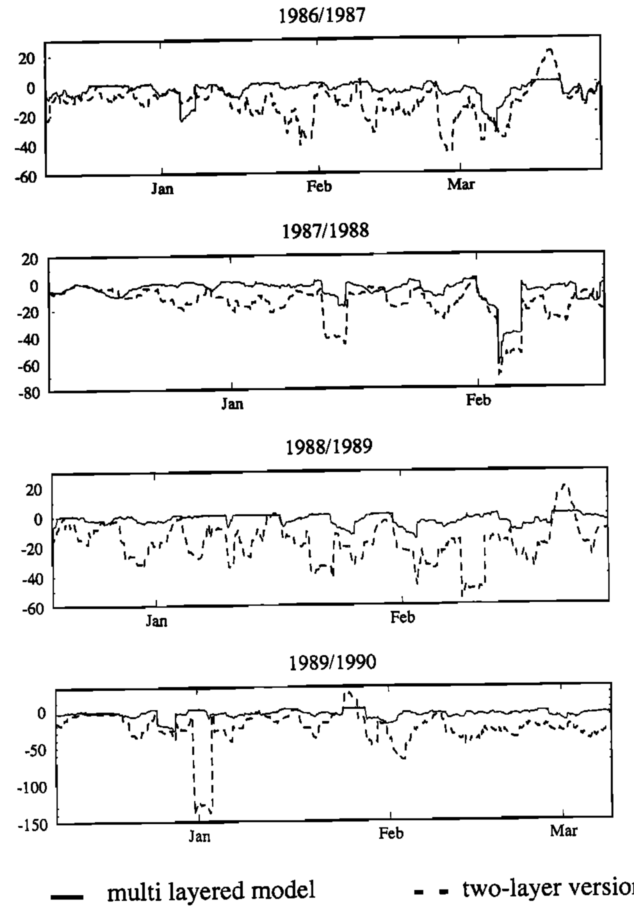

Figure 4c. Same as Figure 4a but for latent heat flux (in W/m^2).

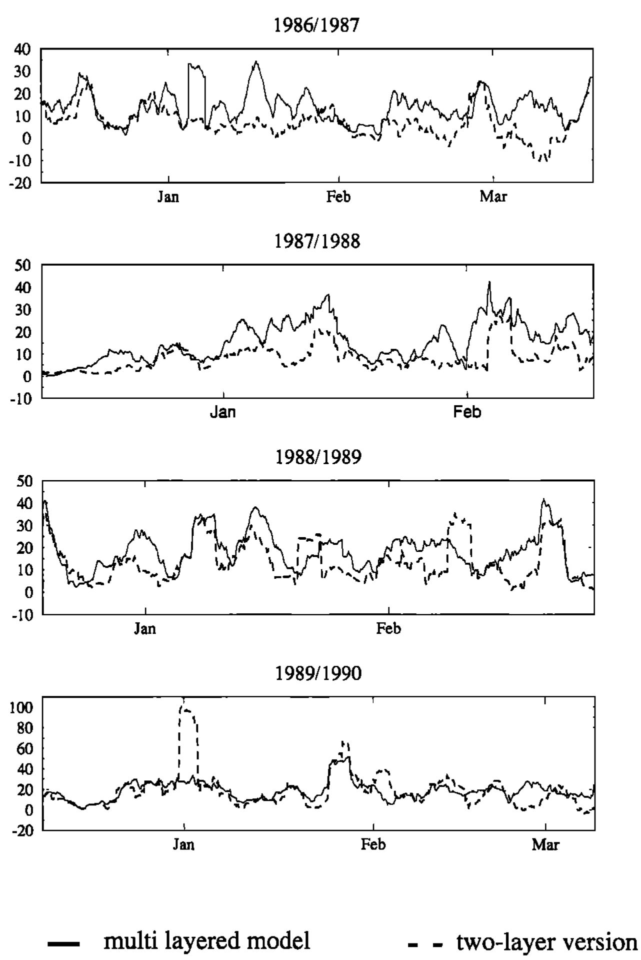

Figure 4b. Same as Figure 4a but for sensible heat flux (in W/m^2).

