Assessment of the land surface scheme in climate models with focus on surface albedo and snow cover

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Publication Date:
1999

Permanent Link:
https://doi.org/10.3929/ethz-a-003813790

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Assessment of the land surface scheme in climate models with focus on surface albedo and snow cover

A dissertation submitted to the
SWISS FEDERAL INSTITUTE OF TECHNOLOGY ZURICH

for the degree of
DOCTOR OF NATURAL SCIENCES

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Zurich, 1999
Acknowledgement

I would like to express deep gratitude to Prof. A. Ohmura, the supervisor of this doctoral study programme, for his supervision and instruction during the studies. I could profit from his scientific thinking and deep understanding of the processes at the Earth's surface. I also appreciated his sense of humour and the friendly atmosphere he creates at the institute.

I am indebted to the encouragement of and discussions with Dr. Martin Wild who accompanied me as the head of the climate modeling group through this work. I greatly enjoy the good atmosphere and spirit we have in the climate modeling group, which also contributes Raelene Sheppard. Be it in the office or in a bar, I always enjoyed numerous stimulating discussions (not only on land surface schemes...).

The two co-examiners, Prof. C. Schär and Dr. D. Jacob gave me many constructive comments and suggestions for the final draft of this thesis. Their help and encouragement is greatly appreciated.

I would like to thank Prof. L. Bengtsson for the open doors at the Max-Planck-Institute and the possibility of frequently visiting his institute. I am grateful to Dr. J.-P. Schulz and Dr. D. Jacob who allowed me to tap their extensive scientific knowledge during my stays at Hamburg. Special thanks are due to Uli Schlese at DKRZ for his support in the technical aspects of the project, and for helping with all my questions regarding the model code.

The Swiss Centre for Scientific Computing in Manno made the high resolution simulations possible. Thanks in particular to Dr. Andrea Bernasconi and Dr. James Brunson for their technical support.

I am also grateful to many researchers from abroad: Dr. E. Heise, DWD in Offenbach, who kindly provided me with the source code of EM and hosted me in his own house during my research stay in Offenbach; Prof. M. Claussen, Potsdam-Institut für Klimafolgenforschung, for his support in vegetation-related problems; Dr. A. Schlosser, who sent me the data from a number of Russian sites. Dr. J. Pomeroy, National Hydrology Research Institute, Saskatoon, Canada and Dr. R.J. Harding, Institute of Hydrology in Wallingford, UK for valuable help concerning snow interception and the albedo of snow covered forests; Dr. A. Slater, Institute for Research in Environmental Science (CIRES) in Boulder, for the active E-mail contact about data sets and albedo of snow covered forests; and Dr. D. Verseghy, Climate Research Branch, Atmospheric Environment Service, Toronto, who gave me useful information on CLASS.

All the colleagues in our Geographical Institute are appreciated for their family-like atmosphere. Manfred Schwarb, Ludwig Zraggen and Marcel Müller shared not only the office, but also the wisdom, joys and hunger in solving scientific, computer-technical and human problems. Special thanks go to Dr. H. Gilgen who made valuable contributions to the paper and gave kind permission to access the GEBA data. I would like to thank Dr. G. Niederbäumer who provided data from the ETH Greenland expedition, Dr. G. Müller for BSRN data from Payerne, and Dr. D. Grebner for many discussions on the Indian monsoon and other climate relevant processes.

Many thanks go to Ms. Denise Scherrer from South Africa, who corrected my English and Ms. R. Sheppard from Australia, who helped me in numerous language problems and helped to improve my scientific English.

My wife, Brigitte Roesch, has given consistent assistance with this thesis. I would also like to give her my special thanks.
Abstract

This work assesses the land surface focusing on surface albedo and snow, along with their representation in numerical models.

Based on comprehensive surface datasets and literature, the work presents a number of improvements for the albedo- and snow parameterization. These algorithms are validated with ECHAM4 climate simulations at a T42 horizontal resolution.

The transformation of the prognostic snow water equivalent into the snow cover fraction was improved over both mountainous and flat areas. Validating the global snow cover fraction gave rise to a substantial overestimation in the snow cover fraction in mountainous areas, while the opposite has been found over flat areas by comparing remotely sensed and ground-based data. Long-term three dimensional ECHAM4 simulations showed that the relationship between snow water equivalent and snow cover fraction largely affects the surface climate in the Northern Hemisphere, including the Monsoon circulation. A model experiment, including the effects of subgrid scale variations in topography on surface temperature, snow water equivalent and surface albedo demonstrated that a significant deviation from the control climate primarily occurs over mountainous areas in autumn.

The incorporation of the BATS snow submodel in ECHAM showed that it is beneficial to introduce a prognostic snow-aging variable for simulating the snow albedo. However, it is crucial to correctly determine the parameter set in regions with minimal snowfall, such as Antarctica. The adoption of the submodel for snow albedo as used in CLASS, combined with a simple snow interception model demonstrated the ability to capture the main physical processes of snow covered canopies, including the albedo. The mean surface albedo over the boreal forests decreases by approximately 0.1 during winter and spring, which is in better agreement with ground-based observations. The enhanced solar absorption at the Earth's surface induces a significant rise in the surface temperature over extended parts of Eurasia and North America in late spring, which yields a faster snowmelt in spring and an accelerated retreat of the snow line.

The separate treatment of the solar radiation and surface albedo in the visible and near-infrared spectrum yielded changes in the total (radiation weighted) surface albedo by up to 0.05, primarily in regions with high seasonal variations in the precipitation and the precipitable water vapour. The inclusion of a reasonable relationship between the surface albedo and soil moisture demonstrated that neglecting direct soil moisture-albedo feedback fails to accurately simulate the annual amplitude of the surface albedo, primarily in regions with both rainy and dry seasons. The model's ability to simulate the snow-rain ratio could be considerably improved by reducing the threshold temperature where the melt of falling snow flakes is initialized.

The main results, which are based on off-line simulations and sensitivity tests using the land surface schemes incorporated in ECHAM and EM, yielded the following results.

The two land surface schemes produced similar results for surface temperature, net radiation and, to a lesser extent, for total water discharge and latent heat flux. Significant differences are found in simulations of the soil moisture content, surface runoff and drainage, as well as transpiration and bare soil evaporation, primarily due to differences in parameterization algorithms and the structure of the soil water model. The diurnal amplitudes of the ground heat fluxes are substantially overestimated in both models due to the reasonably thick first subsurface layer. The EM fails in the simulation of the transpiration rate in late spring due to the neglect of the biophysical control by the stomatal resistance of the canopy.

These sensitivity studies clearly demonstrated that it is of great importance to test the
sensitivities on annual, monthly and daily timescales. As the responses are nonlinear, it is necessary to extend the investigations to a broad parameter range. The impact on surface climate usually weakens with increasing roughness length, leaf area index and field capacity. The ECHAM land surface scheme generally shows a higher sensitivity with respect to the leaf area index and the roughness length than EM, primarily due to a different parameterization of the transpiration rate.
Zusammenfassung

Die vorliegende Arbeit befasst sich mit der Oberflächenalbedo und Schneebedeckung sowie deren Darstellung in numerischen Modellen.


Die Anwendung des Schneemodells von BATS zur Berechnung der Schneealterung und Schneealbedo hat sich als sehr günstig erwiesen. Dabei ist jedoch wichtig, dass die Parameter in Gebieten mit sehr geringem Schneefall, wie beispielsweise der Antarktis, richtig bestimmt werden. Die Einführung des Schneemodells von CLASS und eines einfachen Modells zur Bestimmung der Schneeinterzeption zeigt, dass damit die wesentlichen physikalischen Prozesse von schneebedeckten Wäldern gut beschrieben werden können. Die Albedo von schneebedeckten borealen Wäldern nimmt dabei im Winter und Frühling um rund 0.1 ab, was zu einer besseren Übereinstimmung mit Bodenbeobachtungen führt. Die erhöhte Absorption solarer Strahlung an der Erdoberfläche impliziert im Spätfrühling eine starke Erwärmung grosser Teile von Eurasien und Nordamerika. Dies wiederum lässt den Schnee im Frühjahr eher schmelzen, was zu einem beschleunigten Rückgang der Schneegrenze führt.

Eine getrennte Behandlung der Solarstrahlung sowie der Oberflächenalbedo im sichtbaren Bereich und nahen Infrarot erlaubt, besser auf die spektralen Eigenschaften der Albedo und des Strahlungsitransfers in der Atmosphäre einzugehen. Die (strahlungsgewichtete) Oberflächenalbedo ändert aufgrund dieser Aufteilung um bis zu 0.05, vor allem in Gebieten mit grossen saisonalen Schwankungen des Niederschlages und des atmosphärischen Wasserdampfes. Eine weitere Klimasimulation zeigt, dass die korrekte Behandlung des Einflusses der Bodenfeuchte auf die Oberflächenalbedo vor allem in Regionen mit einer Regen- und Trockenzeit von Bedeutung ist. Das Verhältnis von Schneefall zu Niederschlag stimmt deutlich besser mit Beobachtungen überein, wenn das Modell die Schmelze von Schneeflocken in der Atmosphäre bereits bei Temperaturen unter 2°C erlaubt.

Im folgenden werden die wichtigsten Resultate aus multidimensionalen Modellrechnungen sowie umfangreichen Sensitivitätsstudien mit den Landoberflächenschemata des ECHAM und EM präsentiert.

Die simulierte Oberflächentemperatur und die Nettostrahlung stimmen in beiden Landoberflächenschemata gut überein, während die Unterschiede im gesamten Abfluss und fühlbaren Wärmefluss doch deutlich grösser sind. Wesentliche Differenzen, die meist durch die unterschiedliche Struktur des Bodenmodells und der Parametrisierungsgleichungen entstanden sind, wurden bei den folgenden Grössen festgestellt: Bodenwassergehalt, Oberflächenabfluss, Drainage, Transpiration und Evaporation. Die Tagesamplitude des
Bodenwärmeflusses ist wegen der dicken obersten Bodenschicht in beiden Modellen deutlich überschätzt. Das EM hat vor allem im späten Frühjahr Probleme, die Transpirationsrate richtig zu modellieren, was hauptsächlich auf die fehlende biophysikalische Beschreibung des Stomatawiderstandes zurückzuführen ist.

Die Sensitivitätsstudie hat klar gezeigt, dass es von großer Bedeutung ist, die Sensitivität auf verschiedensten Zeitskalen (Tag, Monat und Jahr) zu testen. Zudem ist es wichtig, die Untersuchungen auf einen genügend grossen Bereich der Parameter auszudehnen, um nichtlineare Abhängigkeiten erkennen zu können. Der Einfluss auf das simulierbare bodennahe Klima nimmt üblicherweise mit zunehmender Rauhigkeit, Blattflächenindex und Feldkapazität ab. Das Landoberflächenschema des ECHAM zeigt häufig eine höhere Sensitivität bezüglich des Blattflächenindex und der Rauhigkeitslänge als jenes des EM, was in erster Linie durch eine stark unterschiedliche Parametrisierung der Transpiration hervorgerufen wird.
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1. Introduction

1.1 Objectives of this study

The land surface areas of the Earth represent significant sources, sinks and reservoirs of heat and moisture with respect to the atmosphere. The evaluation of land-atmosphere transfers of energy and water is therefore an important component of general circulation models (GCMs). The Earth's surface processes are also vital aspects in trace gas cycles. Therefore, the land surface schemes have received increasing attention in recent years.

The land surface schemes in GCMs differ substantially in structure as well as in the number and values of their parameters. Numerous investigations have demonstrated that the simulations of surface climate by GCMs are largely dependent on the formulation of their land surface schemes. Large uncertainties still exist, among other factors, mainly in (i) the parameterization of the components which contribute to total evaporation (i.e. transpiration and evaporation from bare soil) and the partitioning of the available net radiation into latent and sensible heat fluxes, (ii) the accurate representation of vegetation which covers at least half of the total land surface, and (iii) the approaches used for snow-covered situations.

It is, therefore, essential to carefully examine the biases in the simulated surface climate as generated by different land surface schemes. The most comprehensive project in the domain of comparing land-surface schemes is the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) (Henderson-Sellers et al., 1993) which was created in 1992 to provide a mechanism for validating land-surface schemes and to compare existing schemes.

Of key importance in any land surface scheme is the specification of the surface albedo, which influences the absorbed shortwave radiation. Modifications in the radiation balance induce changes in the surface temperature, in the turbulent heat fluxes and thus also in the soil moisture content. A number of studies have demonstrated that the surface albedo may affect the large-scale circulation (Kukla and Kukla, 1974; Laval, 1986). Therefore, a comprehensive understanding and realistic numerical modeling of surface albedo would represent a major step towards improving climate models.

The present work aims at assessing the parameterization of the surface albedo in the General Circulation Model ECHAM (ECHAM stands for European Center Model with Hamburg Physics) with focus on snow covered areas and compares the characteristics of the land surface schemes applied in ECHAM, the Europa-Modell (EM) and alternative model approaches.

The main focuses are as follows:

- to reveal the differences between the land surface schemes incorporated in ECHAM and EM
- to identify the processes responsible for the discrepancies between the ECHAM and EM simulations, with focus on the surface energy balance and the surface hydrology
- to analyse the sensitivities of the surface climate to key surface parameters
- to test available algorithms for an improvement and refinement of surface albedo parameterization
- to develop new parameterizations for improving the present parameterization of surface albedo
• to investigate the impact of snow on the surface albedo of vegetated areas
• to investigate a number of (snow-) albedo models within the framework of three-dimensional GCM simulations and to analyse their impact on the global climate.

The study is organized as follows: the first chapter provides a general introduction into the research field. The observational data sets are discussed in Chapter 2. In the following chapter, a detailed comparison between observed global climatologies and the control simulation of ECHAM4 is given. Problems arising from the comparison of surface albedos derived from GCM grids with in-situ observations are discussed in Roesch et al. (1999) (cf. Appendix).

In the fourth chapter, the characteristics of the land surface schemes used in both ECHAM and EM are given. The analyses are all based on off-line simulations and a detailed study of the isolated parameterization equations. These topics are mainly covered in the MPI-report No. 244 (Roesch et al., 1997) which is, in an improved and shortened form, presented in Chapter 4. This report provides, in the first section, a detailed description of the parameterization equations.

The main part of the thesis consists of validating different parameterization algorithms for the surface albedo, focussing on snow conditions, presented in Chapter 5. In addition to the analyses of isolated algorithms and their comparison with observations, the parameterizations were tested in the framework of long-term three-dimensional model simulations with the ECHAM4 GCM.

1.2 Previous work on sensitivity

In this section, a survey of studies on surface albedo and soil moisture is presented.

Surface albedo

Numerous studies have shown that climate models exhibit a significant sensitivity to surface albedo changes (Charney et al., 1977; Potter et al., 1981; Henderson-Sellers and Wilson, 1983). The first experiment, which showed a significant influence of the surface albedo on the climate conducted with a GCM was recorded by Charney et al. (1977). Hummel and Reck (1979) found a temperature change of 1 K for an alteration in global surface albedo by 0.01. In contrast, Hansen (1981) computed a four times weaker sensitivity of (global) surface temperature with respect to the (global) surface albedo. Reck (1993) found that, based on a 1-D climate model, a shift of 0.025 from the estimated global surface albedo of 0.13 may lead to an uncertainty of 2.5°C in the numerical simulation of global surface temperature.

The vegetation feedback (Claussen, 1997) increases the model’s sensitivity through a positive feedback: An increased albedo (caused by deforestation or overgrazing, thereby exposing more bare soil with higher albedo) yields less net radiation, which, in turn, lowers the surface temperature and stabilizes the lower atmosphere. This causes enhanced subsidence and, therefore, less precipitation which reinforces further vegetation loss (Cess, 1978). The albedo-vegetation feedback is excluded in ECHAM since the model physics does not allow a change in the vegetation mask. The snow/ice-albedo feedback intensifies the cooling due to increasing albedos and thus, less absorption of solar radiation: A higher albedo, and thus lower temperatures, leads to a preference of snowfall over rain and a delay in snow melt, thereby triggering a further increase in surface albedo. The snow/ice and
vegetation feedback suggests that the surface temperature of a 3-D GCM simulation will be more sensitive to surface albedo than in 1-D model runs.

These two positive feedbacks may be counterbalanced by negative feedbacks. The drying of the atmosphere due to an increased surface albedo (vegetation feedback) leads to a decreasing cloud amount which, in the case of low and middle clouds, leads to increased global radiation and hence, a positive temperature bias. 3-D GCM simulations, however, show that an increase in surface albedo will tend to reduce the mean global temperature (Manabe and Wetherald, 1975; Rowntree, 1988).

A number of studies have investigated the impact of surface albedo on evaporation and precipitation (Charney et al., 1977; Rowntree, 1988; Mylne and Rowntree, 1991; Garratt, 1993). The effect of surface albedo on evaporation strongly depends on the vegetation type and the climate zone, i.e. it is of major importance to distinguish between studies with extratropical, tropical or (semi-)arid climate.

Charney et al. (1977) found that, from a series of numerical simulation in semi-arid areas, increased albedo yields reduced evapotranspiration and lower precipitation. Increasing the albedo from 0.14 to 0.35 decreased the evaporation and precipitation by 0.8 mm/day and 2.0 mm/day, respectively. They argued that an increase in albedo acts to reduce the absorption of solar radiation by the ground and therefore the transfer of turbulent heat into the atmosphere. The resulting reduction in convective clouds tends to compensate for the increase of albedo by allowing more solar radiation to reach the ground, but it reduces the downward longwave radiation flux even more. Hence, increased albedo results in reduced net radiation and therefore a decrease of convective clouds and precipitation.

The model study of Sud and Fennessy (1985) depicted that the Indian summer monsoon was significantly weakened by an increase in surface albedo over the Indian region, which is in agreement with Charney’s hypothesis (Charney et al., 1977).

Mylne and Rowntree (1991) summarized model results for regions equatorwards of 20°. They found that evaporation from moist surfaces will be reduced by approximately 0.6 mm/day when the surface albedo increases by 0.1. This indicates that the climate response to variations in the surface albedo is in the same order for both moist and semi-arid regions.

Laval (1986) demonstrated the effect of albedo variations on atmospheric circulation which induced changes in the precipitation pattern.

Soil moisture

Many studies have shown that soil moisture has a large impact on temperature and precipitation (see e.g. Jacquemin and Noilhan, 1990; Garratt, 1993). Soil moisture also largely affects the vegetation cover and is thus crucial for human beings and animals. In the following, the main impacts of soil moisture variations on the climate system are summarized, starting with the moisture-rainfall feedback.

Eltahir (1998) suggests the following hypothesis for the moisture-rainfall feedback. Since wet soils reflect less radiation than dry soils, wet soils tend to enhance shortwave net radiation at the surface. A higher soil moisture content induces more evaporation at the expense of a reduced sensible heat flux, leading to (i) lower surface temperatures and (ii) higher water vapour content in the atmospheric boundary layer. This implies lower upward longwave radiation and more downward flux of terrestrial radiation (greenhouse effect). Hence, wet soils enhance net shortwave heating and reduce net longwave cooling. This increase in net radiation increases the total turbulent flux from the surface into the atmospheric boundary layer, thereby increasing the total energy in the atmospheric
boundary layer, which has been shown to result in an increase in convective instability and thus, a higher precipitation rate (Eltahir, 1998).

Recent budget analyses (Schär et al., 1999), through both conceptual models and numerical studies, demonstrated that, in addition to local evapotranspiration, the atmospheric transport of water plays a major role for the precipitation rate. Mintz (1984) reports that convective precipitation is strongly correlated with the surface evapotranspiration, while water vapour convergence is more important for large-scale precipitation events.

Numerous model studies investigated the impact of soil moisture variations. Mintz (1984) and Garratt (1993) reviewed the soil moisture sensitivity studies. They found the major effects of gross soil moisture changes on the simulated climate to be:

(i) For prescribed dry land surfaces (no evaporation, called 'dry soil case'), precipitation and evapotranspiration over Eurasia and North America are essentially zero. In tropical regions, however, precipitation amounts are maintained due to increased moisture convergence.

(ii) The dry soil case gives rise to far less cloudiness than the wet soil case (potential evaporation), resulting in larger global radiation. The increased solar heating, the enhanced sensible heat flux, and the elimination of the evaporative cooling makes the ground significantly warmer. This result contradicts, in part, the findings of Eltahir (1998) who suggests increased shortwave net radiation over wetter soil. The difference between the two concepts consists in the importance of cloudiness: The above mechanism emphasizes the reduction of cloudiness over dry soils, while Eltahir (1998) initially stresses the direct effect of the surface albedo. In reality, both effects play a major role in modifying shortwave net radiation. It is important to realize that both mechanisms lead to less cloudiness and precipitation over regions with drier soils.

(iii) Experiments with different initial soil moisture have shown that the anomalies in the soil moisture content require several months to return to normal conditions. This may be related to the positive feedback between soil moisture and precipitation (as outlined above).