Table 1. Mean Daily Differences in Energy Fluxes Using a Two-layer version of a Snow-Cover Model and a Multi layered model [*Loth*, 1995] for the period 1986-1987 at Sodankylä (Finland) and Gander (Newfoundland).

Energy Flux	Absolute Differences, 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

Table 2. Parameterizations of the Snow Albedo in the Basic Version [*Loth,* 1995] and in Sensitivity Experiments

	Formula	Coefficients	Reference
Basic version Cold snow	$\alpha_s(t) = \alpha_s(t-1) + a_\alpha \Delta\tau$	$a_\alpha = -0.006/d$	*Gray and Landine* [1987]
Melting period,	$h_s > 25$ cm $\alpha_s(t) = a_\alpha + (\alpha_s(t-1) - a_\alpha)\exp(-b_\alpha \Delta\tau)$	$a_\alpha = 0.5$ $b_\alpha = 0.24$	*Verseghy* [1991]
Melting period,	$h_s \leq 25$ cm $\alpha_s(t) = \alpha_s(t-1) + a_\alpha \Delta\tau$	$a_\alpha = -0.071/d$	*Gray and Landine* [1987]
ECHAM	(temperature function) $\alpha_s(t) = a_\alpha - b_\alpha (T_{sf} - T_r)/(T_M - T_r)$	$a_\alpha = 0.8$ $b_\alpha = 0.4$ $T_r = 263.15$ K $T_M = 273.16$ K	*Roeckner et al.* [1992]
Linear ansatz	$\alpha_s(t) = \alpha_s(t-1) + a_{\alpha 1,2,3} \Delta\tau$ $a_{\alpha 1}$ for cold snow $a_{\alpha 2}$ for melting snow and $h_s > 25$ 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]

Here, $\alpha_s(t)$ and $\alpha_s(t-1)$ are the snow albedo at time t and $t-1$, respectively, $\Delta\tau$ is the time step of the model in days.

3.2. Albedo

The determination of the snow albedo, which is usually the key snow parameter in contemporary climate models, can be divided in four subareas: assessment of new snow albedo, increase in snow albedo in the case of slight snowfall, changes in albedo during dry aging and melting periods, and influence of the underlying soil in the case of a thin snow cover.

The sensitivity to the snow albedo is analyzed with the data set taken from the Col de Porte station (1988-1989). 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 Table 2.

Observed and simulated data for the snow albedo (Figure 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 unrealistically 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 the 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 (Figure 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 (Figure 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 K to 1.5 K. In spring (March to May) the simulated values are systematically smaller (and closer to the observations) compared to those of the basic version.

The energy balance changes systematically due to the modified parameterization of the albedo (Figure 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 (by -2 W/m^2 to -5 W/m^2 in the running average of 24 values) using the ECHAM temperature function for the albedo, the latent heat flux is increased by 5 W/m^2 to 7 W/m^2.

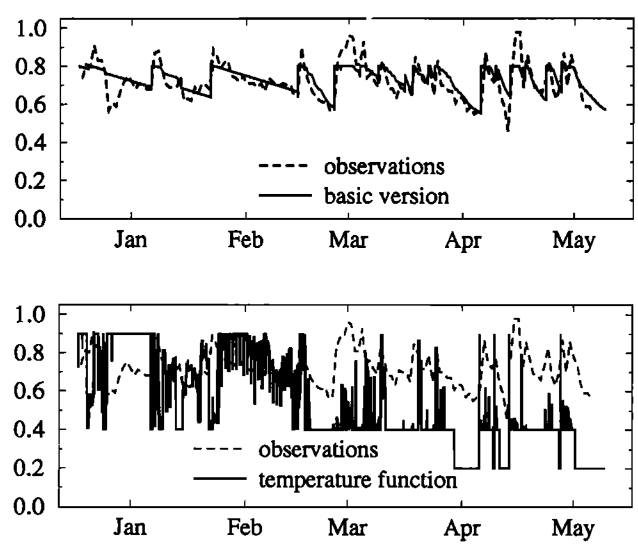

Figure 5. Measured and simulated snow albedo for Col de Porte during the time interval December 17, 1988, to May 8, 1989.The simulations are performed with a timedependent aging function and, alternatively, with a function of temperature for the albedo.