1.3 Definition and measurement of surface albedo

Definition

The word albedo is derived from the Latin albus (meaning white) and is thus a measure of the whiteness or brightness of an object. The term was already used at the turn of the century. For instance, Russell (1916) defined the spherical albedo at wavelength $\lambda$ as the total energy reflected from the whole planet at wavelength $\lambda$ divided by the total energy incident at the planet at wavelength $\lambda$.

Albedo is used for different levels and wavelengths. Various types of albedo are defined below:

- The planetary (or system) albedo is the ratio of total solar energy reflected by a planet to solar energy incident upon it. Since albedos are generally dependent on both the wavelength and incidence angle of incoming solar radiation, it is of major importance to specify the spectral range to which the (integral) albedo is related.

- The surface albedo is the ratio of reflected to incident solar radiation at the Earth's surface. It is of utmost importance to indicate to which spectral range the surface albedo refers. In the ECHAM GCM, the surface albedo is the solar reflectivity integrated over a range between 0.25 $\mu$m and 4.0 $\mu$m. Unless otherwise specified the term 'surface albedo' is used in this sense throughout the whole thesis. A number of land surface schemes
distinguish between the spectral surface albedo in the visible range and the near-infrared. Unfortunately, the spectral range of these bands differ among the models and complicate the task of making direct comparisons.

**Problems in measuring the surface albedo**

Many measurements and climatologies of surface albedo exist (Leonard and Escher, 1968; Federer, 1968; Pinker et al., 1980; Kukla and Robinson, 1980; Henderson-Sellers and Hughes, 1982; Ohmura, 1982; Robinson and Kukla, 1984; and Robinson and Kukla, 1985). However, the successful compilation of the surface albedo data poses a severe experimental problem, as the albedo changes with locality and time. Temporal changes occur for two reasons: Fast changes are due to changes in irradiance and its spectral distribution, and slow changes are related to changes in reflection properties of the surface (e.g., snow cover). Since albedo usually varies with the incident zenith angle, it is insufficient to measure albedo once a day, even when assuming all other conditions to be fixed. The significant spectral variation in surface albedo requires measurements covering the whole spectrum of shortwave radiation. Generally, little effort is made to document the wavelength band for which measured data are appropriate, making the observation less than ideal. Since satellite-based measurements often cover only a narrow wavelength band, an attempt should be made to reconstruct the full solar spectrum albedo.

A further problem arises since the surface albedo relies heavily on the height at which the instrument measuring the reflected radiation is located. Ground-based observations measure the albedo for only small areas and, therefore, do not adequately characterize the albedo of extended areas.

An appropriate technique to determine the surface albedo of large inhomogeneous landscapes is based on the technique of remote sensing. Remote sensing observations collected by satellites deliver information from the top of atmosphere (TOA) albedo. To estimate the surface albedo, radiative transfer models of the atmosphere must be applied (e.g., Barker and Davies, 1989; Cess and Vulis, 1989; Li et al., 1993). The simplification and errors in the parameterization of the radiation transfer, as well as uncertainties in water vapour and aerosol amounts within the atmosphere, may lead to significant errors in derived surface albedo values. Additionally, sampling errors (random and systematic) in satellite measurements arise (Henderson-Sellers and Wilson, 1983). Verstraete et al. (1990) stress that satellite measurements strongly depend on both the position of the sun and the position of the observer relative to the sun since natural surfaces such as vegetation canopies reflect fairly anisotropically. Therefore, surface albedos, determined with remote sensing, are afflicted with rather large uncertainties.

An estimate of the likely error of approximately 11% in surface albedo retrieved from satellite data in the Sahel. However, since ground-based site measurements cannot provide the spatial coverage needed for worldwide validation of GCMs, global radiation monitoring must include satellite measurements.

Other measurement methods suffer from problems as follows: Tower measurements of shortwave radiation fluxes provide albedo values solely for a confined area and aircraft measurements are costly. While aircraft measurements have been used in different experiments, the interpretation of data raises many questions, e.g., fluctuations in the horizontal position of the aircraft may affect the accuracy of the measurements. However, deviations from the horizontal position hardly influence the measurements when an effort is made to fly at low altitudes and in overcast sky conditions. Tower measurements and in-situ observations are suitable for investigating the ground truth over small specified surfaces.
1.4 Main characteristics of surface albedo

Some aspects of surface albedo are detailed in the introductions to the analysed surface albedo parameterizations which have been evaluated in both off-line calculations and in long-term three-dimensional model simulations. A summary is given in the following section.

Parameters which affect the surface albedo

(i) Soil moisture
Since water surfaces generally have lower albedo in comparison with land surfaces, the albedo is dependent on the soil moisture content (Idso et al., 1975; Kondratyev, 1969; Gulf et al., 1995). Regions with bare soils exposed directly to the incoming radiation behave differently from vegetated regions, where the underlying soil is covered by vegetation. However, soil moisture content affects the state of the leaves which may, in turn, alter their reflectivity properties.

(ii) Roughness
Rougher surfaces generally have lower albedos since some of the radiation reflected upward by a rough surface strikes other parts of the surface, being further attenuated. Thus, surface albedo and roughness are usually inversely correlated.

(iii) Wavelength
The surface albedo of most surface types shows a distinct spectral variation (Briegleb and Ramanathan, 1982; Coulson and Reynolds, 1971). The surface albedo depends, consequently, on spectral scattering and absorption of solar radiation in the atmosphere. The presence of clouds also changes the spectral distribution of the incident solar radiation due to a depletion of the incident beam primarily in the regions of liquid and solid phase water absorption bands.

Soil surfaces are characterized by a monotonic increase in spectral albedo from 0.4 to 1.0 \( \mu \text{m} \). With further increasing wavelength, the spectral albedo varies weakly. Green vegetation, in contrast, has a rather specific spectral albedo: Leaf reflectance shows a sharp increase at 0.7 \( \mu \text{m} \) from approximately 0.1 to 0.5. Additionally, leaf albedo varies with leaf structure, age and in the presence of disease (Goudriaan, 1977). Snow cover has a maximum reflectivity in the 0.4 - 0.8 \( \mu \text{m} \) spectral region. At wavelengths exceeding 0.8 \( \mu \text{m} \), the snow albedo decreases sharply. Ice and snow have similar bulk optical properties in the visible and near-infrared spectrum (Barry, 1996). The spectral albedo over water surfaces varies slightly. However, the presence of particles suspended in water may substantially affect the spectral albedo.

(iv) Solar elevation
For most land surface types, the surface albedo varies with the zenith angle. The most pronounced variations are detected for snow, whereas forest albedo barely shows any dependence on the zenith angle due to the high degree of canopy roughness and consequent radiation trapping at all angles (Verseghy et al., 1993). Since the ratio of diffuse to total radiation generally increases with increasing cloud amount, the dependence on the zenith angle is more pronounced for clear sky conditions. Water albedo decreases strongly with increasing solar elevation. The corresponding relationship can be derived theoretically by using the Fresnel equation (Cogley, 1979). However, deviations from this ideal state occur due to the influence of wind, which changes the roughness, and water-mass turbidity.
Albedos classified according to land surface types

(i) Soil albedo
The soil albedo depends not only on the type of soil, but also on its colour, structure and soil moisture content. In general, higher albedos are associated with dry, sandy, smooth surfaces, whereas low albedos are associated with wet, organic, rough surfaces.

(ii) Snow albedo
The albedo of a snow cover is influenced by three factors: (1) the properties of the snow; (2) the zenith angle and spectral distribution of the shortwave radiation; and (3) the albedo of the underlying surface for thin snow packs.

(1) The snow albedo varies widely depending on its structure and state (see e.g. Kondratyev, 1969; Warren and Warren, 1980). The albedo of dry, clean, new snow is often in the range of 80% - 90% and decreases with time owing to the snow aging process which is caused by an increase in grain size and the accumulation of dirt and soot. During melt, the optical properties of snow are changed substantially due to melt-water and an enhanced liquid water content (which enables the growth of clusters of snow grains), thereby leading to considerably lower snow albedos (Sellers et al., 1996a).

(2) Snow albedo is generally high in the near-UV and visible range ($\lambda = 0.4 - 0.7 \mu m$), with the highest value between 0.4 $\mu m$ and 0.6 $\mu m$. A sophisticated model for the spectral albedo of snow is described in Warren and Warren (1980). The spectral albedo decreases sharply in the near-infrared spectrum (Barry, 1996; Henderson-Sellers et al., 1993). Snow albedo increases primarily where solar zenith angles exceed 60° (Lafleur et al., 1997).

(3) When the snow cover is optically-thin, shortwave radiation is transmitted through the snow pack and, as a consequence, absorption by the underlying surface occurs. Furthermore, thin snow packs are usually no longer closed but are rather patchy, primarily during melting conditions, which allows for some part of the vegetation to emerge. This complexity is discussed further in Section 5.1.

(iii) Albedo of forests and vegetated areas
The total albedo of forests and vegetated areas is determined by (1) the spectral composition of the incident solar radiation, (2) the zenith angle of the incident solar beam, (3) the optical properties of individual leaves (e.g., moisture content in leaves), (4) the structure of the canopy (height, crown size of forests), and (5) the orientation of the plant’s leaves. Reviews of various aspects are presented in Ross (1975), Goudriaan (1977) and Ross (1981).

Points (3) and (4) imply that the albedo of leaf canopies (typically 0.1 - 0.15) differs remarkably from that of single leaves (albedo: $\sim$0.3) due to the process of radiation trapping: much of the light reflected from leaves situated below the top layer of the canopy is shaded by other leaves and so becomes further attenuated. Since radiation trapped by multiple reflection is generally at a maximum for (direct) radiation coming from overhead, the albedo values are often at a minimum around noon (Stewart, 1971; Rauner, 1976; Jarvis et al., 1976). However, Verseghy et al. (1993) emphasize that for trees, the diurnal variation of the total albedo is small whereas the albedo of crops and grass is largely dependent on the zenith angle.

For the albedo properties of a snow-covered canopy, experiment EXP7 (Section 5.7) offers considerable information.

(iv) Sea surface albedo
The sea surface albedo depends on several factors: solar elevation, cloud amount, roughness of the water surface, water contamination, and characteristics of the water basins such
as depth and transparency. The findings of Cogley (1979) suggest that the water albedo increases considerably with zenith angle. The albedo of agitated water differs markedly from that of a calm surface.

The value of the sea ice albedo is detailed in Barry (1996) and Allison (1993). The surface albedo of sea-ice depends primarily on ice thickness, puddle coverage, snow cover and the distribution of open water in leads and polynyas (areas of open water and thin ice, surrounded by thick sea ice and by land).

1.5 The importance of albedo in the climate system

Planetary and surface albedo greatly affects the Earth’s climate by determining the energy balance of the entire Earth/atmosphere system and at the surface, respectively. The simplest possible measure of global climatic change is the effective temperature $T_e$. Assuming the Earth, on average, to be in global radiative equilibrium, Equation 1.1, which incorporates the planetary albedo $A_p$, applies:

$$\frac{S}{4}(1 - A_p) = \sigma \cdot T_e^4,$$

where $S$ is the solar constant and $\sigma$ is Stefan-Boltzmann’s constant. It is hence obvious that the temperature of the Earth-atmosphere system is quite sensitive to small changes in its energy balance, such as large-scale changes in surface albedos.

Relatively large surface albedo changes may result from human activities: desertification, salinization, deforestation and urbanization have a significant effect on surface albedo. Sagan et al. (1979) suggest that the albedo changes in the period 1954 - 1979 were capable of depressing global temperature by 0.2°C.

Many simple climate models analyse changes in the climatic regime as a function of albedo variations. For instance, the snow/ice albedo feedback can be crudely investigated by considering the large difference between the snow/ice albedo and albedo of bare soils. Increasing snow cover substantially reduces the surface net radiation and thus yields further cooling which increases the ratio between snowfall and rain. The high snow albedo also contributes greatly to thermally induced anticyclones which develop over vast snow covered regions such as Russia.

The albedo/vegetation feedback is mainly based on significant differences between the albedo values of dry/bare and vegetated soils: Increased albedo due to overgrazing (e.g. in the Sahel zone) results in a net radiative loss which produces general subsidence and drying over the area, thereby inhibiting or reducing the convections necessary for rain. Deforestation experiments are often conducted by assuming that, in a first approximation, clearing forests solely results in a raised surface albedo.

A change in the surface albedo has further consequences. Glaciation may partly result in albedo variations over geological time-scales. Ash deposited by explosive volcanic eruptions might trigger glaciation due to increased planetary albedo. Over very long time periods, effects of changing land-sea distribution due to sea level fluctuations may also substantially change the global (surface) albedo and thus the net radiation balance owing to a very low ocean albedo.

It should be pointed out that many more examples could be added to the above-mentioned list where the albedo plays a major role in climate change.
2. Observational data

Observational data on a global and regional scale are used for model validation. Only when there is a high degree of confidence that land surface datasets, such as snow cover and snow depth, are reliable the observational data can be used to evaluate the performance of GCMs (Foster et al., 1996). However, some of the data sources are satellite-derived quantities, e.g. the global distribution of surface albedo, and should be handled with special care.

2.1 Global data

Snow cover

For the validation of the snow water equivalent, the global snow depth climatology of the U.S. Air Force Environmental Technical Application Center (USAF/ETAC) as documented in Foster and Davy (1988) is used. This dataset provides a mid-monthly mean snow depth climatology with the highest spatial resolution (1 x 1° equal-angle grid) currently available, using a comprehensive set of station data for the months of September through to June. The USAF data is generally considered as being one of the most reliable and accurate snow depth climatologies available (Douville et al., 1995a) and is used in several studies dealing with the validation of snow models (Douville et al., 1995a; Marshall et al., 1994; Foster et al., 1996). Over the United States, Canada and Eurasia, there is high confidence in the observations, as they generally contain more than 5 years of data and are of good coverage. In areas of sparse data coverage, snow depths are estimated using precipitation and satellite analyses of snow extent. These areas are generally assumed to have a low confidence level (e.g., Arctic and Antarctica).

For model comparison, snow depth $h_s$ was transformed to snow water equivalent $S_n$ (in millimetres) according to Verseghy (1991) as snow depth is not simulated in ECHAM:

$$
\rho_s = 188.82 \; kgm^{-3} + 0.419 \; kgm^{-2} \cdot S_n \leq 450 \; kgm^{-3}, \tag{2.1}
$$

where $\rho_s$ is the density of snow and $S_n$ the snow water equivalent. Following Verseghy (1991), snow density does not exceed 450 kgm$^{-3}$ but rather remains constant for $S_n \geq 0.623 \; m$. The snow cover density changes not only by mechanical compaction but also by metamorphism. Temperature induced metamorphism leads to significant density changes to the end of the snow season. Therefore, Eq. 2.1 probably underestimates the snow density in spring since the density increases as the snow pack melts. However, Eq. 2.1 is reasonably well confirmed by using the observed values of snow depth and snow water equivalent at six Russian stations (cf. Section 2.2).

For the global snow cover fraction the weekly values of “weekly digital Northern Hemisphere snow and ice product” compiled by the National Oceanic and Atmospheric Administration (NOAA) and National Environmental Satellite, Data and Information Service (NESDIS) has been used. The Very High Resolution Radiometer (VHRR) launched in 1972 provided imagery with a spatial resolution of 1.0 km. Since November 1978, the Advanced Very High Resolution Radiometer (AVHRR) has provided 1.1 km-resolution data. NOAA charts are based on a visual interpretation of photographic copies of visible satellite imagery by trained meteorologists. Only cells interpreted to be at least 50% snow covered are considered as being snow covered. In general, the NOAA charts are considered to be the most accurate means of obtaining snow extent information on large regional to hemisphere scales. Furthermore, they comprise the longest satellite-based record available.
The principal shortcomings in using shortwave data to chart snow cover are (i) problems when solar radiation is small, (ii) cloudy skies, (iii) snow masking of forests. Moreover, problems arise when the snow cover is unstable or changes rapidly. The dataset has been intensively used in former studies (Gutzler and Rosen, 1992; Iwasaki, 1991; Kukla and Robinson, 1981; Masuda et al., 1993; Robinson et al., 1993).

Surface radiation and albedo

The Surface Radiation Budget (SRB) dataset provides short- and longwave radiation fluxes and thus albedo at the surface. Input data, which cover the period 1984 - 1990, come from the International Satellite Cloud Climatology Project (ISCCP) (Schiffer and Rossow, 1983), which provides diurnal cloud cover data, and from the Earth Radiation Budget Experiment (ERBE) (Barkstrom et al., 1989) which provides broadband measurements of longwave fluxes and albedos at the top of atmosphere (TOA). The SRB data are given on the ISCCP equal-area grid which comprises 6596 gridboxes. Close to the equator, its resolution is 2.5° x 2.5°.

In order to derive surface radiation fluxes and surface albedo from top-of-atmosphere (TOA) radiation fluxes, two different algorithms were developed. The Pinker algorithm is a physical model that uses an interpolative procedure based on delta-Eddington radiative transfer calculations. It uses scaled radiance, cloud amount, precipitable water and ozone parameters for input. A theoretical description of the method is given in Pinker and Laszlo (1992). The second algorithm was developed by Staylor (Darnell et al., 1992) at the NASA Langley Research Center. This algorithm is a parameterized physical model that uses ISCCP-scaled radiances, cloud amount, precipitable water and climatological aerosols to determine the clear atmospheric and cloud transmission characteristics. The surface albedos used in this study are based on the Staylor algorithm since which computes all-sky albedos. The Pinker algorithm, in contrast, generates clear-sky albedos. Further, the comparison of the surface albedo with ground measurements on Greenland, Antarctica and the Amazonian rain forest indicate that the ERBE-derived Staylor algorithm estimates are more accurate (Whitlock et al., 1995). The same paper emphasizes that shortwave surface parameters are not expected to be accurate over either snow/ice or very bright desert surfaces.

The Global Energy Balance Archive (GEBA) has been developed at the Swiss Federal Institute of Technology. This database contains worldwide measured energy fluxes at the Earth’s surface (Ohmura et al., 1989; Ohmura et al., 1991; Gilgen et al., 1997). A quality control procedure has been applied to the data which includes a “physically possible” check and an additional quality control based on empirically derived relationships between cloud cover and irradiances. The shortwave radiation is measured with thermoelectric pyranometers in the wavelength range between 0.3 and 2.8 μm. The random error in the shortwave fluxes due to the calibration procedure and the pyranometer window is estimated to be less than 1% (Gilgen et al., 1998). Currently, the database possesses 220’000 monthly mean data entries for approximately 1600 sites. GEBA has been used in a number of studies focusing on the assessment of the energy balance at the Earth’s surface (Ohmura and Gilgen, 1993; Garratt, 1994; Li et al., 1995; Rossow and Zhang, 1995; Whitlock et al., 1995; Wild et al., 1995; Garratt and Prata, 1996; Konzelmann et al., 1996).
Air temperature

The global distribution of surface air temperature has been compared with the dataset of Legates and Willmott (1990a). The dataset which served as a basis for the interpolation, consists of 17,986 terrestrial station records and 6,955 oceanic records. Most of the data were compiled between 1920 and 1980. The surface air temperatures are interpolated to a 0.5° of latitude by 0.5° of longitude resolution. Besides station inhomogeneities due to station movement, changes in instruments and exposure, and the variable thermometer heights can lead to discrepancies. Oceanic measurements of surface temperature are usually taken at a shipboard height of twelve meters (Woodruff et al., 1987), whereas land measurements refer to a height of about one to two meters above the land surface.

Precipitation

The precipitation data have been compiled by Legates and Willmott (1990b) and comprises 24,635 terrestrial station records and 2,223 oceanic grid-point records. However, most of the stations are located in the countries of North America, Europe and East Asia. All the monthly precipitation data have been corrected for influence of wind and are interpolated to a 0.5° of latitude by 0.5° of longitude grid. Most of the data were observed between 1920 - 1980, and therefore the climatology is largely representative of that 60-year period.

In addition, the GPCP precipitation distribution was used for precipitation comparison. The Global Precipitation Climatology Project (GPCP) was established to provide global datasets of area-averaged and time-integrated precipitation based on all suitable observation (Rudolf et al., 1996). This global analysis contains precipitation estimates on a 2.5° x 2.5° mesh and is based on conventional and satellite measurements. The precipitation from gauge stations (approximately 6,700) have been corrected with regard to the systematic error from evaporation losses and snow drift caused by wind, using simple correction factors according to Legates (1987). Over oceans, satellite observations of radiances were used. In order to transform radiances to rainfall, algorithms that relate radiances from clouds or from raindrops to rainfall were applied.