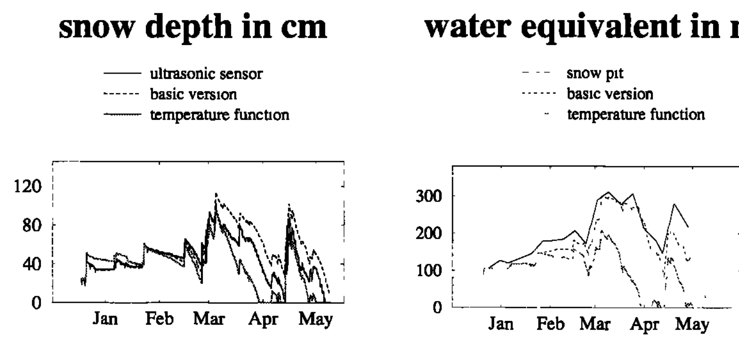

Figure 6. Measured and simulated snow depth and water equivalent for Col de Porte during the simulation period 1988-1989 using different parameterizations for the snow albedo.

3.3. Storage and Transmission of Liquid Water

The penetration of rainwater 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 is important for both the melting period and the time following the liquid-water events.

In contrast to Gander, at Col de Porte melting processes which directly influence the atmosphere only occur between March and May. High amounts of meltwater 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 and -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 Figure 9 for 1988-1989.

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 2 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 significance. 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 analyze 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 with no 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 modeled value of the snow depth and the measurements taken by the ultrasonic sensor (Figure 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,*

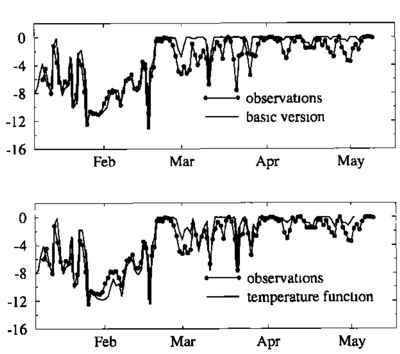

Figure 7. Daily mean values of the snow-surface temperature (in degrees Celsius) in Col de Porte for the time period January 8, 1989, to May 8, 1989, from observed data and simulated results using different parameterizations for the snow albedo (see Table 2).

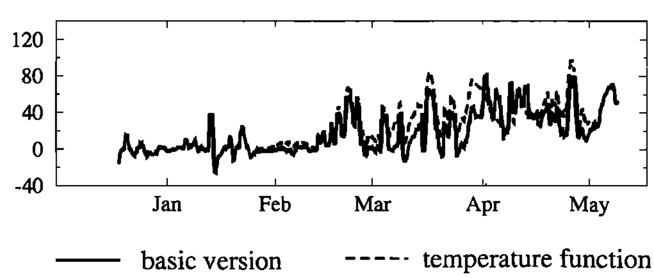

Figure 8. Simulated values of the energy balance of the snow cover (in W/m^2) using different parameterizations for the snow albedo. Shown are the running averages over 24 values for Col de Porte during the time interval December 17, 1988, to May 8, 1989