Re-analysis

Both the U.S. National Centers for Environmental Prediction (NCEP), cooperating with the National Center for Atmospheric Research (NCAR), and the European Center for Medium-range Weather Forecast (ECMWF) have carried out independent re-analysis projects, with more than a terabyte of data generated. This massive atmospheric data archive comprises a 40 year record (1957 - 1996) for the NCEP data and analyzed data from 1979 - 1993 for the ECMWF re-analysis. During the ECMWF re-analysis project all forecast cycles with the ECMWF model over the period 1979 - 1993 have been repeated to establish a consistent picture of the state of the atmosphere over a period of 15 years. The model version that was used incorporates the code described in ECMWF (1996). Details of the technical procedure are given in Majewski (1985). The NCEP re-analysis (Kalnay et al., 1996) is one of the most important observational datasets that is available for studying the three-dimensional variability of standard atmospheric variables. In this study, both the snow water equivalent and the soil moisture content are retrieved from the NCEP re-analysis and represent, therefore, 40-year-averages from 1957 - 1996.
2.2 In-situ snow cover observation

The data of six Russian stations (Table 2.1) for the period 1978 - 1983 were chosen as one of the principle sources for regional model validation. All sites are within ten degrees of latitude, but are widely spaced longitudinally. This high quality dataset is described in detail in Robock et al. (1995) and is often used for the validation of snow models (Douville et al., 1995b; Yang et al., 1997). Each of these stations is located on a plot with natural vegetation (grass). The grass covering was allowed to grow uninhibited and was rarely cut (Slater et al., 1998). Snow cover measurements were made on the 8th, 18th and 28th day of each month during the winter. Meteorological data (air temperature, soil (snow) surface temperature, precipitation, wind speed, air pressure, lower cloud cover fraction and total cloud cover fraction) for the period 1978 - 1983 were measured regularly 8 times a day at the same time (Moscow legal time, Greenwich time plus three hours) for all six stations. The actinometric data (incoming solar radiation, net radiation and surface albedo) were taken 6 times per day (12:30 a.m., 6:30 a.m., 9:30 a.m., 12:30 p.m., 3:30 p.m., and 6:30 p.m.) at the mean local solar time.

Table 2.1: Russian stations

<table>
<thead>
<tr>
<th>station</th>
<th>latitude (-° S)</th>
<th>longitude (-° W)</th>
<th>altitude (mamsl.)</th>
<th>period of measurement</th>
<th>surface type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kostroma</td>
<td>57.8</td>
<td>41.0</td>
<td>464</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
<tr>
<td>Yershov</td>
<td>51.4</td>
<td>48.3</td>
<td>104</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
<tr>
<td>Uralsk</td>
<td>51.3</td>
<td>51.4</td>
<td>82</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
<tr>
<td>Ogurtsovo</td>
<td>54.9</td>
<td>83.0</td>
<td>74</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
<tr>
<td>Tulun</td>
<td>54.6</td>
<td>100.6</td>
<td>131</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
<tr>
<td>Khabarovsky</td>
<td>48.5</td>
<td>135.2</td>
<td>0</td>
<td>1978 - 1983</td>
<td>grass</td>
</tr>
</tbody>
</table>
3. Validation of ECHAM4

3.1 Introduction

The principal goal of this chapter is to analyze the model's climate and to reveal the main errors by comparing the simulated surface climate with observations.

The comparison is limited to the global fields which are of interest in the following discussions of the ECHAM4 simulations, that is precipitation, temperature, surface albedo and snow water equivalent. For a detailed assessment of snow cover characteristics the reader is referred to Section 5.1. A thorough validation of the ECHAM models can be found in Roeckner et al. (1992) for ECHAM3 and in Roeckner et al. (1996) for ECHAM4.

The following comparison is based on the ECHAM4/T42 simulation of present day climate ("control run") with prescribed annual mean cycle of sea surface temperature (SST) and sea ice coverage from the atmospheric intercomparison project (AMIP) (Gates, 1992) which comprises data from the period 1979 - 1988. Some of the results are also founded on the ten year simulation of the present climate with ECHAM4/T106. However, the principle features remain the same using either T42 or T106 resolution, respectively: Significant differences are primarily detected on a regional scale.

ECHAM model

The ECHAM GCM model evolved from the spectral numerical weather forecasting model of the European Centre for Medium-Range Weather Forecasts and has been modified extensively for climate applications at the Max Planck Institute (MPI), Hamburg. This model is described in detail in Roeckner et al. (1992) for the ECHAM3 version and Roeckner et al. (1996) for ECHAM4.

The land surface scheme of ECHAM4 is described in detail in Section 4.2.

3.2 Precipitation

Precipitation directly influences the soil water content, the runoff and the latent heat flux. Additionally, the atmospheric circulation is influenced by the release of latent heat. However, a completely reliable spatial representation of global precipitation remains unavailable (Willmott et al., 1985). Over oceans, the mean precipitation at any point may be uncertain by a factor of two (Dickinson, 1992) due to huge spatial data gaps. Land areas are heavily instrumented with rain gauges in some regions and hence, mean values of precipitation are known within a few tens of percent (Dickinson, 1992). In mountainous regions the uncertainty in the mean precipitation is more significant because of the large spatial variability of precipitation and the problems due to snowfall, particularly at high wind speeds.

As an indication of the uncertainties of present-day precipitation climatologies, two global precipitation distributions are used for comparison. Fig. 3.1 shows the zonally averaged precipitation of the ECHAM4/T42 control experiment, Legates and Willmott (1990b) climatology (LW90 hereinafter) and the data obtained within the Global Precipitation Climatology Project (GPCP) (cf. Section 2). There is generally good agreement between the observation and T42 in the Northern Hemisphere, whereas the Southern Hemisphere, with its large ocean areas, shows distinctly more scatter. While the climatologies agree
reasonably well over land (Fig. 3.2), there are large discrepancies over oceans. Whereas LW90 gives a large peak at 60°S during DJF, the GPCP climatology and ECHAM4 show an average of only about 50 mm/month (Fig. 3.1). This substantial difference between both climatologies at these latitudes stresses the low reliability of the observations over oceans. The observed double maximum in the tropics during DJF is, in ECHAM, distinctly shifted to the south. In JJA the tropical maximum is similar to LW90 (shifted somewhat to the north) whereas the GPCP climatology is significantly lower. The difference of nearly 50 mm/month is mainly due to different analyzed precipitation amounts over the oceans. In the latitude belt between 0 and 30°S the model is close to the GPCP distribution whereas LW90 shows more precipitation in these regions. Again, these uncertainties may not be significant due to large data gaps in ocean measurements. During JJA, ECHAM4 appears to overestimate precipitation between 30° and 60°S. Some of the differences may be due to different resolutions of the datasets used in this study.
A comparison of the simulated geographical distribution of seasonal mean precipitation with LW90 is shown in Fig. 3.3. In DJF, the simulated precipitation is smaller than in LW90 over large parts of the Southern Hemisphere oceans. Over the North Hemisphere oceans the bias is also negative, but is smaller than over the SH oceans. However, over the Indian Ocean close to the Equator, a huge area shows excessive model precipitation compared to both analyses. No systematic errors are found over land area of the Northern Hemisphere. Precipitation appears to be too high in the dry regions of Australia, South Africa and extended parts of South America excluding the Brazilian rain forests. This also applies to the GPCP climatology (not shown). The strong maxima in the equatorial Pacific close to the intertropical convergence zone (ITCZ) in LW90 has not been found in either the GPCP global precipitation field and the ECHAM4 simulation and is expected to be unrealistic. It has to be emphasized that LW90, the GPCP data and ECHAM4/T42
have different spatial resolution and thus the gridboxes of the global datasets do not coincide. This yields artificial differences between the simulated and observed distribution and may be partly the reason for the differences.

An excessive model precipitation is found during DJF over the Rocky Mountains and Andes. This is related to either poor analyses or wrong circulation patterns. During DJF, the model underestimates precipitation over the Mediterranean Sea while large parts of central, eastern and northern Europe are too wet.

During JJA (Fig. 3.3b), less precipitation is simulated than suggested by LW90 throughout the tropics, except in the eastern equatorial Indian Ocean. In the mid-latitudes of the Southern Hemisphere, ECHAM4 simulates distinctly more precipitation than suggested by LW90, which is different to the lack observed in DJF in these regions. While the lack of precipitation over the SH mid-latitudes in DJF is not confirmed by the GPCP analysis, the positive deviation of precipitation in JJA can be seen compared to both climatologies. Over the Brazilian rain forest there is a lack of precipitation throughout the entire year. However, the reliability of the analysis is very low due to the small number of available measurements over rain forest domains. Over North America, the difference pattern is rather patchy. More systematic are the deviations from LW90 over Eurasia where the summer precipitation is generally too low in the simulation. During the European summer, the model produces less precipitation than is observed due to the summer drying of soils (Wild et al., 1996). In these territories, the model does not reproduce the correct annual cycle with more precipitation in summer than in winter.

A definite weakness of ECHAM4 is the poor simulation of the summer monsoon over India and South East Asia (Roeckner et al., 1996). While the model tends to produce a precipitation maxima along the equatorial Indian Ocean, the analysis shows a maxima further north over India and the Bay of Bengal. These characteristics lead to a lack of precipitation over the equatorial Indian Ocean whereas the adjacent land masses are too dry. The huge dry tongue-like areas on the western side of the continents due to cold ocean currents and therefore subsidence influence, look very similar in both the model and the GPCP-analysis. On the contrary, LW90 displays far smaller regions with low precipitation (not shown). This finding may signify that the satellite-derived dataset (GPCP) is more reliable over the ocean than the interpolation of LW90 with poor data coverage and therefore the GPCP-analysis should be preferred over ocean areas. Model simulated precipitation in the lee of large mountain chains is often higher than in the climatologies. This may be influenced by too low mountains in ECHAM/T42 which allows the air, flown over the mountain chain, to contain too much water. This feature seems to be most pronounced in the lee of the Rocky Mountains and the Himalayas. During both summer and winter, ECHAM4 underestimates the precipitation rate near the slopes of Antarctica. This may be due to an underestimation of orographic precipitation at steep slopes caused by poor orographic resolution. However, the same pattern is found using the ECHAM4/T106 control experiment (not shown).

### 3.3 Surface temperature

Surface temperature is one of the most important climatic elements. It is an important attribute of climate and also very widely available. Fig. 3.4 displays the climatological differences between the surface temperatures of ECHAM4/T42 and LW90 for both the summer (JJA) and winter season (DJF). The error patterns over land are less spatially coherent than over the oceans. This is mainly related to the variations in the topography over land which are absent over sea. The effect of different resolution (ECHAM4/T42: \( \sim 2.8° \), LW90: 0.5°) may also play a minor role. In Northern Hemisphere (NH) winter
(DJF), surface temperatures are far too cold in the Antarctica and extended areas of the Arctic regions such as Greenland, the northern parts of Canada and Siberia. The cold bias often exceeds three degrees and amounts to about five degrees in both above-mentioned regions. However, it should be emphasized that LW90 used a very poor interpolation method for temperature over mountainous regions which leads to large errors in the temperature. Comparing the temperature map displayed in Ohmura (1987) for Greenland with LW90 reveals a difference in the mean annual temperature of more than nine degrees. In fact, Ohmura et al. (1996a) report that in the case of air temperature, ECHAM3 performs reasonably well. Merely minor overestimation has been detected in the interior of Greenland. However, a more comprehensive evaluation based on the map of annual mean temperature over Greenland displayed in Ohmura (1987) reveals a noteworthy relationship between observed and modeled surface temperature (Fig. 3.5): For low temperatures the model
significantly overestimates the observation while areas close to the coast line are slightly too cold. The simulated annual mean temperature over the whole Greenland ice sheet is approximately -20.3°C while the estimate derived from LW90 is -11°C. This is mainly caused by the incorrect interpolation method applied in LW90 which overrepresents the low-lying stations close to the coast line. Similar problems occur over the Antarctic ice sheet where the model produces a small overestimation for the Antarctica Peninsula and Ross and Filchner Ice Shelves when compared with data from Giovinetto et al. (1990). Summarized, LW90 temperature data are useless for describing the temperature distribution over the Greenland and Antarctica ice sheets.

The Sahara and Arabian desert are substantially too cold in DJF. This cold bias is less pronounced in NH spring and autumn (not shown) whereas in JJA, the temperature is overestimated in the ECHAM4 control simulation. The negative temperature bias in winter may be due to a too low surface albedo which reduces the available radiation for heating the soil and the adjacent boundary layer. In the Himalayas, an underestimation of the surface temperature is also found. This may be connected to the problem that most of the measurement sites in mountainous regions are situated in valleys. This effect may also contribute to the negative bias in the Andes and Rocky Mountains. The patchy pattern often found over ocean surfaces is due to the spectral representation of the surface height in the ECHAM GCM, which leads to an artificially waved ocean surface. The warm DJF bias in Europe found in ECHAM3/T42 due to anomalous warm air advection (Roeckner et al., 1992), is strongly reduced in the new model version.

In the Antarctic region, the difference pattern between the observation and ECHAM4 are fairly similar in both the summer and winter season. The warm bias off the Antarctic coast includes the regions covered with sea ice. Problems with detailed data of sea ice coverage or the poor parameterization of the processes over sea ice in ECHAM4 may lead to this overestimation in surface temperature. The principal deficiencies in the parameterization of sea ice processes in ECHAM4 are due to the neglect of snow melt and snow accumulation.
on sea ice, the omission of open water in leads and polynyas and the described sea ice coverage leaving out computation of sea ice thicknesses. These processes strongly influence the evolution of surface albedo (Allison, 1993) and as a result, the surface temperature.

Figure 3.6: Global distribution of the surface temperature and precipitation bias between the 10-year ECHAM4/T106 simulation and the LW90 climatology for the period JJA.

In large areas of Eurasia and North America, a warm bias has been detected in JJA. This is consistent with the lack of precipitation, the excessive summer drying and a too low cloud amount in ECHAM4. Large parts of Africa are slightly too warm except for the savanna zone north of the ITCZ which is influenced by the south west monsoon. This monsoon leads to rather strong precipitation which decreases the surface albedo due to moister soils. Since the (background) surface albedo is a constant for every grid point in ECHAM4, an overestimation of surface albedo during the wet season is assumed. This again leads to a decreased net shortwave radiation and thereby a cooling.
The impact of increased resolution can be seen when comparing Fig. 3.6 with Figs. 3.3 and 3.4 for both precipitation and surface temperatures during the NH summer season. The model biases of both the coarse (T42) and fine mesh (T106) look very similar during JJA. However, the global difference pattern of surface temperature and precipitation between T106 and LW90 show a more detailed structure than the corresponding result of T42 due to the finer mesh. Concerning surface temperature, the most apparent differences between the two resolutions occur in the Antarctica where the cold bias of T42 is significantly reduced in T106. This may indicate that the T42 resolution is somewhat too coarse to correctly describe the processes occurring at the slopes of Antarctic.

With regard to the precipitation, slightly larger differences between the two resolutions have been detected. The lack of JJA precipitation in Europe is more pronounced in the finer resolution. The excess of precipitation over the Indian Ocean and the Caribbean Sea is more pronounced in the T42 resolution. In summary, both model resolutions reproduce the mean climatology of precipitation and surface temperature very similarly. The most apparent differences occur in mountainous regions where the better resolution is beneficial.

### 3.4 Snow water equivalent

Of all the surface conditions, snow shows the largest spatial and temporal fluctuations. Over 50% of Eurasia and North America can be seasonally covered by snow (Robinson et al., 1993). Snow depth exhibits a strong seasonal cycle with a distinct interannual variability in the middle latitudes.

Snow is an important component in the Earth’s heat balance. The incoming shortwave radiation is strongly reflected due to the high snow albedo. The induced radiative cooling is reinforced by the high thermal emissivity of the snow pack which increases static stability in the boundary layer and consequently reduces turbulent fluxes. This effect is reinforced by a reduced roughness of snow covered vegetation when compared to the snow free condition.

Snow plays a key role in hydrology as well. During winter a large water storage reservoir is filled up. During spring this stored water is released in the form of meltwater over a long period, thereby influencing total runoff and soil moisture.

Many studies have shown the importance of snow for weather forecasts as well as for climate simulations. The sensitivity of the Indian monsoon to the Eurasian snow cover has been confirmed by several numerical experiments (e.g., Barnett et al., 1989). Walsh and Ross (1988) tested the sensitivity of 30-day forecasts to continental snow cover and found great sensitivity over Eurasia.

Snow extent is related to many feedbacks (Randall et al., 1994), the most obvious being the snow albedo feedback: A positive temperature bias leads to more snow melt and favours rain over snowfall which leads to a decrease of planetary albedo. This allows absorption of more solar radiation and therefore reinforces further warming. Snow is a particularly good diagnostic for evaluation since in order to correctly model snow thickness both the temperature and precipitation distribution need to be realistic.

### Comparison with observations

The following comparison between the model and the simulation focuses on the condition during NH winter (DJF) and the melting season. Greenland must be excluded since the ECHAM model computes no snow melt and snow water equivalent ($S_n$) on ice but allocates grid element covered with sea ice a constant snow water of $S_n = 10$ m.

The comparison of the simulated snow water equivalent and the global snow depth clima-
ology compiled by USAF (cf. Section 2.1) is displayed in Fig. 3.7. Note that snow mass and not snow depth is the modeled variable. This means that assumptions have to be made concerning snow density in order to connect depth to mass (see Section 2.1). In North America, both model and USAF map show the snow boundary line, which corresponds about to the 0.5 cm contour line of the snow water equivalent, close to the 40°N latitude parallel. However, in the USAF-analysis, snow cover protrudes well south of this parallel at the southern end of the Rocky Mountains. The ECHAM4/T42 simulation accumulates most snow in the northern Rocky Mountains. On the other hand, the observed snow pack attains its maximum thickness in the eastern part of Canada southeast of Hudson Bay and in central Siberia. In these areas, \( S_n \) exceeds 16 cm. The model bias (lower left hand panel in Fig. 3.7) is most pronounced in the Rocky Mountains where the simulation produces a snow mass in excess of 6 cm water equivalent.

FIGURE 3.7: Simulated and observed snow water equivalent in DJF. Simulation: Control run with ECHAM4/T42; Observation: USAF snow depth climatology. Lower panels: mean and standard deviation of the differences between observed and simulated snow water equivalent. The standard deviation is normalized with the 10-year mean in DJF.
This large positive bias is in line with overestimated precipitation compared to both the LW90 and GPCP analysis. This may be due to the smoothed orography in the T42-model which hinders the humid oceanic air from losing its water close to the coast line. The model bias may, however, be somewhat smaller since the measurement sites are located predominantly in valleys and, therefore, tend to give an underestimation of the area-averaged snow depth.

A distinct lack of snow mass (more than 2 cm snow water equivalent) is found in the south eastern part of Canada east of Winnipeg and the region expanding from Siberia to the coastline of the Japanese Sea. The large underestimation in central/east Siberia can be observed from the beginning of the snow season in October (not shown) and increases during winter. On the contrary, the modeled snow deck is too thick in both Labrador and most of Canada in early winter till December. The difficulty in reliably portraying the snow cover condition is inherent to other GCMs as well (Foster et al., 1996). Since the temperature during DJF is well below freezing point in the above-mentioned regions, errors in surface temperature and the parameterization of melt processes should have a minor influence on the snow deck discrepancies. Precipitation in autumn and winter seems to be captured reasonably well by the ECHAM4 model (cf. Section 3.2). The most likely source of error results therefore in the computation of snow density as well as in a very sparse station network within the highly forested areas of Siberia and Canada. Furthermore, erroneous calculation of sublimation and the omission of snow interception processes may play a certain role.

Regions with shallow snow decks during winter show a large interannual variability, making the comparison between the observed and simulated snow water equivalent untrustworthy (cf. Fig. 3.7). In March, at the beginning of the melting season, the model bias looks very similar to the bias during the DJF season (not shown). Fig. 3.8 illustrates the distribution of the snow mass in April, as simulated and observed in the Northern Hemisphere. Whereas the USAF analysis shows in April substantial lower snow water equivalents than in March, the model produces only a slight decrease in snow mass. This leads to a strong overestimation in $S_n$ by more than 6 cm over Russia, Siberia and Canada. The delayed melting of snow during spring may have different reasons. Surface temperature is mostly too low in April which favors snowfall instead of rainfall. The positive snow albedo feedback may reinforce this process. In addition, precipitation is distinctly overestimated in the northern parts of the former Soviet Union and North America during April (not shown). The excess of precipitation amounts to 10 mm/month for vast continental areas north of the 50° parallel. Therefore, it may be that the snow is melting at a rate similar to that of the observed melt, but snowfall is too heavy in the model. However, it is supposed that the parameterization of snow melt also suffers from certain deficiencies. Therefore, snow melt processes need to be examined further in off-line experiments (cf. Section 4.8). Additionally, a source of error may be deficient sublimation due to inadequate wind fields or stability conditions. Associated with the cooling, there is often an increase in the static stability of the lower atmosphere. This curtails turbulent flux transfer and as a result, sublimation. Another weakness is the model’s partitioning of precipitation into rain and snowfall which is based on the hypothetic assumption that the melt of snow flakes is initialized when the mean layer temperature exceeds 2°C. A reduction of this value to e.g. 1°C, would lead to enhanced snow melt, thereby reducing snowfall at the Earth’s surface (cf. Section 5.8).

The model bias in snow accumulation in the Rockies and Himalayas and the Alpine region, respectively, is different. Whereas the Himalayas and Rockies show an excessive snow amount in T42, snow amount is underestimated in the Alpine region. This is mainly caused by the fact that T42 resolution is not sufficient to accurately represent the height
distribution of the Alps. The highest grid element in the Alpine region, using the T42-grid, is approximately 900 m a.s.l. A further reason for incorrect snow simulation over the Alps is as follows. The Alps consist, in the T42 resolution, of a few isolated grid points and thus, in a 3D-simulation the hypothetical snow covered grid elements are strongly influenced by the surrounding (mostly snow free) grid points. This may lead to a suppression of feedbacks inherent to more extended snow cover areas, e.g. the positive snow-albedo feedback. The so-called seeding effect may enhance the lack of snow in early winter: In nature, snow cover is first "seeded" at the highest elevations, and once an initial snow cover is in place the surface energy balance may be modified in such a way to encourage growth of the cover. This effect may be less important for the large mountain complexes as the Rockies and Himalayas.

Some information on zonal averages of $S_n$ is given in Fig. 3.9. The annual cycle of observed and modeled snow water equivalent is displayed for both Eurasia and North America. Note that North America does not include the Canadian Archipelago north of the 70°N parallel. Therefore, the latitude belt between 70° and 80°N only includes about 5 grid elements for
North America while Eurasia includes 40 grid boxes in the same area. The land mass of Eurasia between the 40° and 80°N parallel is about twice as large as North America. The most striking difference between the two land masses is the thicker snow pack over North America (except for the latitude belt north of 70°N (which is not representative). The observed snow pack in the latitude belt between 60°N and 70°N attains its maximum thickness in March. ECHAM4, however, shows the maxima one month later. The difference between the calculated and observed $S_n$ is considerably larger in northern Canada than in Siberia. From October to February in Siberia, snow mass is slightly underestimated, whereas in northern Canada, the model substantially overestimates $S_n$ the whole year round. Note that the model produces a maximal snow water equivalent of 17 cm but the observation attains only 12.5 cm. In midsummer (July and August), only a remnant of seasonal snow can be found along the Arctic coast in central and far eastern Siberia as well as the Canadian Archipelago (not shown). In general, it can be stated that ECHAM4 does a reasonably good job in fall and winter but suffers from many problems in correctly portraying snow during spring. The latitude belt between the 50° and 60°N parallel covers the main part of southern Canada and huge parts of the former Soviet Union with steppes and Taiga vegetation.

Figure 3.9: Annual cycle of zonal averages of the snow water equivalent for Eurasia as well as North America. The four displayed figures refer to 10° latitude belts between 40°N and 80°N. Thick lines: USAF snow depth climatology; thin lines: monthly means from the ECHAM4/T42 control simulation.
The discrepancy between the model and the USAF-analysis seems to be distinctly smaller in southern than northern Canada. This feature may be reinforced by interpolation problems in the data void regions of Canada and by the strong temperature gradient since temperature is an important parameter for the computation of $S_n$ from snow depth. However, temperature information is not incorporated in Eq. 2.1.

Note the large difference in the observed $S_n$ between Canada and Russia. This may be partly explained by the difference in size between Eurasia and North America. During winter a huge anticyclone is formed over the Asian continent, leading to a dry atmosphere and rare snowfalls. On the contrary, over the smaller land mass of North America only a weak high pressure system is normally generated during the winter season. In fact, from September till April, North America shows substantially more precipitation than Eurasia does between 50°N and 60°N (not shown). Large differences of surface temperature between same latitude belts of both continents may lead to unrealistic features due to the computation of snow density neglecting the temperature effect (Eq. 2.1). However, a computation of the annual cycle of the surface temperature for both continents (not shown) suggests the effect to be unimportant. The maximal snow water equivalent of close to 14 cm is well captured by ECHAM4/T42.

In the latitude belt between 40°N and 50°N, the simulated snow water equivalent in North America closely agrees with the observation, whereas in Eurasia a substantial overestimation is observed during winter and spring. The excellent agreement in North America is somewhat surprising since a transition zone, with high interannual variability of snow depth, is involved.

A source of errors in the portrayed evolution of simulated and observed snow depth may be due to the neglect of temperature in Eq. 2.1. An underestimation of snow density would typically occur during the snow melt in April, while too high densities would be the rule during the autumn season. Since ECHAM4 tends to substantially overestimate $S_n$ in spring, assuming a higher snow density would reduce the model's bias.

The key features remain the same when comparing the USAF-analysis with the ECHAM4 control simulation with substantially increased horizontal resolution. The delayed melting of snow during spring, the overestimation of snow depth in the Rockies and the Himalayas, as well as the excess of snow in extended areas of Canada and parts of Siberia during DJF are also found in the T106 experiment.

Substantial differences do, however, occur on the regional scale and in mountainous regions where the better resolution resolves more orographic details. The delayed snow melt in spring seems to be even more pronounced in the higher resolution.

### 3.5 Surface albedo

Numerous studies deal with surface albedo measurements for selected regions (e.g., Federer, 1968; Leonard and Escher, 1968; Pinker et al., 1980; Robinson and Kukla, 1984; Allison, 1993 and Konzelmann, 1994), but only some cover the whole globe (e.g., Hummel and Reck, 1979; Kukla and Robinson, 1980; Robinson and Kukla, 1985). These global datasets are compiled by a geographical classification plus measurements at selected "typical" locations. However, the quoted global surface albedo climatologies fail to provide the resolution necessary for GCM validation. Since surface albedo variations are spatially very incoherent, both interpolation and averaging are hazardous. The only two global datasets available with both a sufficient resolution and long-term measurements (period from 1984 - 1990) are based on the Pinker-algorithm (Pinker and Laszlo, 1992) and the Staylor-algorithm (Staylor and Wilber, 1990). These climatologies are derived from re-
mately sensed observations sampled during the International Satellite Cloud Climatology Project (ISCCP) (Schiffer and Rossow, 1983 and Section 2.1).

The surface albedo "measured" by satellites shows many deficiencies and uncertainties as has been outlined in the previous section and the general introduction (Section 1.3) and should, therefore, be treated very cautiously. Nevertheless, the comparison between the observed climatology and the ECHAM output yields some interesting results. The global

![Surface albedo map](image)

**FIGURE 3.10: Global distribution of observed mean surface albedo for January and July. The observations are based on SRB from 1984 - 1990. The data are interpolated from the ISCCP equal-area grid onto the T42-grid.**

The surface albedo map derived from SRB for January and July is presented in Fig. 3.10. The SRB data are given on the ISCCP equal-area grid and are interpolated onto the T42-grid. This procedure is questionable primarily in arctic and antarctic regions where the longitudinal extent of the gridboxes strongly increases with increasing latitude. Most of the Antartica and Greenland ice sheet have surface albedos above 0.8 in its corresponding winter time. Long-term albedos exceeding 0.7 are observed in northern Siberia and Canada.
with tundra cover during winter and in an isolated region of western Russia along the 50° parallel latitude where the dry climate allows for only sparse vegetation. The satellite-derived averaged winter albedo is between 0.6 and 0.7 for extended parts of western Russia and northern Scandinavia. Regions with dense boreal forests have (winter-) mean albedos lower than 0.6. Typical albedo values of approximately 0.4 found over boreal forests also occur over isolated areas of the Sahara desert, with albedos above 0.35 occurring over large parts of the Sahara and some regions of the Arabian desert. The albedo values found over the boreal forest are somewhat surprising since they are substantially higher than observed surface albedos from in-situ measurements (e.g., Federer, 1968; Betts and Ball, 1997). The simulated surface albedo over the Himalayan Mountains is substantially overestimated while the albedo in Western Europe is generally too low during winter. However, the coarse resolution of the SRB-grid (and T42-mesh) hinders detailed orographic information over the Alps. During summer, under snow free conditions, most regions excluding deserts exhibit a surface albedo between 0.15 and 0.2. Only densely forested regions have a satellite-derived surface albedo lower than 0.15. The surface albedo of the boreal forests amounts to approximately 0.13. The albedo for the Amazonian rain forest equals approximately 0.15. This is consistent with other studies, remembering the range of uncertainty adherent to remotely sensed measurements. However, Henderson-Sellers and Wilson (1983) gives a range (from a review of published measurements) from 0.07 - 0.15 for tropical rain forests. Therefore, the remotely sensed mean albedo equal to 0.15 may be somewhat overestimated.

The comparison between the ECHAM4/T42 control simulation and the observation for three selected periods is presented in Fig. 3.11. For an interpretation of the results, it is of principal importance if the bias in albedo can be related to biases of other variables, mainly the snow cover fraction and the snow water equivalent. Therefore, the model deviations from both the SRB albedo and the snow cover fraction (compiled by NOAA) were related to each other. However, the expected strong correlation between the albedo bias and the snow cover bias was not detected. This may be due to different reasons:

1. Instrumental problems inherent to remotely sensed measurements
2. Atmospheric effects and erroneous radiation transfer and reflectance models
3. Problems induced by interpolation on the same grid resolution
4. Uncertainties in the forest fraction allocated to each GCM grid element.

The uncertainty in the forest fraction may significantly influence the modeled albedo. The forest fraction in ECHAM3 and ECHAM4 differs greatly north of 20°N. Moreover, satellite-based derivations of surface parameters over forested regions are rather hazardous. In fact, omitting the grid elements with a forest fraction of over 20% provides a far better correlation between the bias of surface albedo and snow cover. For Eurasia, a correlation coefficient of close to 0.5 is achieved during the winter months, while the inclusion of all grid boxes lowers this value to approximately 0.2.


The problem relating to different observation periods may not be neglected. Bradley and Jones (1993) report that the decade of the 1980s had some of the warmest years of the century and perhaps of the past 400 years. This was confirmed by a computation of the decrease of snow cover fraction between the periods 1984 - 1990 and 1973 - 1996, respectively (Table 3.1). The evaluation detected a considerable decrease in the snow cover fraction f_s over the snow transition zone in October and spring (April, May). Assuming an albedo of 0.2 for snow free land and 0.7 for snow, the identified decrease in snow cover by close to 5% yields an alteration of land surface albedo of 0.025 over the snow transition zone. Note that the change in snow cover fraction is strictly limited to transition months.
TABLE 3.1: Changes in the mean snow cover fraction between the period 1973 - 1996 and the period 1984 - 1990 for the remotely sensed NOAA climatology. The specifications in percentages are given for Eurasia (EUAS) and North America (NA) for the regions with (1) $0.1 \text{ cm} \leq S_n \leq 2 \text{ cm}$ and (2) $S_n \geq 3 \text{ cm}$.

<table>
<thead>
<tr>
<th>month</th>
<th>EUAS (1)</th>
<th>EUAS (2)</th>
<th>NA (1)</th>
<th>NA (2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>-0.4</td>
<td>-0.5</td>
<td>0.2</td>
<td>-0.6</td>
</tr>
<tr>
<td>February</td>
<td>1.7</td>
<td>0.4</td>
<td>0.8</td>
<td>-0.2</td>
</tr>
<tr>
<td>March</td>
<td>3.0</td>
<td>0.3</td>
<td>3.2</td>
<td>-0.3</td>
</tr>
<tr>
<td>April</td>
<td>2.4</td>
<td>0.3</td>
<td>4.1</td>
<td>0.0</td>
</tr>
<tr>
<td>May</td>
<td>4.6</td>
<td>1.3</td>
<td>3.7</td>
<td>2.1</td>
</tr>
<tr>
<td>October</td>
<td>4.3</td>
<td>0.8</td>
<td>1.4</td>
<td>0.6</td>
</tr>
<tr>
<td>November</td>
<td>0.2</td>
<td>0.1</td>
<td>0.5</td>
<td>0.1</td>
</tr>
<tr>
<td>December</td>
<td>-0.2</td>
<td>-0.4</td>
<td>-1.0</td>
<td>0.0</td>
</tr>
</tbody>
</table>

whereas during winter, a slight decrease in the snow cover fraction has been revealed. It is obvious that for regions with thick snow cover, the changes are less significant. These findings signify that the SRB climatology produces lower albedos during winter when the sampling period would have been the same as for the NOAA data (1973 - 1996) or the USAF snow depth climatology (not specified, but probably representative for the period 1960 - 1985).

Despite the problems outlined in the previous paragraphs, some remarks about the surface albedo comparison, displayed in Fig. 3.11, are given. At the beginning of the snow season, the model's albedo for Eurasia and North America (excluding the Canadian Archipelago) is systematically higher than in the observation. Over Eurasia, this positive albedo bias may be related to excessive snow in northern Russia. In contrast, the excessive snow over northern Canada is not related to an overestimated surface albedo. The Sahara and the Arabian desert have significantly lower simulated than observed albedos throughout the year. However, one has to keep in mind that shortwave surface parameters are not expected to be accurate over either snow/ice or very bright desert surfaces (Whitlock et al., 1995).

During winter (Fig. 3.11b), the difference pattern between observed and simulated surface albedo differs considerably from that in October. In extended regions in Canada, excluding the Hudson-Bay, northern and eastern Europe as well as western Russia, the model underestimates the surface albedo. This is not well reflected in the corresponding characteristics of the snow water equivalent. Note the large overestimation of the albedo over the Himalayan area as well as the mountain ridges spreading between the Alps and the Caucasus. This seems to be related to a strong underestimation of the snow cover fraction in steep mountainous areas. This problem is further investigated in Section 5.2.

In April, the strong overestimation in snow due to a delayed snow melt in spring is fairly well confirmed in the albedo's bias. The substantial underestimation of $S_n$ in southern Scandinavia relates well to the underestimation of the surface albedo albedo. Discrepancies between the albedo and snow depth biases are mainly found over northwestern Siberia and northwestern Canada. This is likely due to the oversimplified Eq. 2.1 which fails to capture the complicated processes during snow melt correctly.

Some further information about the albedo differences between observation and simulation can be attained by comparing annual cycles for selected specific regions. For a thorough discussion of the albedo biases, it is useful to refer to the respective biases in snow cover fraction, snow water equivalent or precipitation. The corresponding charts are, however, not shown.
The modeled annual average of surface albedo for Eurasia is approximately 0.8% higher than the observed one. The most significant discrepancies are found in April and October. The delayed snow melt in spring and early snowfall in autumn may be partly responsible for these characteristics. However, the excess in $S_a$ in spring should generate a positive albedo anomaly, primarily in March and May. The snow cover fraction is significantly underestimated during winter (October through to February) which is not reflected in both the surface albedo and snow water equivalent. This can be interpreted as a hint to carefully investigate the parameterization for the snow cover fraction in the ECHAM GCM (see Section 5.1). The overestimation of the surface albedo in October may be partly related to the extraordinary warm years during the 1980s which led to a decrease of the snow cover fraction during that time (see Table 3.1) and, therefore, makes a comparison with SRB data, sampled from 1984 through the end of 1990, somewhat questionable. In fact, the underestimation in snow cover extent in October is considerably reduced.
Figure 3.12: Annual cycles of surface albedo for nine selected regions. Solid lines indicate observed values (SRB climatology, 1984-1990); dashed lines modeled values (T42, control simulation). R1 = Eurasia; R2 = North America; R3 = Eurasia, only grid elements with a forest fraction $a_f \leq 30\%$; R4 = Taiga; R5 = Western Europe; R6 = Sahara desert; R7 = North America, no snow deck; R8 = Antarctic; R9 = the Himalayas.

selecting the NOAA-charts from 1984-1990 only. In winter the modeled albedo is in good agreement with the observed albedo. However, there is some hint for error cancellation (see discussion below).

Excluding all grid elements with a forest fraction $\geq 30\%$ (signifying, on principle, the exclusion of the boreal forests, see panel R3 in Fig. 3.12), implies some changes in the albedo's bias. From October through to April, the model is approximately 0.02 higher than SRB. During autumn and winter, this result is confirmed by the evolution of the snow cover fraction that is more reliable due to the neglect of largely forested grid boxes. From April to May, the solid and dashed curves intersect each other, which is possibly related to a deficiency in the parameterization of the snow albedo. Since both the snow water excess, the negative temperature bias and the excessive snow cover fraction in May indicate an overestimation of the snow depth, the underestimation of the surface albedo in May is inconsistent. This discrepancy may be related to the assumption in ECHAM4, that the snow albedo decreases linearly between $-10^\circ$C and freezing point. A polynomial relationship (a detailed discussion is given in Section 5.3) between these two variables would yield a better agreement between model and observation in May.
Limitation to western Europe (panel R5 in Fig. 3.12) reflects, again, the overestimation in the snow free surface albedo by approximately 0.02. From January through to March, the SRB albedo is slightly higher than in ECHAM4, primarily due to discrepancies in the Alpine snow cover extent. The snow cover fraction is less in ECHAM4 than in the observation due to the coarse resolution in T42. The annual mean equals 2.6% and 4.5% in ECHAM4 and the remotely sensed NOAA observations, respectively.

The Taiga, the boreal forests in Siberia, has been defined using the Olson classification of vegetation types. The strong underestimation of the modeled albedo in winter is evident (region R4 in Fig. 3.12). However, the high values of remotely sensed albedo over the boreal forest are extremely unreliable. In-situ measurements (Betts and Ball, 1997; Robinson and Kukla, 1984) have revealed a very low mean surface albedo for snow covered boreal forests in the order of 0.2. Even assuming the selected Taiga region to be only half forested (and the remaining forest free land to be totally snow covered) would yield lower albedo values as suggested by the Staylor climatology. Moreover, satellite-based sampling of surface albedo over dense forests is very inaccurate. Further, very low incident (and thus reflected) radiation, as well as often cloudy weather conditions, may lead to poor results.

In the following, an attempt is made to estimate the winter surface albedo over the Taiga. The calculation is based on (i) $\alpha = 0.25$ for snow covered forests, (ii) $\alpha = 0.75$ for snow covered forest free land, (iii) $\alpha = 0.15$ for snow free land, (iv) $S_n$ and the forest fraction as in the ECHAM4 control simulation, and (v) Eq. 5.1. This leads to a mean winter albedo over the Taiga of approximately 0.4 which is substantially less than in both the observation and the control simulation.

The observed surface albedo over North America is in line with the model during April and May (region R2 in Fig. 3.12). This is surprising, since the snow depth is considerably underestimated in late spring. The substantial overestimation of precipitation the whole year round excluding autumn as well as the too cold winter temperatures imply the overestimation in snow depth to be realistic. The correct simulation of the surface albedo in spring may be due to an error cancellation. The excessive snow amount is cancelled out by a poor transformation of $S_n$ to $f_s$ which yields an underestimated snow cover fraction. The characteristics during summer are very similar to those detected over the Taiga and western Europe. This overestimation by close to 0.02 is consistent with the findings in Roesch et al. (1999) (see Appendix), where a comparison between some GEBA sites and ECHAM3/T106 is provided. The SRB albedo increases slightly from April to September. A similar increase has been detected in the development of the albedo over snow free North America. The decrease in soil moisture during this period (e.g., Idso et al., 1975), cannot be the reason, since it leads to a decrease instead of an increase in the surface albedo. The radiation trapping by vegetation is thought not to be the key factor, since maximum trapping of radiation by multiple reflection occurs at smaller zenith angles (especially when the canopy structure is vertical). This would yield lowest albedos around the summer solstice which has also not been observed. Considering the growth of the leaves, one may argue that during summer, the upper canopy leaves become larger, thereby exhibiting intrusion of radiation into the canopy (M. Claussen, pers. comm.). This process may be the reason for the increasing albedo. A change in the radiation spectrum, leading to a change in the broad band albedo, may be excluded since no trend of the ratio between the visible and near infrared incoming radiation is simulated.

In the Himalayan area, the monthly surface albedo strongly differs between model and observation (R9 in Fig. 3.12). Note that this region is composed of 68 T42 grid boxes. It comprises thus more than a limited area of the Himalayas but rather covers a larger area including foot-hills. While the simulated surface albedo is subject to a pronounced annual cycle with an amplitude of close to 0.2, the remotely sensed albedo barely shows
any dependence on the season. The difference in the mean annual surface albedo equals 0.07. The verification is not a simple task but according to the results of Douville et al. (1995b), the transformation of $S_n$ to snow cover fraction in ECHAM4 (Eq. 5.1), which does not incorporate any treatment of the snow cover over steep slopes, is probably too high. A more detailed investigation of that problem is given in Section 5.2.