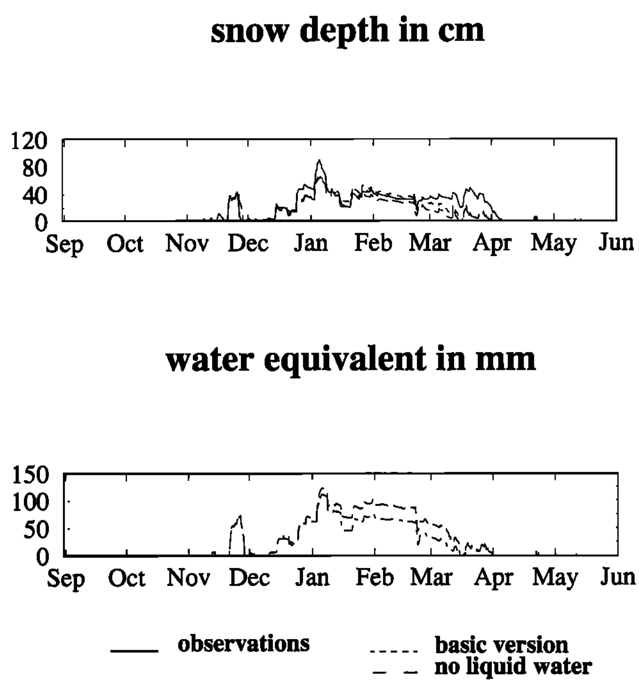

Figure 9. Snow depth and water equivalent at Gander for the simulation period 1988-1989. Shown are observed values of snow depth and simulation results using the basic version [*Loth*, 1995] and neglecting the storage of liquid water.

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. Because of 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 (Figure 11). This is not depen-dent 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 (Figure 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 (Figure 13). The reason for this model's behavior 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.

3.4. 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, the snow cover is very thin at Spitsbergen. At Zugspitze, with maximum snow depths between 300 cm and 500 cm, the ground properties can be important too. Owing 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 analyze the influence of the soil state (frozen or nonfrozen) 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/(mK). 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/(mK) and 0.9 W/(mK).

The higher thermal conductivity of the soil results in a more intense heat exchange between the snow cover and the underlying

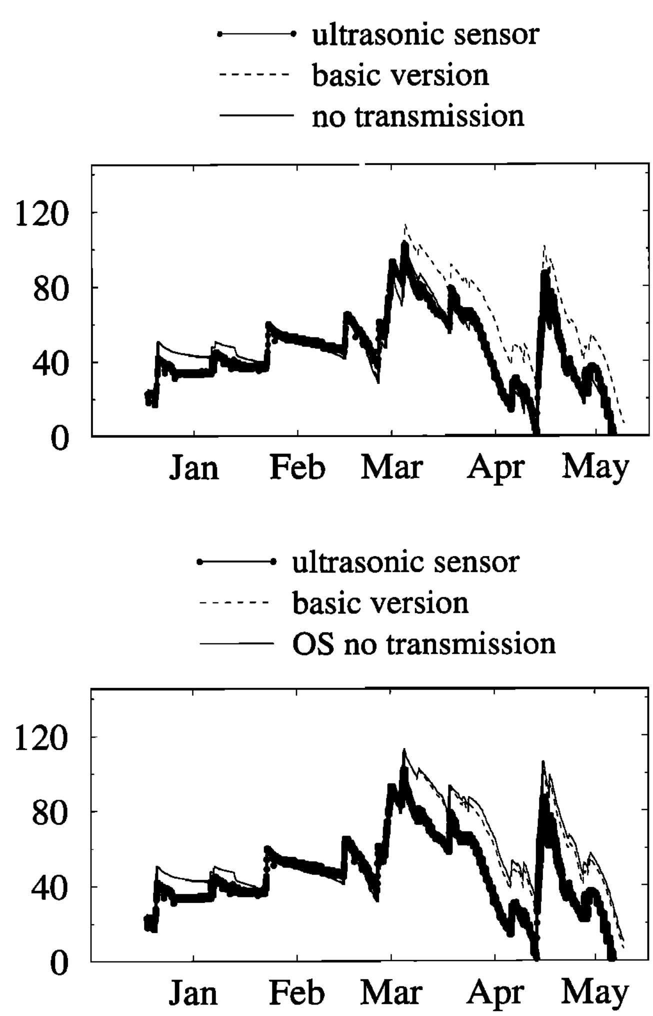

Figure 10a. Observed and simulated values of snow depth (in centimeters) for Col de Porte during 1988-1989 using different parameterizations for the transmission of liquid water within the snow cover. OS symbolizes the surface layer of the snow cover.