Over the Antarctic ice sheet, ECHAM4 simulates a surface albedo of close to 80% the whole year round. The SRB, in contrast, shows two noticeable peaks in February/March and September/October. Since surface temperatures are very low and persistently below freezing point, melting processes do not occur. However, the albedo’s annual cycle closely follows the annual cycle of snowfall. High snowfall rates in spring and autumn (of about 2 mm/d) are related to high albedos and vice versa. The correlation coefficient equals 0.91. The observed range (0.76 - 0.87) is in good agreement with the albedo range given for the ice caps of Antarctica in the literature (0.75 - 0.89, see Henderson-Sellers and Wilson, 1983). Radiation measurements on Greenland, performed at the ETH camp during 1991 (Konzelmann, 1994), reflect a minimum monthly albedo in July (74.2%) and a maximum in May (85.9%) which, once more, agrees well with the above estimates.

In the Sahara desert (R6 in Fig. 3.12), ECHAM underestimates the surface albedo, primarily in late spring. The annual difference amounts to slightly more than 0.02, with the annual average of SRB being close to 0.39. However, Whitlock et al. (1995) caution that the accuracy of remotely sensed surface albedo may be low over very bright desert, such as the Sahara.

To summarize, the comparison between the observed and modeled albedo reveals deficiencies which may be due to errors in both the measurements and in the model. The model’s bias over snow free vegetated areas of Eurasia and North America is close to 0.02. The reflectance over snow covered boreal forest is likely to be overestimated in both the SRB and ECHAM4 (cf. Section 5.7). The Himalayas show a weak annual cycle in the SRB albedo whereas ECHAM suggests a significant increase in winter. It is likely that differences in the observation period can lead to substantial discrepancies due to the warm 1980’s.
4. Assessment of the ECHAM and EM land surface schemes in an off-line mode

4.1 Introduction

While the capability of climate models to reproduce observed climate has significantly increased in recent years, a number of problems related to model physics remain. Among these problems, the interaction of the land surface with the overlying atmosphere has become of increasing interest, particularly in the case of land-use changes, which today have a strong anthropogenic component. The inclusion of realistic land-surface schemes in numerical models has in recent times taken on an added importance because of the increased interest in land-use activities and their impact on climate. Therefore, a number of land surface models for GCMs have been developed during recent years, e.g., Dickinson (1978), SIB [Sellers et al. (1986)], BATS [Dickinson et al. (1986)], BEST [Pitman et al. (1991)], CLASS [Verseghy (1991)] and SECHIBA [Ducoudré et al. (1993)]. Numerous investigations have demonstrated that the simulated surface climate using general circulation models, numerical weather prediction models and regional atmospheric modelling systems is very much dependent on the formulation of the land surface schemes (e.g. Meehl and Washington, 1988; Gallimore and Kutzbach, 1989; Yamazaki, 1989; Barnett et al., 1989).

The three land-surface properties - albedo, roughness and hydrology (degree of soil wetness) - have been the subject of individual sensitivity studies. Rowntree (1983) reviewed 11 studies with emphasis on surface albedo and soil moisture content. The paper of Garratt (1993) provides an overall summary of current GCM surface schemes including the main results from many sensitivity studies undertaken with GCMs in the last two decades. He mentions the well-known relationship that continental evaporation and precipitation tend to decrease with increased albedo and decreased soil-moisture availability. For example, results from numerous studies give an average decrease in continental precipitation of 1 mm/day in response to an average albedo increase of 0.13.

Many studies deal with the impact of deforestation in tropical and boreal forests (e.g. Henderson-Sellers and Gornitz, 1984; Mylne and Rowntree, 1991; Bonan et al., 1992). In testing the sensitivity to vegetation cover, Dickinson (1978) found that evapotranspiration can increase by a factor of two as the amount of vegetation cover increases. Jacquemin and Noilhan (1990) found vegetation cover to be the most sensitive surface parameter, strongly influencing the Bowen ratio and the amount of sensible and latent heat flux. Contrary to this result, Wetzel and Chang (1988) and Siebert et al. (1992) noted a higher sensitivity to the amount of soil water content rather than to vegetation. Pitman (1994) claims that it is impossible to give a general ranking of the sensitivity to surface parameters. Moreover, he emphasizes that surface-atmosphere feedbacks are significant for determining these sensitivities. Additionally, the type of model simulation and the model complexity strongly influence sensitivity results.

Many sensitivity studies have shown that the regional and global climate is strongly influenced by albedo, surface roughness, soil moisture content and vegetation cover. However, the majority of the studies are based on short time integration in the range of several weeks or months, and merely examine the influence on one single parameter. Furthermore, most studies only use the maximum and minimum values of the range and can not specify the non-linear responses to the surface parameters.

The present study is based on an exhaustive analysis of the surface parameterization in both the ECHAM3 General Circulation Model (GCM) and the Europa-Modell (EM).
ECHAM is based on the ECMWF spectral weather forecast model and has been extensively modified for climate application at the MPI for Meteorology in Hamburg. It has been used for climate-related investigations by the ETH Climate Modeling Group in Switzerland through a joint collaboration. The EM, on the other hand, is an operational weather forecast model developed at the DWD (German Weather Service). Both models were run in an off-line mode to exclude effects resulting from feedbacks within the atmosphere.

The objective of this study is to show the dependencies of different important model variables on certain input parameters. The emphasis lies on a detailed description of the parameterizations used, a comparison of the model results and the sensitivities to different parameters. Testing the sensitivity is a useful tool to detect important differences and similarities between land-surface models. Since changing more than one variable will obscure the sensitivity in the experiments, only one land surface parameter is changed at one time. For example, changing the vegetation ratio also implies a change in the surface albedo, the surface roughness length and the soil-moisture availability. The inclusion of a land surface parameterization implies that the number of parameters required to provide initial and boundary conditions for a numerical simulation increases substantially. The uncertainty of the parameters is often quite large (e.g. leaf area index or field capacity). Therefore, it is of major importance to understand the impact of these parameters on the surface and soil temperatures, fluxes of latent and sensible heat, or on runoff. Without such assessments, it is difficult to correctly interpret comparisons of model output and observations. The present study allows for the evaluation of the effect of land surface changes on regional climate characteristics simulated by ECHAM or EM, and to understand the main differences of the land-surface schemes in these models.

Off-line experiments are better adapted to investigate regional characteristics in more detail. The main advantage consists of the exclusion of the land surface-atmosphere coupling processes which often make interpretation awkward. Since the model is not allowed to interact with the atmosphere, the performance of the land surface scheme can be tested in more detail.

Most of the results presented in this chapter appeared as MPI report 244, see Roesch et al. (1997).

4.2 Comparison of the land surface schemes of ECHAM and EM

This results in this chapter are based on the land surface schemes implemented in the ECHAM3 GCM of the Max Planck Institute for Meteorology, Hamburg (Roeckner et al., 1992) and the Europa-Modell (EM) (Edelmann et al., 1995). A detailed description and comparison of the land surface schemes is given in this section.

4.2.1 Soil temperature and soil heat flux

The calculation of soil temperatures is based on completely different methods in ECHAM and EM. ECHAM includes five layers (of increasing thickness with depth) for temperature. The uppermost layer is 6.5 cm thick, whereas the lowest has a thickness of 5.7 m. The thickness of the uppermost layer is important when determining the surface heat flux and
the diurnal cycle of surface temperature. The discretized heat conduction equations are

\[ \frac{\Delta T_i}{\Delta t} = \frac{F_s}{\rho c g \Delta z_i} + \frac{2\kappa(T_i - T_{i+1})}{\Delta z_i(\Delta z_i - \Delta z_{i+1})} \quad \text{(layer 1)} \]

\[ \frac{\Delta T_i}{\Delta t} = \frac{-2\kappa(T_i - T_{i-1})}{\Delta z_i(\Delta z_{i-1} - \Delta z_i)} \quad \text{(layers 2 to 5)} , \]

with

- \( \rho c g \) Heat capacity of soil per unit volume (2.4 x 10^6 Jm^{-3} K^{-1})
- \( F_s \) Sum of the radiative and turbulent fluxes at the surface if there is no snow; heat flux from the snow to the deep soil if snow depth exceeds 0.025 m water equivalent [Wm^{-2}]
- \( T_i \) Temperature of the i-th soil layer [K]
- \( \kappa \) Thermal diffusivity of the soil (7.5 x 10^{-7} m^2 s^{-1})
- \( \Delta z_i \) Thickness of soil layer i [m], requiring a zero heat flux at the bottom of the lowest soil layer. This ensures that no artificial heat sources and sinks affect the energy balance. The surface temperature is set equal to the temperature of the uppermost soil layer for snow free conditions. In the case of snow, the surface temperature is extrapolated from the temperature \( T_{sn} \) in the middle of the snow pack.

The Europa-Modell (EM) includes only two soil layers for temperature. The equations for the temperatures in the middle of the soil layers are shown below, positive fluxes being directed towards the surface

\[ \frac{(\rho c \Delta z_s) \partial}{\partial t}(T_s + T_B) = G_s - G_{SB} + G_{melt} \]
\[ \frac{(\rho c \Delta z_s) \partial}{\partial t}(T_B + T_M) = G_{MB} + (1 - f_s)G_B + f_s G_{SB} + G_{melt} \]
\[ \frac{(\rho c \Delta z_M) \partial}{\partial t}(T_M + T_U) = G_{UM} - G_{MB} , \]

with

- \( B \) Subscript for upper soil layer
- \( G_B \) Sum of radiative and turbulent fluxes at the snow free surface [Wm^{-2}]
- \( G_{melt} \) Heat flux produced by a redistribution of heat caused by melting snow [Wm^{-2}]
- \( G_{MB} \) Heat flux across the boundary between upper and lower soil layer [Wm^{-2}]
- \( G_S \) Sum of radiative and turbulent fluxes at the snow surface [Wm^{-2}]
- \( G_{SB} \) Heat flux from soil to snow [Wm^{-2}]
- \( G_{UM} \) Heat flux across lower boundary of lower soil layer [Wm^{-2}]
- \( M \) Subscript for lower soil layer
- \( f_s \) Snow covered fraction of the grid element
- \( S \) Subscript for snow
- \( T_B \) Temperature of snow free surface [K]
- \( T_M \) Temperature at the boundary between upper and lower soil layers [K]
- \( T_S \) Temperature of the snow surface [K]
- \( T_U \) Temperature at the lower boundary of the lower soil layer [K]
- \( \Delta z \) Thickness of soil layer [m]
- \( \rho c \) Heat capacity of the soil per unit volume [Jm^{-3} K^{-1}].
Temperature is linearly interpolated in snow and soil layers. In EM, the heat conduction equation is solved using the so-called 'Extended Force Restore' method (EFR-method) (Jacobsen and Heise, 1982). The EFR-method computes, for two preselected frequencies of harmonic forcing, the same surface temperature as the full heat conduction equation. The thickness of the soil layers is related to the periods $\tau_1$ and $\tau_2$ of the harmonic forcing:

$$ \Delta z_B = \frac{D_{1,2}}{1 + x}, $$
$$ \Delta z_M = \Delta z_B \left( \frac{\alpha}{\lambda_m} - 1 \right), $$

(4.3)

with

$$ D_{1,2} = \sqrt{2\omega_{1,2}}, $$
$$ \omega_1 = \frac{2\pi}{\tau_1}, $$
$$ \omega_2 = \frac{2\pi}{\tau_2}, $$

(4.4)

and

$$ \alpha = \omega_1 (1 + x + x^2), $$
$$ \beta_m = \omega_1 x \sqrt{1 + x^2} e^{\frac{x}{1 + x}}, $$
$$ x = \frac{\tau_1}{\tau_2}, $$

(4.5)

Heat conductivity of the soil [W K$^{-1}$ m$^{-1}$].

Soil heat fluxes in ECHAM can be determined by taking into account that the heat flux at the lower boundary of the deepest soil layer equals zero. Therefore, the surface ground heat flux $G$ can be computed as

$$ G = \sum_{K=1}^{5} \rho_g C_g \Delta z_K \frac{\partial T_K}{\partial t}, $$

(4.6)

with

$C_g$ Specific heat of soil [J kg$^{-1}$ K$^{-1}$]
$K$ Number of soil layer
$T_K$ Temperature of soil layer K [K]
$t$ Time [s]
$\Delta z_K$ Depth of soil layer K [m]
$\rho_g$ Density of soil (dependent on soil type) [kg m$^{-3}$].

EM applies a different method for calculating the soil heat fluxes. The EFR-method allows the computation of the soil heat fluxes as

$$ G_{UM} = \frac{\Delta z_B + \Delta z_M}{D_1} (T_M - T_U) $$
$$ G_{MB} = \frac{\Delta z_B - \Delta z_M}{D_2} \left[ -x(T_M - T_U) + (1 + x + x^2) \right], $$

(4.7)

where $G_{MB}$ represents the soil heat flux at a depth of $\Delta z_B$, while $G_{UM}$ is the corresponding flux at the lower boundary of the (modeled) soil. The heat flux at the lowest layer boundary is, in contrast to ECHAM, not zero. This distinction is one of the main differences in climate and weather forecast models and is based mainly on the request of the forecast model to save computer time.

Equation 4.3 and some transformations of Equation 4.7 lead to

$$ G_{UM} = -\sqrt{\lambda \rho_g} \tau_1 \frac{\sqrt{\pi x}}{1 + x} e^{\frac{x}{1 + x}} (T_M - T_U) $$
$$ G_{MB} = -\sqrt{\lambda \rho_g \tau_1} \frac{\sqrt{\pi x}}{1 + x} \left[ -e^{\frac{x}{1 + x}} - x(T_M - T_U) \cdot \frac{[(1 + x + x^2) - x\sqrt{1 + x^2} e^{\frac{x}{1 + x}}](T_B - T_U)}{(1 + x + x^2) - x\sqrt{1 + x^2} e^{\frac{x}{1 + x}}} \right], $$

(4.8)
abbreviations being defined in the context of Equations 4.2 and 4.3. Equation 4.8 can be simplified assuming $\tau_1 = 24 \text{ h}$ and $\tau_2 = 5\tau_1$:

\[
\begin{align*}
G_{UM} &= -\sqrt{\lambda\rho c/\tau_1}[0.68(T_M - T_U)] \\
G_{MB} &= -\sqrt{\lambda\rho c/\tau_1}[-1.28(T_M - T_U) + 1.58(T_S - T_U)].
\end{align*}
\tag{4.9}
\]

The parameters are typically $\lambda = 1.5 \text{ WK}^{-1}\text{m}^{-1}$, and $\rho c = 1.8 \cdot 10^6 \text{ JK}^{-1}\text{m}^{-3}$ at the location of Cabauw. This assumption leads to an estimate of

\[
\begin{align*}
G_{UM} &= 3.8 \text{ Wm}^{-2}\text{K}^{-1}(T_M - T_U) \quad \text{and} \\
G_{MB} &= 7.2 \text{ Wm}^{-2}\text{K}^{-1}(T_M - T_U) - 8.8 \text{ Wm}^{-2}\text{K}^{-1}(T_S - T_U).
\end{align*}
\tag{4.10}
\]

The heat flux at the lowest layer boundary increases by about $4 \text{ Wm}^{-2}$ when $T_M$ changes by $1^\circ\text{C}$.

4.2.2 Soil hydrology

The soil hydrology in ECHAM is based on three budget equations for

i. Snow water equivalent $S_n$ accumulated at the surface

ii. Water amount $W_l$ intercepted by the vegetation during rain or snow melt episodes (the so-called skin reservoir)

iii. Soil water amount $W_k$.

Changes in the snow water equivalent (land, no ice) are computed as

\[
\frac{\partial S_n}{\partial t} = \frac{E_{Sn} + P_{Sn} - M_{Sn}}{\rho_w},
\tag{4.11}
\]

with

- $E_{Sn}$ Evaporation rate per unit area over the snow pack [$\text{kgm}^{-2}\text{s}^{-1}$]
- $M_{Sn}$ Snow melt rate per unit area [$\text{kgm}^{-2}\text{s}^{-1}$]
- $P_{Sn}$ Snow fall rate per unit area [$\text{kgm}^{-2}\text{s}^{-1}$]
- $\rho_w$ Density of water [1000 $\text{kgm}^{-3}$].

Over glaciers and sea ice the snow water equivalent does not change.

The maximum content of the skin reservoir is calculated by

\[
W_{lmax} = W_{lmax}((1 - C_v) + (C_v\text{LAI})),
\tag{4.12}
\]

with

- $C_v$ Grid fraction covered with vegetation
- $\text{LAI}$ Leaf area index
- $W_{lmax}$ Maximum amount of water held in one layer of leaves or bare ground [$2 \cdot 10^{-4} \text{ m}$]
- $W_{lmax}$ Maximum skin reservoir content [m].

The rain water and melting snow on the leaves are intercepted by the canopy until its water holding capacity $W_{lmax}$ is exceeded. The budget equation is given by

\[
\frac{\partial W_l}{\partial t} = \frac{E_{skin} + C_{tp}C_v(C_uP_R + M_{Sn})}{\rho_w},
\tag{4.13}
\]

with
Fractional area wetted by rain during a timestep (currently 100% for large-scale rain and 50% for convective rain)

Coefficient of efficiency of rain and snow interception (currently 100%)

Fraction of the grid box covered with vegetation

Evaporation rate from the skin reservoir \([\text{kgm}^{-2}\text{s}^{-1}]\)

Rainfall rate per unit area \([\text{kgm}^{-2}\text{s}^{-1}]\).

The excessive water from rain and snow melt, which does not replenish the skin reservoir, is used to calculate the amount of soil infiltration and surface runoff.

The soil water content \((W_s)\) is enhanced by precipitation and is decreased by evapotranspiration and runoff:

\[
\frac{\partial W_s}{\partial t} = \frac{E - E_{\text{skin}} + P_R - P_{Ri} + M_{sn} - M_{sn1} - R_R - R_D}{\rho_w},
\]

with

\(E\) Grid-mean evapotranspiration rate per unit area

\(M_{sn1}\) Snow melt rate per unit area intercepted by the skin reservoir

\(P_{Ri}\) Rainfall rate per unit area intercepted by the skin reservoir

\(R_R\) Surface runoff rate per unit area from precipitation events and snow melt

\(R_D\) Runoff rate per unit area from drainage processes.

The parameterization of drainage distinguishes between fast and slow drainage, the threshold value being at \(W_s = 0.9 \cdot W_{smax}\). The exact formulae are

\[
\frac{R_D(\text{slow})}{\rho_w} = d_{min} \frac{W_s}{W_{smax}} \quad \text{if} \quad (W_{smin} < W_s < W_{dr})
\]

\[
\frac{R_D(\text{fast})}{\rho_w} = d_{min} \frac{W_s}{W_{smax}} + (d_{max} - d_{min}) \left( \frac{W_s - W_{smin}}{W_{smax} - W_{dr}} \right)^d \quad \text{if} \quad (W_s \geq W_{dr}),
\]

with

\(d = 1.5\)

\(d_{min} = 0.0005 \text{ mmhr}^{-1}\)

\(d_{max} = 0.05 \text{ mmhr}^{-1}\)

\(W_{dr} = 0.9 \cdot W_{smax}\)

\(W_{smin} = 0.05 \cdot W_{smax}\)

\(W_{smax} = \text{field capacity}\).

The surface runoff is calculated following the scheme by Dümenil and Todini (1992), which takes into account the subgrid scale heterogeneity of the grid box by introducing a structure parameter \(b\). This parameter depends on the standard deviation of the terrain height. The introduction of the fractional saturated area \(s/S\) allows a non zero surface runoff for not fully saturated soils:

\[
\frac{s}{S} = 1 - \left( 1 - \frac{W_s}{W_{smax}} \right)^b,
\]

with

\(b\) Structure parameter

\(s\) Saturated area of the grid box

\(S\) Total area of the grid box.
Rain water is transported away by runoff in the fractional saturated area \( s/S \), while the rain water infiltrates in the area \( 1-(s/S) \).

The EM incorporates a different set of parameterization formulae to describe the soil hydrology. The EM allows the implementation of two (or more) soil layers for soil moisture instead of the bucket assumption of ECHAM. The prognostic equations for snow amount, skin reservoir and soil moisture are, however, similar in both models. In EM the following set is applied:

\[
\rho_w \frac{\partial W}{\partial t} = PR - ABF_{WS} - FL_{SI} \tag{4.17}
\]

\[
\rho_w \frac{\partial W}{\partial t} = PR + r_i E_{p,i} - ABF_{WI} - INFIL + FL_{SI} \tag{4.18}
\]

\[
\rho_w \Delta z_{WB_1} \frac{\partial W}{\partial t} = (1-f_S-r_I)(E_b + E_{tr_1}) - ABF_1 + INFIL + FL_2 \tag{4.19}
\]

and for the deeper soil layers (\( K = 2, ..., NLWB \)):

\[
\rho_w \Delta z_{WB,K} \frac{\partial W}{\partial t} = (1-f_S-r_I)E_{tr,K} - ABF_K + F_{L,K+1} - F_{L,K} \tag{4.20}
\]

with

\( ABF \) Runoff
\( E_p \) Potential evaporation \( (E_p > 0 \) implies the formation of dew or hoarfrost) \([\text{kgm}^{-2}\text{s}^{-1}]\)
\( E_b \) Bare soil evaporation \([\text{kgm}^{-2}\text{s}^{-1}]\)
\( FL_K \) Water flux between soil layers \( K-1 \) and \( K \) \([\text{kgm}^{-2}\text{s}^{-1}]\)
(positive fluxes are directed toward the surface)
\( INFIL \) Flux from snow to skin reservoir caused by melting and freezing processes \([\text{kgm}^{-2}\text{s}^{-1}]\)
\( PR \) Precipitation (index S: snow, R: rain) \([\text{kgm}^{-2}\text{s}^{-1}]\)
\( r_i \) Water covered fraction of the grid element
\( f_S \) Snow covered fraction of the grid element
\( E_{tr,K} \) Transpiration \([\text{kgm}^{-2}\text{s}^{-1}]\)
\( W_i \) Water content of the skin reservoir \([\text{m}]\)
\( W_K \) Water content of soil layer \( K \) \([\text{m}]\)
\( W_S \) Water equivalent of the snow layer \([\text{m}]\)
\( \Delta z_{WB,K} \) Depth of soil layer \( K \) \([\text{m}]\)
\( \rho_w \) Density of water \( (1000 \text{ kgm}^{-3}) \).