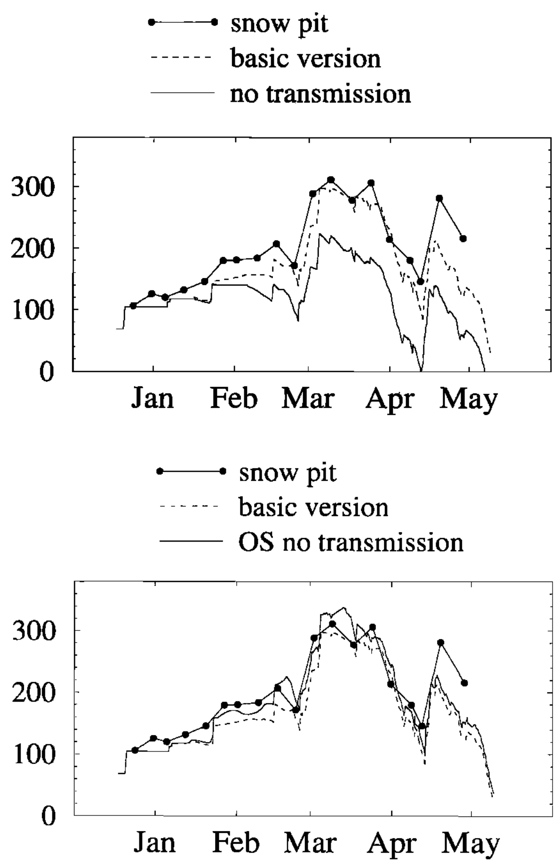

Figure 10b. Same as Figure 10a but for water equivalent (in millimeters).

soil. The changes in the energy fluxes at the snow surface are 2 and 4 W/m^2 for an increase in the thermal conductivity of the soil to 0.6 W/(mK) and to 0.9 W/(mK), 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. Figure 14 shows these results of the period 1988-1989. The increase in the thermal conductivity of the soil from 0.3 W/(mK) to 0.6 W/(mK) results in the earlier occurrence of the melting processes and an earlier end of 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.

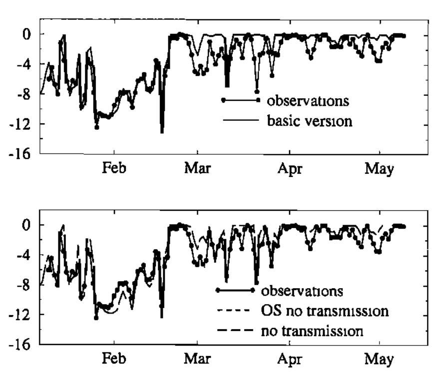

Figure 11. Observed and simulated mean daily values of the snow-surface temperature (in degrees Celsius) using different parameterizations for the transmission of liquid water at Col de Porte for January 8, 1989, to May 8, 1989.

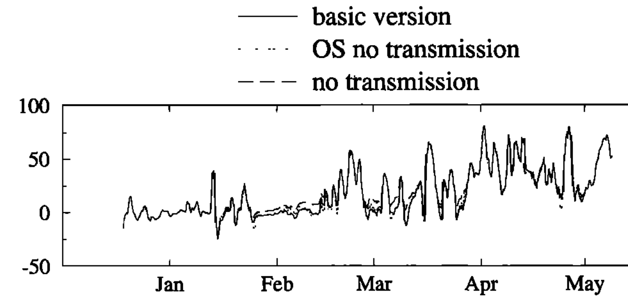

Figure 12. Energy balance of the snow cover (in W/m^2) dependent on the parameterization of the liquid-water transmission at Col de Porte during the time period December 17, 1988, to May 8, 1989 (running averages over 24 values).

At Zugspitze a higher soil temperature leads to systematic changes in the snow depth and in the water equivalent (Figure 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 2 weeks and 1 month. The mean ground heat flux increases from 3 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 (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 (Figures 14 and 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.

3.5. Internal Heat Conduction

Further studies were performed using different parameterizations for the internal thermal conductivity of snow (Table 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/(mK). Owing to the aging 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 (Figures 16 and 17) show the snow cover's strong dependence on the thermal conductivity of snow at midlatitudes (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 aging and melting periods. The snow depth changes between 2 cm and 10 cm for thin snow covers and between 100 cm and 150 cm 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

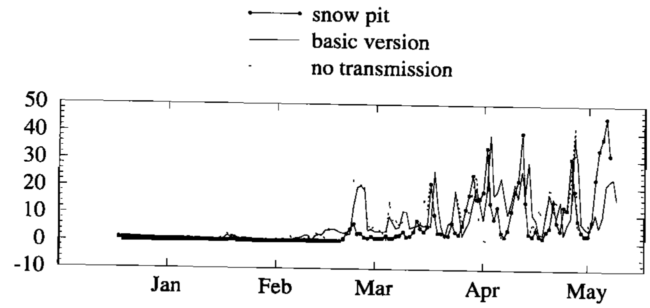

Figure 13. Mean daily values of the observed and simulated runoff (in millimeters) using different parameterization for the liquid-water transmission at Col de Porte during the time period December 17, 1988 to May 8, 1989.

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 result in changes in the turbulent fluxes (5 W/m^2 to 15 W/m^2). 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 midlatitudes.

At Zugspitze, snow mass differences also occur due to a change of the density-dependent 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/(mK) 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.

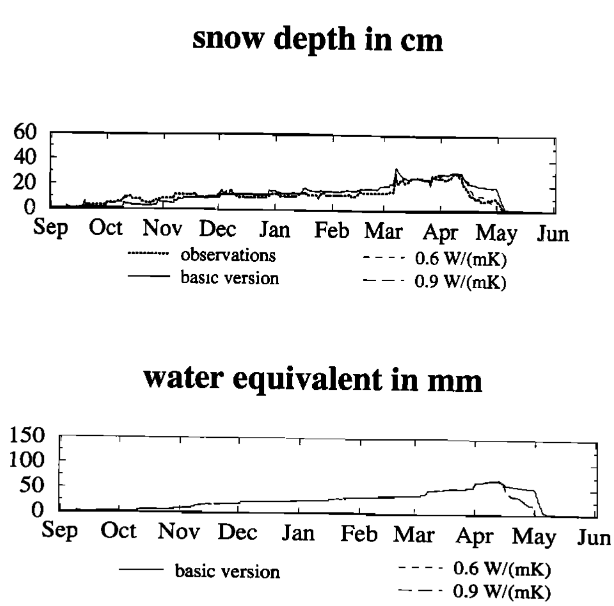

Figure 14. Simulated values of snow depth and water equivalent using different values of the thermal conductivity of the ground. Shown are the observed values of snow depth and simulation results using a thermal conductivity of soil of 0.3 W/(mK), 0.6 W/(mK) and,0.9 W/(mK) for Svalbard Lufthavn during 1988-1989.

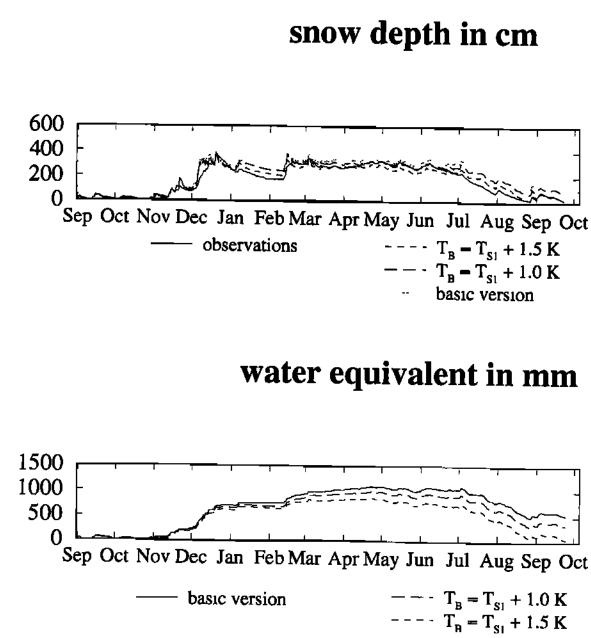

Figure 15. Snow depth and water equivalent at Zugspitze for 1988-1989 using different soil temperatures at 5 cm depth. Shown are observed snow depths and simulations of snow depth and water equivalent.