The maximum skin reservoir content \( W_{I,\text{max}} \) is calculated by

\[
W_{I,\text{max}} = W_{I,MB}(1 + 5\sigma_{PLNT}), \tag{4.21}
\]

with \( W_{I,MB} = 0.5 \text{ mm} \). The vegetated area of the grid element is represented by \( \sigma_{PLNT} \), leading to a maximum skin reservoir content of 3 mm. The formula excludes any dependence on the leaf area index. Assuming \( \text{LAI} = 2 \), Equation 4.12 yields a maximum skin reservoir content equal to 0.4 mm in ECHAM.

The EM allows for a water flux between the different soil layers. The transpiration originates not only from the uppermost soil layer but also from lower ones (depending on the root depth of plants).

The surface runoff includes the following components:
- precipitation which exceeds the infiltration capacity
- runoff due to overfilled skin reservoir
- excessive melt water (which does not infiltrate)
- runoff of the uppermost soil layer.

The maximum infiltration rate is determined by a simplified Holtat equation (only for surface temperatures above freezing point):

\[ N_{max} = SVORO \cdot \max(0.5, \sigma_{PLNT}) \cdot K_1 \frac{PV - W_1}{PV} + K_2, \]  

where

\( K_1 \) Parameter for hydraulic conductivity [kg m^-2 s^-1]
\( K_2 \) Minimum infiltration rate [kg m^-2 s^-1]
\( N_{max} \) Maximum infiltration rate [kg m^-2 s^-1]
\( PV \) Porosity (total water-holding capacity) [m^3 m^-3]
\( SVORO \) Influence of subgrid scale orography (presently set equal to 1)
\( W_1 \) Relative water content of the uppermost soil layer
\( \sigma_{PLNT} \) Vegetated area of the grid element.

If more than half of the grid area is vegetated, the maximum infiltration rate is linearly increasing with vegetated area.

The infiltration of water from the skin reservoir into the soil is parameterized as

\[ VERS_{IN} = \begin{cases} 
0 & ; T_B \leq T_0 \\
W_{I, \rho_W} & ; T_B > T_0, 
\end{cases} \]  

with

\( T_B \) Temperature of snow free surface [K]
\( T_0 \) Freezing temperature [273.15 K]
\( VERS_{IN} \) Infiltration of water into soil [kg m^-2 s^-1]
\( W_{I, \rho_W} \) Initial skin reservoir content reduced by the evaporation [m]
\( \tau_{VER} \) Time constant (set equal to 1000 s)
\( \rho_W \) Density of water [1000 kg m^-3].

This equation implies that after approximately 15 minutes, a full interception reservoir drains completely. Model experiments have shown that most of (liquid) precipitation (more than 99%) immediately infiltrates into the soil. Multiplying the time constant \( \tau_{VER} \) by five reduces the infiltration rate by the factor 5; the average of the skin reservoir content rises by a factor 5 and evaporation from the skin reservoir increases more than double.

If rain falls on unfrozen soil, the part \( \alpha P_{RR} \) of the rain replenishes the skin reservoir while the remaining rainfall \( (1 - \alpha) P_{RR} \) infiltrates. The parameter \( \alpha \) depends on the skin reservoir content \( W_I \) and its maximum \( W_{I, \text{max}} \):

\[ \alpha = \left( 1 - \frac{W_I}{W_{I, \text{max}}} \right)^{0.5}, \]  

An additional correction term is added to avoid a decrease in the skin reservoir content during rainfall. A number of experiments demonstrate that \( \alpha \) is generally > 0.9.

Runoff occurs when the field capacity of a soil layer is exceeded. If one sets \( FL_1 = -INFIL \), where \( INFIL \) is the infiltration into the soil, the difference of water fluxes can be written as

\[ \Delta F_{L_K} = F_{L_{K+1}} - F_{L_K}. \]
If $\Delta FL_K > 0$ and $W_K > FC$, then

$$ABF_K = \frac{W_K - FC}{PV - FC} \Delta FL_K,$$

(4.26)

where $ABF_K$ is runoff from soil layer $K$.

The transport of water between the soil layers is based on the Darcy equation, which includes the influence of gravity and capillary force:

$$FL = -\rho_W \left[D_W(W) \frac{\partial W}{\partial z} + K_W(W)\right],$$

(4.27)

with the hydraulic diffusivity $D_W$ ($[m^2 s^{-1}]$) and the hydraulic conductivity $K_W$ $[m s^{-1}]$ and

$$D_W(W) = D_0 e^{D_1 PV - ADP},$$

(4.28)

$$K_W(W) = K_0 e^{K_1 PV - ADP}.$$  

The four constants $D_0$, $D_1$, $K_0$ and $K_1$ depend on the soil type. According to Eq. 4.28, $K_W$ and $D_W$ increase exponentially between the air dryness point $ADP$ and the porosity. The water flux between the layers is not allowed to exceed a threshold of 10% of the porosity during one timestep.

4.2.3 Prediction of surface albedo and snow processes

Snow temperature

The computation of the snow temperature in ECHAM is different for snow water equivalents $S_n < 0.025$ m and $S_n > 0.025$ m, respectively. In the first case, Equation 4.1 is solved assuming the characteristics of bare soil. If $S_n > 0.025$ m, an extra heat conduction equation is applied according to

$$\frac{\partial (T_{S_n})}{\partial t} = \frac{F_S}{\rho_{S_n} C_{S_n} S_n},$$

(4.29)

with

- $F_S$: Sum of the radiative and turbulent fluxes at the surface
- $S_n$: Depth of the snow pack
- $T_{S_n}$: Temperature in the middle of the snow pack
- $\rho_{S_n} C_{S_n}$: Heat capacity of the snow per unit volume
  (using a density of snow $\rho_{S_n} = 300 \text{ kgm}^{-3}$).

In EM, the prediction of the snow temperature follows Eq. 4.2. In contrast to ECHAM, the heat equation for snow in EM allows for a heat flux from snow to soil and a heat flux resulting from melting snow and refreezing melt water.

Surface albedo without snow

In ECHAM, the surface albedo for every grid element under snow free conditions is constant. Therefore, the albedo does not depend on land use.

The parameterization of the surface albedo in EM distinctly deviates from the approach used in ECHAM. The albedo of snow free grid elements is not constant as in ECHAM, but depends on the soil water content of the uppermost soil layer, as well as on the albedo of vegetation and bare soil. The albedo under snow free conditions is computed from

$$\alpha_{sf} = \sigma_{PLNT} \cdot \alpha_{PLNT} + (1 - \sigma_{PLNT}) \left(\alpha_{bas} - \frac{LK \cdot W_1}{KR}\right),$$

(4.30)

with
KR  Term which includes heat capacity and heat conductivity of the soil
LK  Linear decrease coefficient
W1  Relative water content of the uppermost soil layer
αbas Surface albedo for bare soil (for dry soil, dependent on soil type)
αPLNT Albedo of vegetated areas (constant, equal to 0.15)
σPLNT Fraction of vegetated area
αsf Albedo under snow free conditions.

Surface albedo for snow covered land

In snow covered regions, the surface albedo in ECHAM is modified according to

\[ \alpha_{surf} = \alpha_{sb} + (\alpha_s - \alpha_{sb})f_s, \]  

(4.31)

with

\[ f_s = \frac{S_n}{S_n + 0.01 m} \]  

(4.32)

where \( S_n^* = 0.01 \text{ m} \).

The albedo of snow and ice surfaces \( \alpha_s \) is a function of the surface type, the surface temperature and fractional forest area (Equation 4.33). Below \( T_m = -10^\circ C \), the snow albedo is at its maximum value \( \alpha_{smax} \), while a minimum albedo (\( \alpha_{smin} \)) is assumed for \( T_m = 0^\circ C \) with a linear interpolation being applied between \(-10^\circ C\) and \(0^\circ C\). Over land, a weighted albedo of the fraction covered with forest and the remaining part of the grid element is applied:

\[ \alpha_s = \alpha_{smax} - (\alpha_{smax} - \alpha_{smin}) \cdot \frac{T_m - T_0}{T_m - T_0}, \]  

(4.33)

with

\[ \alpha_{smin}(a_f) = a_f \cdot \alpha_{smin}(a_f = 1) + (1 - a_f) \cdot \alpha_{smin}(a_f = 0) \]
\[ \alpha_{smax}(a_f) = a_f \cdot \alpha_{smax}(a_f = 1) + (1 - a_f) \cdot \alpha_{smax}(a_f = 1), \]

where

\[ a_f \quad \text{Fractional forest area} \]
\[ T_0 = 273.15 \text{ K} \]
\[ T_m = 263.15 \text{ K}. \]

The albedo for snow covered land is given by

\[ \alpha_{surf} = f_s \alpha_s + (1 - f_s)\alpha_{sf}, \]  

(4.34)

where

\[ \alpha_s \quad \text{Albedo of snow (set to 0.7)} \]
\[ f_s \quad \text{Snow covered fraction of grid element}. \]
The striking difference between ECHAM and EM is the missing temperature dependence of snow albedo and the lacking distinction between vegetated and forested areas. Figure 4.1 illustrates how the surface albedo varies with the fractional forest area and the snow water equivalent.

The two models also differ in the parameterization of the snow cover fraction $f_s$ (Fig. 4.2). Assuming a powdery snow pack of about 5 cm height (corresponding to 0.5 cm water equivalent), EM and ECHAM compute $f_s = \sim 90\%$ and $f_s = \sim 30\%$, respectively. These differences yield to drastic biases in the respective surface albedos.
Snow melt

The parameterization of snow melt is solved in a different way in ECHAM and EM. In ECHAM, snow melts when both the temperature in the middle of the snow pack and the temperature of the uppermost soil layer reach 0°C. Snow which is thinner than 0.025 m may melt if the temperature of the uppermost soil layer equals 0°C.

The EM distinguishes snow melt from the upper and lower boundary of the snow pack. A sophisticated algorithm which allows for vertical redistribution of heat in the snow pack and in the uppermost soil layer, guarantees a realistic parameterization of snow melt. Two cases are distinguished:

1. No snow melts if the temperature of the upper boundary of the snow layer \( T_{S,N} \) is above 0°C, while the temperature of the lower boundary \( T_{B,N} \) is below the freezing point. \( T_{B,N} \) increases as a consequence of a vertical redistribution of heat. Snow melt is initialized only if \( T_{B,N} \) exceeds 0°C.

2. The following amount of energy is available for snow melt if \( T_{S,N} < 0°C \) and \( T_{B,N} > 0°C \):

\[
WN_M = WM_S + WM_B. \tag{4.35}
\]

Snow melts only if \( WN_M > 0 \text{ J m}^{-2} \). If the mean temperature of the snow pack is above 0°C, the released heat available for snow melt is

\[
WM_S = \rho_{CS} c_S \Delta z_S \left[ \frac{T_{S,N} + T_{B,N}}{2} - (T_0 - \epsilon) \right], \tag{4.36}
\]

with

- \( c_S \) Heat capacity of snow [J kg\(^{-1}\) K\(^{-1}\)]
- \( S \) Subscript for snow
- \( T_{B,N} \) Temperature of lower boundary of the snow pack [K]
- \( T_{S,N} \) Temperature of upper boundary of the snow pack [K]
- \( T_0 \) Melting point (273.15 K)
- \( \Delta z_S \) Snow depth [m]
- \( \epsilon \) Very small number \((10^{-10} \text{ K})\)
- \( \rho_S \) Density of snow.

The reduction of \( T_{B,N} \) to the freezing point yields the release of the following heat amount which is available for melt:

\[
WM_B = (\rho_c \Delta z)_B [T_{B,N} - (T_0 - \epsilon)]. \tag{4.37}
\]

The calculation of the snow depth requires, in addition to the snow water equivalent, the snow density. In ECHAM, a constant snow density of 300 kgm\(^{-3}\) is assumed, whereas in EM the following relationship is applied:

\[
\rho_S = \rho_{S_0} + \min(\rho_{S_{max}}, W_S \frac{\partial \rho_S}{\partial W_S}), \tag{4.38}
\]

with

- \( \rho_S \) Density of snow
- \( \rho_{S_0} \) Minimum snow density \((500 \text{ kgm}^{-3})\)
- \( \rho_{S_{max}} \) Maximum snow density \((800 \text{ kgm}^{-3})\)
- \( \frac{\partial \rho_S}{\partial W_S} \) Coefficient of snow density change with water equivalent of snow pack.

This formula describes a linear increase of snow density with snow water equivalent. The maximum snow density is obtained for \( W_S \geq 1 \text{ m} \).
4.2.4 Parameterization of water vapour fluxes

In ECHAM and EM, evaporation is parameterized using a heat transfer coefficient $C_h$. However, the approaches in both models show some substantial differences.

Evaporation from snow and the skin reservoir is at the potential rate in both the ECHAM and EM:

$$E_p = \rho C_h |v_h| (q_v - q_{sat}), \quad (4.39)$$

where $C_h$ is the heat transfer coefficient, $q_v$ the water vapour mixing ratio, $\rho$ the density of air, $q_{sat}$ the water vapour mixing ratio at saturation and $v_h$ the horizontal wind vector. The water vapour mixing ratio $q_v$ and $|v_h|$ refer to the lowest atmospheric model level while $q_{sat}$ refers to the surface temperature $T_s$ and surface pressure $p_S$.

In order to compute the evaporation from dry bare soil (no water in the skin reservoir) in ECHAM, the relative humidity at the surface $h$ is introduced.

$$E_b = \rho C_h |v_h| (q_v - h q_{sat}), \quad (4.40)$$

where $h$ is defined as

$$h = \max \left[ 0.5 \left( 1 - \cos \left( \frac{\pi W_s}{W_{smax}} \right) \right), \min (1, \frac{q_v}{q_{sat}}) \right]. \quad (4.41)$$

The EM bare soil evaporation is expressed as:

$$E_b = (\beta_E)^2 \rho C_h |v_h| (q_v - q_{sat}). \quad (4.42)$$

For ice covered areas, $E_b$ is equal to the potential evaporation, whereas evaporation vanishes for the soil type named 'rocks'. The coefficient $\beta_E$ is represented by a linear function of the soil water content of the uppermost soil layer: $\beta_E$ vanishes for soil moistures below the air dryness point and is at its potential rate for saturated soils above field capacity.

The evaporation from dry vegetated areas (transpiration) is parameterized in ECHAM using the stomatal resistance and the water stress factor:

$$E_{tr} = \rho C_h |v_h| E(q_v - q_{sat}). \quad (4.43)$$

The evaporation efficiency $E$ is expressed as:

$$E = \left[ 1 + \frac{C_h |v_h| R_{co}(PAR)}{F(W_s)} \right]^{-1} = [1 + C_h |v_h| r_{st}]^{-1}, \quad (4.44)$$

where the water stress factor $F(W_s)$ is a function of soil moisture and $R_{co}$ is dependent on the photosynthetically active radiation (PAR), which is 55% of the net shortwave radiation at the surface. $r_{st}$ is the stomatal resistance, with a minimum value $R_{co}$. The water stress factor is equal to 1 if the soil moisture is above the critical value and is 0 for a soil moisture below the permanent wilting point, the point where wilting of plants is initialized.

In EM, a less sophisticated parameterization for transpiration is applied:

$$E_{tr,K} = (\beta_{B,K})^2 r_{WT,K} \rho C_h |v_h| (q_v - q_{sat}). \quad (4.45)$$

For soil types 1 and 2 (ice and rock) and surface temperatures below 0°C, transpiration is zero. The factor $\beta_{B,K}$ is represented similar to $\beta_E$. $\beta_{B,K}$ is parameterized by a linear function between the permanent wilting point ($\beta_{B,K} = 0$) and the so-called turgor loss point $TLP$ ($\beta_{B,K} = 1$), which is dependent on the soil type and potential evaporation. The function $r_{WT,K}$, ranging between 0 and 1, controls the water uptake of the roots.
in the different soil layers and is strictly between 0 and 1. The factor \( r_{WT,K} \) leads to a significant dependence of transpiration on rooting depth.

\[
\begin{align*}
\ r_{WT,K} &= \min \left( BWT, \Delta z_{WB,1} \right) / \sum_{k=1}^{NLWB} \Delta z_{WB,k} \quad ; K = 1 \\
\ r_{WT,K} &= \left( BWT - \sum_{k=1}^{K-1} \Delta z_{WB,k} \right) / \sum_{k=1}^{NLWB} \Delta z_{WB,k} \quad ; K > 1,
\end{align*}
\]

with

- \( BWT \) Rooting depth
- \( NLWB \) Number of hydrological soil layers
- \( \Delta z_{WB,k} \) Depth of (hydrological) soil layer \( k \).

Note that EM allows for a more sophisticated parameterization for transpiration and bare soil evaporation, which is based on a simplified version of the Dickinson parameterization (cf. Section 4.3.3).

### 4.2.5 Parameterization of boundary layer transport

In both the ECHAM and the EM, the parameterization of fluxes within the boundary layer is based on Louis (1979). The surface fluxes of heat, moisture and momentum are calculated by using the Monin-Obukhov theory. The transfer coefficients for heat and momentum are dependent on the roughness length \( z_0 \), the von-Kármán-constant \( k \) and a stability-correction function \( f \).

\[ C_h = \left( \frac{k}{ln(z_0)} \right)^2 f_h, \]

**Figure 4.3:** Heat transfer coefficient \( C_h \) (a) for unstable conditions, (b) Ratio between \( C_h(\text{ECHAM}) \) and \( C_h(\text{EM}) \). Parameters: \( z = 30 \text{ m}, z_0 = 0.1 \text{ m}, k(\text{ECHAM}) = 0.40, k(\text{EM}) = 0.36. \)
and for momentum

\[ C_m = \left( \frac{k}{\ln(z/z_0)} \right)^2 f_m. \] (4.48)

In EM, \( C_h \) is modified by introducing a specific roughness length \( z_h \) for heat and moisture. In the off-line model simulations, however, \( z_0 \) and \( z_h \) are assumed to be equal.

The correction functions \( f \) has been formulated by Louis (1982) as

\[ f_m = 1 - \frac{2bR_i}{1 + 3bc[\ln(z/z_0)+1]^2{\sqrt{\left[z/z_0+1\right]}R_i}} \] (4.49)

for ECHAM, and

\[ f_m = 1 - \frac{2bR_i}{1 + 3bc[\ln(z/z_0)+1]^2{\sqrt{\left[z/z_0+1\right]}R_i}} \] (4.50)

for EM in the unstable range. The functions \( f_h \) are very similar, with a minor difference in the numerator:

\[ f_h = 1 - \frac{3bR_i}{1 + 3bc[\ln(z/z_0)+1]^2{\sqrt{\left[z/z_0+1\right]}R_i}} \] (4.51)

for ECHAM, and

\[ f_h = 1 - \frac{3bR_i}{1 + 3bc[\ln(z/z_0)+1]^2{\sqrt{\left[z/z_0+1\right]}R_i}} \] (4.52)

for EM in the unstable range. The formulae for \( f_h \) and \( f_m \) look very alike in both models, the only difference being a simplification which has only been implemented in ECHAM. This simplification takes advantage of \( z_0/z << 1 \) where \( z \) is the height of the lowest atmospheric model level.

Under stable conditions, the expressions for \( f_m \) and \( f_h \) are identical in ECHAM and EM:

\[ f_m = \frac{1}{1+2bR_i/\sqrt{1+dR_i}} \]
\[ f_h = \frac{1}{1+3bR_i/\sqrt{1+dR_i}} \] (4.53)

with the constants \( b = c = d = 5.0 \).

In both model, the Richardson number \( R_i \) is calculated with its discretized form, the bulk Richardson number \( R_{ib} \):

\[ R_{ib} = \frac{g(\Delta T/\Delta z + g/c_p)}{T(\Delta e/\Delta z)^2}. \] (4.54)

where \( g \) is the gravitational acceleration, \( T \) the air temperature, \( c_p \) the specific heat of dry air (\( =1005 \text{ Jkg}^{-1}\text{K}^{-1} \)) and \( v \) the wind speed. \( \Delta \) refers to the difference between the Earth's surface and the lowest atmospheric model layer.