Table 3. Parameterization of the Thermal Conductivity of Snow λ_s in the basic version [*Loth*, 1995] and the Sensitivity Experiments

	Formula	Coefficients	Remark
Basic version *Anderson* [1976]	$\lambda_s = a_A + b_A\rho_s^2$	$a_A = 0.02$ $b_A = 2.5 \cdot 10^{-6}$	
Yen [1981]	$\lambda_s = a(\rho_s / \rho_w)^b$	$a = 2.22$ W/(mK) $b = 1.88$ $\rho_w = 1000$ kg/m^3	
	$\lambda_s = 0.176$ W/(mK)		constant value

Here, ρ_s is the snow density in kg/m^3.

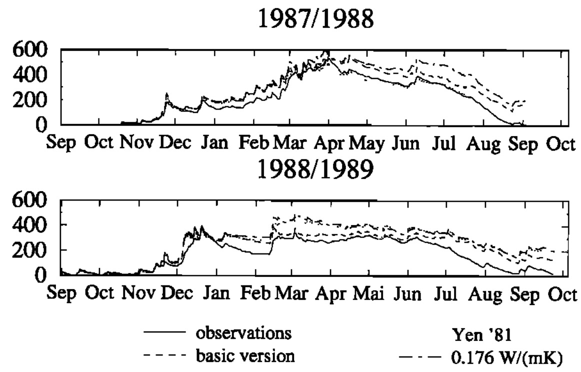

Figure 17. Simulated snow depth using different parameterizations of the thermal conductivity of snow. Shown are the observed values snow depth (in centimeters) at Zugspitze and simulation results using parameterizations from *Anderson* [1976], *Yen* [1981], and a constant value of 0.176 W/(mK).

4. 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 midlatitudes requires the resolution of the internal vertical gradients of temperature, density, and liquid water. Otherwise, the surface temperature is not adequately simulated, the runoff behavior 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 midlatitudes.

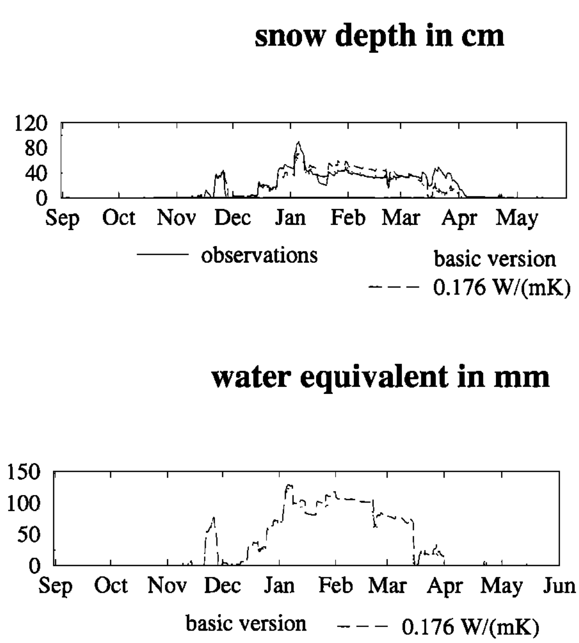

Figure 16. Snow depth and water equivalent at the midlatitude station Gander using different parameterizations of the thermal conductivity of snow for 1988-1989

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 K to 4.2 K (Col de Porte, 1988-1989). The end of the ablation period is shifted by approximately 8 days.

At midlatitudes the internal processes (liquid-water processes, heat conduction, aging) 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 midlatitudes 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 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 midlatitudes 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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