It is noteworthy that the values of the von-Kármán-constant \( k \) differ: the ECHAM GCM assumes \( k = 0.4 \) whereas EM sets \( k = 0.36 \). Under stable conditions, this yields a difference of \( (0.4^2 / 0.36^2 - 1) \cdot 100\% = 23\% \) in the transfer coefficient for heat and momentum.

A review of the relevant literature yields a \( k \)-value of slightly more than 0.4. The von-Kármán-constant as suggested in ECHAM is thus more reliable than the lower value in EM.

Fig. 4.3 illustrates how \( C_h \) is modified under unstable conditions.
4.3 Comparison of the sensitivities of ECHAM and EM

The sensitivity $S$ of a function $f(x)$ with respect to a parameter $x$ is defined as

$$S = \frac{f(x_2) - f(x_1)}{x_2 - x_1}.$$  \hspace{1cm} (4.55)

This is the derivative of $f$ expressed in the discrete form. Eq. 4.55 can be used to test the sensitivity of the land surface scheme with respect to an output variable $x$. To do so, a control experiment (CTRL) (see Section 4.3.1) was defined and off-line model simulations were performed varying one parameter $x$ over the range and step widths given in Table 4.1, fixing all other parameters as in CTRL. The last four parameters in Table 4.1 have proved to be of less importance than the others and have, hence, not been subject to the comprehensive sensitivity.

In each model simulation a number of output variables are calculated and stored in files. An overview of these variables is given, together with their units, in Table 4.5.

Some cases can be clarified using relative sensitivities which are computed as follows:

$$S_{rel} = \frac{f(x_2) - f(x_1)}{(x_2 - x_1)f(x_1)},$$  \hspace{1cm} (4.56)

where $f$ represents any output variable and $x$ the parameter.

The percentage change ($\Delta$) in a variable due to the change in a single parameter is calculated as

$$\Delta = \frac{PERT - CTRL}{CTRL} \times 100\%,$$  \hspace{1cm} (4.57)

where CTRL is the control value and PERT is the value simulated after a certain change in one parameter.

Sensitivities to parameter values can also be derived analytically using the derivation of the function $f(x)$ with respect to the variable $x$ ($\frac{df}{dx}$). This method excludes all model feedbacks but allows for a thorough discussion of the equation's structure.

Finally, it may help to investigate how model results are influenced by a substitution of a parameterization algorithm. This procedure gives further insight into the quality of the parameterization and may improve the understanding of the physical processes.

<table>
<thead>
<tr>
<th>parameter</th>
<th>range</th>
<th>step width</th>
</tr>
</thead>
<tbody>
<tr>
<td>albedo</td>
<td>0.15 - 0.35</td>
<td>0.05</td>
</tr>
<tr>
<td>snow albedo</td>
<td>0.5 - 0.9</td>
<td>0.05</td>
</tr>
<tr>
<td>roughness length</td>
<td>0.1 - 0.7 m</td>
<td>0.05 m</td>
</tr>
<tr>
<td>leaf area index</td>
<td>1 - 6</td>
<td>1</td>
</tr>
<tr>
<td>vegetation ratio</td>
<td>0.0 - 1.0</td>
<td>0.1</td>
</tr>
<tr>
<td>field capacity</td>
<td>0.20 - 0.40 m</td>
<td>0.05 m</td>
</tr>
<tr>
<td>wilting point</td>
<td>10 - 40% (of the field capacity)</td>
<td>5%</td>
</tr>
<tr>
<td>critical value</td>
<td>30 - 70% (of the field capacity)</td>
<td>10%</td>
</tr>
<tr>
<td>maximum amount of water</td>
<td>0.05 - 0.25 mm</td>
<td>0.05 mm</td>
</tr>
<tr>
<td>per leaf layer</td>
<td>0.0 - 1.0</td>
<td>0.1</td>
</tr>
<tr>
<td>coefficient of interception of precipitation</td>
<td>0.0 - 1.0</td>
<td>0.2</td>
</tr>
</tbody>
</table>

4.3.1 Control simulations

It is by no means simple to define the values of the parameters used in the control experiment (CTRL). Most problems are inherent to the different structures of the land surface
schemes in ECHAM and EM. The parameters of CTRL for the Cabauw site are listed in Table 4.2. Numbers 1) to 3) in the last column of Table 4.2 correspond to the following description.

Table 4.2: List of the parameters used in the ECHAM and EM land surface scheme for the control simulation at Cabauw. Numbers 1) to 3): see text below the table.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>description</th>
<th>ECHAM</th>
<th>EM</th>
</tr>
</thead>
<tbody>
<tr>
<td>ALB</td>
<td></td>
<td>albedo</td>
<td>0.156</td>
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<tr>
<td></td>
<td></td>
<td>depends on</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td>soil moisture</td>
<td></td>
<td></td>
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<tr>
<td>AZO</td>
<td>m</td>
<td>roughness length</td>
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<td>VGRAT</td>
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<td>vegetation ratio</td>
<td>0.886</td>
<td>6.886</td>
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<td>FORESTM</td>
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<td>fractional forest area</td>
<td>0.0</td>
<td>not defined</td>
</tr>
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<td>WMAX</td>
<td>mm</td>
<td>maximum moisture content of the skin reservoir</td>
<td>0.2</td>
<td>0.5</td>
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<td>WSMAX</td>
<td>m</td>
<td>maximum soil moisture content</td>
<td>0.2</td>
<td>1</td>
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<td>CVLT</td>
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<td>leaf area index</td>
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<td>2.0</td>
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<td>CVINTER</td>
<td></td>
<td>efficiency of interception of precipitation</td>
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<td>1.0</td>
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<td>CVA, CVB</td>
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<td>constants for definition of stomatal resistance</td>
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<tr>
<td>CVC</td>
<td>sm^-1</td>
<td>minimum stomatal resistance</td>
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<td>-</td>
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<td>ZEMISS</td>
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<td>emissivity of soil</td>
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<td>ALBSN</td>
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<td>min./max. of snow albedo (bare soil)</td>
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<td>0.7</td>
</tr>
<tr>
<td>ALBSNVEG</td>
<td></td>
<td>min./max. of snow albedo (vegetated areas)</td>
<td>0.3 esp. 0.4</td>
<td>-</td>
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<td>SNCRIT</td>
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<td>critical snow depth</td>
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<td>-</td>
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<td>SNDENS</td>
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<td>density of snow</td>
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<td>500 - 800</td>
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<td>SNHEAT</td>
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<td>Eq. 4.38</td>
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<tr>
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<td></td>
<td>structure parameter (Eq. 4.16) in the runoff scheme</td>
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</tbody>
</table>

1) Whereas in ECHAM it is straightforward to equate the maximum soil water content with the bucket depth, the EM-parameterization includes two different variables which are related to the maximum soil water content: the porosity, the water amount stored in the soil when all pores are filled with water, and the field capacity. Potential evaporation and runoff already reach their maximum when the soil water content exceeds the field capacity. Therefore, the relative soil moisture can be either defined as the ratio of the soil water content to the porosity or the ratio of the soil water content to the field capacity. Most studies define the relative soil moisture as the ratio between the soil moisture content and the field capacity. Thus, the relative soil moisture is, throughout the whole study, the soil moisture content with respect to the field capacity. Note that this definition allows for a relative soil moisture above 1.

2) The water stress in EM is computed with a formula which is dependent on the soil type and on potential evaporation. The soil parameters in the EM control simulation are the same as those used in EM for loamy sand which is the most frequent soil type in western Europe.

3) In EM, the heat capacity is a function of the soil water content. The value of dry soil depends on the soil type.

The data used in this study have been collected at the 213 m meteorological mast at Cabauw (51° 58'S, 4° 56'E) in the Netherlands and cover a one-year-period from 1.1.1987 - 31.12.1987. This site is located in flat terrain consisting mainly of grassland interrupted by narrow ditches. Up to a distance of 200 m from the mast, there are no obstacles or perturbations of any importance; further away scattered trees and houses are located in most directions (see Driedonks et al. (1978) for a more detailed description).
The model simulations were all run with a forcing at the lowest level of the model's atmosphere which is about 30 m above the ground. Averages over 30 minute intervals of the following variables were used:
- downward shortwave radiation
- downward longwave radiation
- total precipitation at the surface
- atmospheric temperature at the lowest model level
- zonal and meridional wind at the lowest model level
- specific humidity at the lowest model level.

Each model run comprises a period of five years. The forcing was cyclically repeated each year. The model results of the fifth year are presented in this study. This procedure is used to ensure that both models are in an equilibrium state.

The EM additionally needs the prescription of the temperature at the lowest soil layer \((T_U)\). In order to determine the annual cycle of \(T_U\) the following procedure was applied: In a first step, the measured temperatures at a depth of 2 cm at Cabauw were smoothed using a 31 day running mean. Then, the heat conduction equation (Eq. 4.73) was used to calculate the phase lag between the temperature at a soil depth of 2 cm and 47 cm which is the soil depth where \(T_U\) (see Fig. 4.4) is prescribed in EM. The second value corresponds to the depth where the temperature \(T_U\) is determined (the soil type in the control experiment is loamy sand). The calculation led to a phase lag of 34 days. In the following chapters the terms 'temperature of second soil layer' and 'temperature of third soil layer' are used. These temperatures are related to the definition of soil layers in ECHAM. The corresponding values for EM have been linearly interpolated from the temperatures \(T_B, T_M,\) and \(T_U\) (see Fig. 4.4). Temperature \(T_3\) in ECHAM will be compared to \(T_U\) in EM.

Figure 4.4: Schemes of soil layers and temperatures in EM and ECHAM (only the upper three soil layers are shown) for snow free conditions. The soil type used for all simulations with the EM land surface scheme is 'loamy sand' with layer depths as follows: \(\Delta z_B = 10.3\) cm and \(\Delta z_M = 37.0\) cm.
4.3.2 Sensitivity to the roughness length

The roughness length $z_0$ is an important parameter for GCMs and weather forecast models. The surface fluxes of momentum, heat and moisture are strongly determined by the roughness length. However, relatively little is known concerning exact values of roughness lengths for momentum, heat and moisture. The roughness length related to momentum is distinctly larger than $z_0$ related to heat and moisture fluxes (sensible and latent heat flux). Nevertheless, in ECHAM3 and ECHAM4, there is no distinction between these values. In the EM, distinction is possible but in the version used here both values are set on equal values. It should be emphasized that, although the roughness length is closely related to the geometry and height of the surface roughness elements, it is actually a matter of empirical coefficients. This emphasizes the importance of performing sensitivity studies for this parameter.

In this study, the roughness length was altered from 0.1 m to 1.0 m, with a step width of 0.1 m. These values correspond to a field and a forest in spring, respectively.

Roughness length is incorporated in the parameterizations for the turbulent fluxes using the formalism of the transfer coefficients (Eqs. 4.47 to 4.52).

![Figure 4.5: Annual response of latent and sensible heat flux, transpiration and bare soil evaporation to roughness length. The x-axis denotes which $z_0$-values [m] are used in the model simulation to compute the displayed sensitivities according to Eq. 4.55. Black: EM; shaded: ECHAM.](Attachment)
The general structure of the sensitivities can be better discussed on the basis of annual means because short term characteristics and accidental results can be avoided. Therefore, the following results are based on annual means (or sums). However, annual and diurnal characteristics should also be discussed.

Figure 4.5 depicts the sensitivity of turbulent heat fluxes and its components to $z_0$. The sensitivity of both water vapour fluxes and sensible heat to $z_0$ decrease exponentially for increasing roughness lengths. Latent and sensible heat fluxes show a higher (absolute) sensitivity in ECHAM whereas transpiration and bare soil evaporation are more sensitive in EM. These properties can be related to the derivative of the parameterization equations to the roughness length. The equation for water vapour fluxes and sensible heat flux incorporate the transfer coefficient $C_h$. The transfer coefficient increases for stable and unstable conditions with increasing roughness due to an enhanced turbulence over rough areas. The derivative $\frac{\partial C_h}{\partial z_0}$ (Fig. 4.6) is high for low roughness lengths and asymptotically approaches zero for very high $z_0$. The same characteristic is found for stably stratified boundary layers. A significant difference exists between the parameterizations of tran-

![Figure 4.6: a) Transfer coefficient $C_h$ (stable case) versus roughness length; b) derivative of $C_h$ with respect to $z_0$, based on the ECHAM parameterization. $v = 5 \text{ m s}^{-1}$ (wind speed on the lowest atmospheric model level), the temperature difference between the lowest model level and the surface is set to 0.5°C.](image)

Abstraction in ECHAM and EM. While the simple parameterization of EM puts the entire dependence on $z_0$ into the potential evaporation, the more sophisticated parameterization of the ECHAM model incorporates the evaporation efficiency $E$ which is again a function of $z_0$. $E$ becomes smaller with increasing roughness length, and therefore reduces the sensitivity of transpiration with respect to $z_0$. A typical situation is shown in Figure 4.7. It is obvious that the curve of the derivative quickly approaches zero and barely changes for $z_0 > 0.2$ m. Therefore, the sensitivity of transpiration with respect to $z_0$ decreases
The sensitivities of a number of variables to the roughness length are presented in Fig. 4.8. It shows the impacts caused by an increase in $z_0$ from 0.1 m to 0.2 m, which approximately covers the range of the roughness length's uncertainty at the Cabauw site. The ECHAM model generally shows, on an annual basis, greater sensitivities. The impact of $z_0$ on the evaporation fluxes from plants (transpiration), bare soil, the skin reservoir and snow are positive since rougher surfaces yield more turbulence which leads to higher turbulent fluxes (as long as soil water supply is sufficient). The percentage change (Eq. 4.57) in skin reservoir evaporation is more pronounced than the percentage changes in transpiration and bare soil evaporation. This is due to the explicit dependence of potential evaporation on the roughness length, while transpiration and bare soil evaporation incorporate further resistances which reduce its response to $z_0$. The absolute annual change in bare soil evaporation is larger in ECHAM, but the relative change is smaller since the annual bare soil evaporation of ECHAM is less than 60% of that in EM. The absolute sensitivity of the skin reservoir content is more than fifty times higher in ECHAM, the reason being the very fast infiltration of skin reservoir water in the EM. This results in a significant reduction of potential evaporation from the skin reservoir water in the EM compared to ECHAM.

Surface temperature and albedo, as well as short- and longwave net radiation are only slightly changed. The annual mean of surface temperature decreases by about 0.01°C
FIGURE 4.8: Simulation of the effect on a number of surface variables (annual basis) of changing the roughness length from 0.1 m to 0.2 m. Black: EM; shaded: ECHAM. Units as in Table 4.5.

for an increase in roughness length of 0.1 to 0.2 m (Fig. 4.8). The changes are very similar in both models. The main reason for low changes in surface temperature is the forced air temperature on the lowest model level (~30 m above the ground). A higher impact on temperature would likely be found when using 3-dimensional model runs. This is a severe deficiency in the model experiments and should be performed at a later stage. The sensitivity of soil temperatures and ground heat fluxes are also negligible. The surface albedo in ECHAM is merely affected by the snow water equivalent, whereas the EM albedo is additionally influenced by soil moisture in the uppermost soil layer. Nevertheless, the albedo of ECHAM is more sensitive since snow depth is distinctly more sensitive to roughness changes in ECHAM than in EM.

In order to close the energy balance, the higher latent heat flux must be compensated by a lower sensible heat flux. In the ECHAM model, the annual average of latent heat increases by ~2 Wm⁻² for a 0.1 m rougher grid element while the sensible heat flux decreases by a similar amount. The percentage change Δ is larger for the sensible heat flux as the annual average of the sensible heat flux is noticeably lower than the mean in the latent heat flux. Since more water is extracted from the soil over rougher surfaces by evapotranspiration, there is a decrease in the soil moisture and runoff fluxes which is shown by the negative values in Figure 4.8. Because water vapour fluxes are more sensitive to z₀ in ECHAM than in EM, the same applies (in order to achieve a correct water balance) to runoff at the surface and to drainage processes. While the absolute change in runoff is similar for both models, changes in surface runoff are (absolutely and relatively) far lower in the ECHAM model. The sensitivity of surface runoff, however, decreases much faster in ECHAM.

Snow melt and snow depth produced by the two models respond differently to changes
ECHAM shows a much higher sensitivity. While the percentage change (Eq. 4.57) in snow melt and snow water equivalent, caused by an increase in \( z_0 \) by 0.1 m, amounts to approximately 1% in EM, the simulations with ECHAM reveal a fourfold larger value. Snow depth and snow melt decrease with rougher surfaces because snow sublimation is significantly increased.

In EM, the response of ground heat fluxes to variation in \( z_0 \) is more than ten times larger than in ECHAM, the reason being the prescribed deep soil temperature in EM. The zero-flux condition at the lower boundary in ECHAM provides a better framework to level off forced changes. The EM loses (or gains) energy by the non-zero ground heat flux through the lowest soil layer.

Figure 4.9 demonstrates how the impact on the surface climate levels off with increasing roughness length. The bars represent the annual sensitivity ratios of \( z_0 = 0.45 \) m to \( z_0 = 0.15 \) m. The model runs used to compute the sensitivity for \( z_0 = 0.15 \) m (i.e. for the smoother surface) are \( z_0 = 0.10 \) m and \( z_0 = 0.20 \) m, and \( z_0 = 0.40 \) m and \( z_0 = 0.50 \) m for the roughness length of 0.45 m (i.e. the rougher surface). The response generally decreases with increasing roughness length. This corresponds with bar heights of less than 1 in Figure 4.9. For most variables, the ratio is approximately 0.5, i.e. the sensitivity for the rougher surface is half the sensitivity than for the smoother surface. The negative values for the surface and soil temperatures in ECHAM indicate that these variables decrease with increasing \( z_0 \) for relatively smooth surfaces, whereas the opposite applies to rough surfaces (represented here by the sensitivity for \( z_0 = 0.45 \) m). Differences between ECHAM and EM may be attributed to the described deep soil temperature in EM. The ratio in the albedo sensitivities are random since the effect of the roughness on albedo is minimal. However, differences between both models are mainly related to the inclusion of soil moisture in the albedo parameterization used in EM.

![Figure 4.9: Annual sensitivity ratios of \( z_0 = 0.45 \) m to \( z_0 = 0.15 \) m. Black: EM; shaded: ECHAM.](image)
Annual cycles

Most of the output variables show a reasonably distinct annual cycle of the sensitivities. Figures 4.10 and 4.11 review some of the results. They display the differences in a number of output variables due to a change in the roughness length from 0.1 m to 0.5 m. Runoff due to drainage is almost independent of the surface roughness from July to October since drainage is negligible during late summer/early autumn. In contrast to EM, the drainage in ECHAM shows a negligible sensitivity to \( z_0 \) during early summer. This difference is closely related to the rapid drying of the soil in ECHAM when compared to EM, which inhibits drainage processes. In November and December, ECHAM is more than three times more sensitive than EM. This difference is mainly caused by the incorporation of fast drainage in the ECHAM parameterization.

The distinct seasonal fluctuation of the skin reservoir content is only observed in ECHAM, whereas EM shows almost no dependence on the roughness length throughout the year. The fast infiltration in EM minimizes the effect of roughness length on the skin reservoir content. The pronounced annual cycle in ECHAM disappears completely when displaying percentage changes rather than absolute differences.

The response of soil moisture to changes in roughness length also differs considerably between both models. The absolute change is fairly constant during the year in EM and amounts to about 15 mm. The soil moisture content in ECHAM is, however, far less sensitive during winter (December to April), whereas during the remaining months, the sensitivity is comparable to that in EM. Figure 4.10 illustrates that the soil moisture simulated in ECHAM responds slightly to \( z_0 \) if the water supply is guaranteed, i.e. a change in roughness length affects the single components of water discharge and evaporation, but
the total water content of the bucket is relatively stable. During summer, in contrast, when turbulent fluxes are generally higher, both models do not compensate for the additional loss introduced by rougher surfaces.

Snow depth is generally very low at the Cabauw site, with the exception of January where more than ten days showed a measurable snow pack. In spite of a larger snow depth in the EM control simulation, ECHAM displays almost double the sensitivity of EM (Fig. 4.10). This also applies to snow melt and snow sublimation.

The response of surface temperature to an increase in $z_0$ from 0.1 m to 0.5 m is displayed in Fig. 4.11. The pronounced annual cycle is approximately equal in both models. During summer, increasing roughness yields lower temperatures, whereas the opposite applies for winter.

The differences in the net long- and shortwave radiation between the rough ($z_0 = 0.5$ m) and smooth surface ($z_0 = 0.1$ m) exhibit a pronounced annual cycle in both models, due mainly to the longwave part being changed by the surface temperature. Shortwave net radiation is very slightly reduced (to the order of 0.1 Wm$^{-2}$) in EM since drier soils lead to higher albedos. The ECHAM shortwave net radiation is affected only during winter as a result of variations in the snow cover.

The bare soil evaporation as computed with ECHAM is particularly sensitive during March and April (Fig. 4.11). This is due to the high soil moisture and the already rather high shortwave radiation in spring. During summer, however, such strong sensitivities are no longer observed because of the dry soil. Sensitivity of bare soil evaporation has an entirely different evolution in both models. EM shows the expected development: The greater the roughness, the higher the bare soil evaporation. A completely different impact of bare soil
evaporation to roughness length is simulated in ECHAM. During summer, rougher surfaces lead to less evaporation whereas the opposite applies during winter. This is associated with an effective drying of the soil during winter and spring over rough surfaces due to an enhanced evapotranspiration, leading to an excessive summer drying which limits bare soil evaporation. An interesting detail to note is that transpiration during summer also increases over rougher surfaces, caused by the capability of roots to draw water from deeper soil layers. The parameterization of ECHAM is - in spite of the bucket model representation - able to accurately capture this feature. The sensitivity of ground heat flux to $z_0$ is closely related to variations in the simulated surface and soil temperatures. During summer, the land surface becomes warmer over smoother surfaces, and thus, an additional downward heat flux is observed. During winter, the opposite conditions are found. This implies a decrease in the annual amplitude of monthly mean surface temperatures which, in turn, also reduces the annual amplitude in the surface ground heat flux. The prescribed deep-soil temperature $T_U$ in EM produces larger differences between soil temperatures and surface temperatures, and therefore, typically higher sensitivities of the ground heat flux to $z_0$. Prescribing the same surface temperatures and soil temperatures in both models leads to a close agreement in the simulated ground heat fluxes, in spite of completely different parameterization schemes.

Sensitivities during day- and nighttime

In addition to significant seasonal variations in the sensitivities, substantial differences in the surface variables due to a change in a parameter also occur between day and night. 'Day' is defined as being the time where shortwave incoming radiation exceeds $2 \text{ W m}^{-2}$. Figure 4.12 compares the effects of $z_0$ on a number of surface variables in both ECHAM

![Figure 4.12: Ratios in the annual sensitivity of ECHAM to EM. Black: day; shaded: night. Sensitivities are based on the model experiments using $z_0 = 0.1 \text{ m}$ and $z_0 = 0.2 \text{ m}$, respectively. The bars of snow melt and skin reservoir content are cut off for brevity.](image-url)
and EM. Positive ratios imply that the sensitivity of ECHAM and EM has the same sign, whereas negative values imply opposite signs for the model's sensitivities. Values close to 1 indicate that the respective sensitivity is similar in both models. The most striking variable is snow melt: In ECHAM, snow melt increases with increasing roughness length during the day, but decreases at night. The opposite applies for EM. In addition, the sensitivity is substantially larger in ECHAM mainly during night. This discrepancy is likely to be caused by the completely different structure of the snow melt parameterization.

The large positive ratios for the skin reservoir content (Fig. 4.12) are due to the substantial amount of skin reservoir water in ECHAM which is evaporated at the potential rate, whereas in EM, most of the water immediately infiltrates after a precipitation event. At night, bare soil evaporation is about twice as sensitive to $z_0$ in ECHAM than in EM but during the day the sensitivity is distinctly lower in ECHAM. This may be related to the difference as follows: In EM, the bare soil evaporation is controlled by the soil moisture of the uppermost soil layer whereas in ECHAM, the whole bucket determines the rate of bare soil evaporation. Furthermore, the bucket soil moisture hardly changes between the day and night whereas the thin uppermost soil layer introduced in EM, responds faster to changes in the radiation input or the precipitation frequency.

The different signs for shortwave radiation and albedo during the day are due to the impact of soil moisture on surface albedo which is excluded in ECHAM.

#### 4.3.3 Sensitivity to the leaf area index

The leaf area index (LAI) is the surface of all leaves projected on a horizontal plane. The determination of the LAI is a serious task and is afflicted with large uncertainties. The LAI evolves a distinct annual cycle in many regions of the Earth (e.g. in Europe). Despite this, the LAI is kept constant in time in both the ECHAM3 and ECHAM4 versions.

The LAI is not well adapted for capturing all the characteristics of vegetation. This is an insufficient parameter to describe certain physical processes (e.g., surface energy balance). Therefore, it may be useful to introduce additional indices in order to describe the vegetation more comprehensively. The vegetation indices were developed to reduce the number of parameters present in multispectral measurements to one parameter. A number of such vegetation indices have been proposed in the relevant literature (Verstraete, 1994).

Vegetation indices are based on remote sensing data and attempt to take advantage of the spectral contrast of green vegetation. The most widely used indices are the Simple Ratio (SR) and the Normalized Difference Vegetation Index (NDVI), defined below:

\[
SR = \frac{\rho_{NIR}}{\rho_{RED}} \tag{4.58}
\]

\[
NDVI = \frac{\rho_{NIR} - \rho_{RED}}{\rho_{NIR} + \rho_{RED}} = \frac{SR - 1}{SR + 1} \tag{4.59}
\]

where $\rho_{RED}$ and $\rho_{NIR}$ are the measured reflectances in the red and near-infrared spectral regions, respectively. The main advantage of these indices is that they can provide useful information at a low cost. They suffer a number of drawbacks, however, which include relatively high sensitivity to soils, atmospheric conditions or illumination. Some of these indices have been improved to reduce these sources of error. A recently proposed index is the Global Environment Monitoring Index (GEMI), which is a rather sophisticated function of two spectral bands (Verstraete, 1994).
Dickinson evaporation

All the EM-experiments of this chapter are based on the EM, incorporating the Dickinson transpiration (Dickinson, 1984; Edelmann et al., 1995). The parameterization used operationally for evaporation in EM is simple and does not take into account the leaf area index (Eqs. 4.42 and 4.45). As for ECHAM, each grid element was allocated a constant LAI, neglecting the annual cycle which is a reasonable assumption for sensitivity studies and which optimizes the interpretation.

A short description of the Dickinson parameterization allows for a better understanding of the following results. The parameterization for transpiration takes into account the stomatal resistance as the main element, which is a function of the soil moisture content, the photosynthetically active radiation, the temperature and the humidity. The formula for transpiration is

$$E_{tr} = \frac{R_{net} + \rho_0 \frac{C_A C_F}{C_A + C_F} B_e \partial q}{1 + \frac{C_A C_F}{C_A + C_F} B_e B_t}, \quad (4.60)$$

with

- \( B_e = \frac{\rho_0}{L} \left( \frac{\partial q}{\partial q} \right)^{-1} \)
- \( C_A = C_h \rho_h \quad [\text{ms}^{-1}] \)
- \( C_F = LAI \tau_{la}^{-1} \quad [\text{ms}^{-1}] \)
- \( C' = 0.05 \quad (\text{ms}^{-1})^{0.5} \)
- \( \tau_{la}^{-1} = C'(u_s)^{0.5} \quad [\text{ms}^{-1}] \)
- \( L: \) Specific heat of vaporization [Jkg\(^{-1}\)]
- \( q: \) Specific humidity [kgkg\(^{-1}\)]
- \( E_{tr}: \) Transpiration [mm\text{s}^{-1}]\)

The reduction factor \( \tau' \) contains the stomatal resistance \( r_{st} \) and the function \( \tau_{la} \). The formula can be simply transformed into an equation which is easier to derive with respect to the leaf area index:

$$E_{tr} = \frac{LAI (R_1 + Q_1 C_A) + R_1 C_A \tau_{la}}{(1 + B_e)[LAI + \tau_{la} C_A] + C_A B_e r_{st}}, \quad (4.61)$$

with

- \( Q_1 = \rho_0 B_e \partial q \quad [\text{kgm}^{-3}] \)
- \( R_1 = R_{net}/L \quad [\text{mm}\text{s}^{-1}] \)

and all other abbreviations as in Equation 4.60.

Discussion of parameterizations including the LAI

In this section, the parameterization formulae which are directly dependent on the leaf area index, are briefly discussed.

The maximum content of the skin reservoir \( W_{max} \) is linearly related to the LAI (Equation 4.12) and is parameterized in the same manner for both models when using the Dickinson parameterization in EM. A different approach for \( W_{max} \), which is independent of the LAI, is implemented in the operational EM (Equation 1.21). The water content on the leaves (skin reservoir content) is not only dependent on the potential evaporation rate and the precipitation amount but also on the distribution of total precipitation. Both models neglect interception of snow on the canopy.

The parameterization of transpiration includes the LAI. The parameterization of the stomatal resistance is the main part of both model’s formulae but they differ significantly. In ECHAM, the following formula is applied

$$r_{st}^{-1} = F(W_s) \frac{1}{k \cdot c} \left[ b \frac{d}{d + PAR} \ln \left( \frac{d \cdot e^{-k \cdot LAI}}{d + 1} + 1 \right) - \ln \left( \frac{d + e^{-k \cdot LAI}}{d + 1} \right) \right], \quad (4.62)$$
with

\[ a = 5000 \, \text{Jm}^{-3}; \quad b = 10 \, \text{Wm}^{-2}; \quad c = 100 \, \text{sm}^{-1}; \quad d = \frac{2+bc}{c^2\text{PAR}}; \quad k = 0.9; \]

\( F(W_s) \): Water stress factor;

\( \text{PAR} \): Photosynthetically active radiation.

This parameterization allows for the exponential decrease of shortwave radiation penetrating into a canopy layer: With increasing penetration into the canopy, the shortwave radiation, and therefore transpiration, is weakened. In contrast to the ECHAM algorithm, the Dickinson parameterization of the stomatal resistance excludes the leaf area index (although the whole transpiration formula does), but includes the dependence on temperature and humidity:

\[ r_{st}^{-1} = r_{s,max}^{-1} + (r_{s,min}^{-1} - r_{s,max}^{-1}) \cdot F_{WG}F_{ST}F_{TB}F_{DQ} \quad (4.63) \]

\[ F_{WG} = \begin{cases} 0 & ; \text{WBR} \leq \text{PWP} \\ \frac{\text{WBR} - \text{PWP}}{\text{FC}_B - \text{PWP}} & ; \text{PWP} < \text{WBR} < \text{FC}_B \cdot \text{FC} \\ 1 & ; \text{WBR} \geq \text{FC}_B \cdot \text{FC} \end{cases} \quad (4.64) \]

\[ F_{ST} = \begin{cases} 0 & ; \text{PAR} = 0 \\ \frac{\text{PAR}}{\text{RC}} & ; 0 < \text{PAR} < \text{RC} \\ 1 & ; \text{PAR} \geq \text{RC} \end{cases} \quad (4.65) \]

\[ F_{TB} = \begin{cases} 4(T_{2m} - T_0)[(T_{end} - T_{2m})] & ; T_{2m} \leq T_0 \\ \frac{(T_{end} - T_{2m})}{(T_{end} - T_0)^2} & ; T_0 < T_{2m} < T_{end} \\ 0 & ; T_{2m} \geq T_{end} \end{cases} \quad (4.66) \]

\[ F_{DQ} = \begin{cases} 0 & ; q_{2m} \leq q_{sat,2m} \cdot Q_{BRT} \\ \frac{q_{sat,2m} \cdot Q_{BRT}}{(1-Q_{BRT})q_{sat,2m}} & ; q_{sat,2m} < q_{2m} \leq q_{sat,2m} \\ 1 & ; q_{2m} \geq q_{sat,2m} \end{cases} \quad (4.67) \]

with

- \( \text{FC} \): Field capacity
- \( \text{FC}_B \): Fraction of field capacity below which water stress begins to work (= 0.75)
- \( \text{PWP} \): Permanent wilting point
- \( q \): Specific humidity [kgkg\(^{-1}\)]
- \( Q_{BRT} \): Fraction of saturation humidity below which stomatas are closed (= 0.75)
- \( \text{RC} \): 100 Wm\(^{-2}\)
- \( T \): Temperature [K]
- \( T_{end} \): 313.15 K
- \( T_0 \): Freezing point
- \( \text{WBR} \): Higher value of relative soil moisture of the two upper soil layers
- \( r_{s,max} \): Maximum stomatal resistance (= 4000 sm\(^{-1}\))
- \( r_{s,min} \): Minimum stomatal resistance (= 100 sm\(^{-1}\)).

These four functions control the dependence of transpiration on soil moisture, radiation, temperature and humidity (Fig. 4.13). All functions, with the exception of the temperature function, are linear and increase with increasing value. The function \( F_{TB} \) has a parabolic shape, and reaches a maximum at \( T = 20^\circ\text{C} \).
The stomatal resistance is equal to $r_{s,\text{max}}$ if
- temperature is below freezing point or over $40^\circ$C
- at night (no shortwave incoming radiation)
- soil moisture falls below the permanent wilting point, or
- specific humidity falls below 75% of the saturation value.

The last condition may be too restrictive. For the Cabauw site, the function $F_{DQ}$ is often close to zero and thus, strongly influences the transpiration. The original work of Dickinson omits this function because the relationship between the stomatal resistance and humidity is very uncertain and poorly explored. Therefore, the results presented here are based on the assumption $F_{DQ} = 1.0$. The critical point (start of water stress) is set to equal 75% of the field capacity for EM and 50% for ECHAM which is a significant difference. Remember, however, that the ECHAM soil can not exceed field capacity whereas the soil moisture of EM is only saturated at the pore volume.

**Discussion**

The typical exponential decrease of the transpiration's sensitivity to the LAI is illustrated in Fig. 4.15. As previously discussed, this feature is expected from a physical point of view. It is evident that the increase in transpiration is small when increasing the LAI from 9 to 10 due to an almost total shading of the lowest leaf layer by the remaining foliage.
FIGURE 4.14: Transpiration calculated with the ECHAM algorithm (Eq. 4.43 with Eqs. 4.44 and 4.62) as well as the Dickinson parameterization (Eqs. 4.60 and 4.61). Parameters (typical for a sunny summer day): PAR = 75 Wm\(^{-2}\), \(v_h = 6\) ms\(^{-1}\), \(C_h = 0.08\), \(R_l = 0.00012\) kgm\(^{-2}\)s\(^{-1}\), \(Q_1 = 0.002\) kgm\(^{-3}\), \(B_c = 0.5\), \(r_{st} = 263.0\) sm\(^{-1}\), \(C_A = 0.06\) ms\(^{-1}\), \(r_{al} = 25.82\) sm\(^{-1}\).

This characteristic can be shown by plotting the transpiration and its derivative for both models using the isolated formulae for transpiration (Fig. 4.14). It illustrates that \(\Delta\)Transpiration / \(\Delta\)LAI decreases faster when using the ECHAM parameterization rather than the Dickinson parameterization which indicates an almost linear decrease. Moreover, distinctly higher transpiration rates are computed for the Dickinson parameterization. The absolute values are likely to be overestimated since the calculation is based on a windy and sunny day over 24 hours. The evolution of the transpiration’s sensitivity (in EM) for LAI>3 is surprising. The changes are negligible and do not follow the characteristics found in Fig. 4.14, where the EM shows a higher sensitivity than ECHAM in the case of dense vegetation. This can be explained by analyzing the evolution of the soil water content: The high transpiration rates in the EM simulation strongly dries out the soil. Thus, transpiration no longer increases with higher leaf area indices due to a negative feedback.

The soil moisture content decreases with increasing LAI (see Fig. 4.15) due to higher rates of transpiration. The decrease rate (sensitivity) is more than three times higher for EM (LAI = 1.5) than for ECHAM. This rapid decrease in soil moisture leads to a lower relative soil water content for dense vegetation in EM when compared to ECHAM (not shown): During the late summer period (August through to September) the relative soil moisture in EM reaches unrealistically low values of approximately 30%. The water extracted from the soil by the roots during that dry period comes mainly from the uppermost soil layer. This is directly related to the algorithm used in EM: The water uptake from deeper soil layers is only allowed when the soil moisture falls below 1.25 times the wilting point, which
Soil water content is a key factor for bare soil evaporation, drainage and surface runoff. Annual means of these variables also show an exponential decrease of the sensitivity with increasing LAI as found for transpiration. However, surface runoff in EM behaves differently. For LAI > 3, the surface runoff increases substantially with increasing LAI, whereas the drainage decreases strongly and approaches values close to 0. Remember that most of the surface runoff consists of water discharge from the uppermost soil layers while drainage consists of the water discharge from deeper soil layers. Thus, the model results imply that the precipitation does not reach deeper soil layers when the soil moisture content is too low, but is lost to surface runoff and transpiration.

The skin reservoir content shows a drastically higher sensitivity for the ECHAM model (Fig. 4.15). This is due to the fast infiltration of water in the skin reservoir (EM), whereas in ECHAM the water remains on the foliage and evaporates at the potential rate.

The response of surface temperature to changes in the LAI is demonstrated in Fig. 4.16. Since an increase in the latent heat flux leads to a cooling of the surface due to the energy consumption for evaporating the water, the surface temperature decreases with increasing LAI. However, the temperature changes are minimal owing to the described temperatures on the lowest level in the model's atmosphere. The sensitivities of surface and soil temperatures to the LAI are very similar in ECHAM, but decrease faster in EM as a result of the prescribed deep soil temperature.
In ECHAM, the amount of foliage has no effect on the surface albedo. The surface albedo in EM responds slightly to changes in the soil moisture of the uppermost soil layer, i.e. higher LAIs lead to higher albedos since drier soils reflect more shortwave radiation than wet soils.

The impact of an increase in the LAI from 1 to 2 on a number of surface variables is displayed in Figs. 4.17 and 4.18.

As previously mentioned, variations in the LAI strongly influence the soil water content in EM. Total runoff constitutes the sum of surface runoff and drainage. The effect of LAI changes on total water discharge is, in a first approximation, equal in both models. However, the impact on drainage and surface runoff differs substantially. While surface runoff is far more sensitive in ECHAM, drainage is more sensitive in EM. In ECHAM, the decrease in the fractional saturated grid area due to a lower soil moisture content predominates the respective decrease in drainage caused by rarer occurrences of fast drainage. In EM, however, the water for the higher transpiration due to a denser foliage is extracted from the uppermost soil layer throughout the year since the soil water content remains above the threshold of 1.25 times the wilting point. This leads to a depletion of the available water in the uppermost soil layer and, consequently, to a rapid decrease of waterflow from the first to the second soil layer, thereby decreasing the lateral water discharge from the second and third soil layer (drainage).

The impact of the LAI on the turbulent heat fluxes is slightly higher in ECHAM when compared to EM. The sum of the latent and sensible heat flux barely responds to changes in the LAI since any increase in evapotranspiration is compensated by a lower sensible heat flux, which is caused by a cold temperature bias. The sensitivity of net radiation
to the LAI is similar for both models but for different reasons. The change in surface temperature is slightly larger for EM and leads to less upward long-wave radiation. This is compensated by a slightly higher EM-albedo due to a drier soil, resulting in increased reflected shortwave radiation.

The impact of the leaf area index on bare soil evaporation is significantly larger in ECHAM (Fig. 4.17) which can be attributed to different approaches in the algorithm for bare soil evaporation (cf. Section 4.6): The Dickinson parameterization for bare soil evaporation is less sensitive to the LAI than the parameterization applied in ECHAM.

Fig. 4.18 displays the same changes as in Fig. 4.17 but for changes in the percentage (Eq. 4.57), i.e. the simulated differences are normalized with the annual means of the control simulation. The graph illustrates that absolute and percentage changes in surface variables differ largely for some variables. The LAI generally has a significant influence on the water vapour fluxes, runoff processes and soil moisture. Increasing the LAI by 1 increases the annual evapotranspiration and the annual transpiration by approximately 10% and 25% (EM), respectively. This significantly reduces the annual water discharge. The net effect of an increasing LAI on the soil water is negative, i.e. the mean soil moisture is reduced by 6% and 16% in ECHAM and EM, respectively.

Since the annual mean of sensible heat flux is small (~6 Wm⁻² for LAI = 1 and ~3 Wm⁻² for LAI = 2), its percentage change is large. The fairly high Δ for the ground heat fluxes in EM are mainly due to the prescribed deep soil temperature.

The simulated sensitivity ratios for a dense (LAI = 4.5) and a transparent canopy (LAI = 1.5) are displayed in Figure 4.19. It shows that the sensitivity is generally de-
Figure 4.18: As Fig. 4.17 but expressed as percentage changes, i.e. modeled differences in the surface climate are normalized with the annual mean of the control simulation.

increasing with increasing LAI, which is expressed with bar heights less than 1.

The EM generally produces lower ratios: For transpiration, ground heat fluxes, turbulent heat fluxes, drainage and temperature, the quotient is close to zero. The surface climate in EM levels off when the LAI exceeds a value of 3 to 4. This is reasonable from a physical and meteorological point of view. There exists an optimum amount of foliage, i.e. if the canopy is (too) dense, no light can penetrate the lower canopy and a high transpiration rate draws (too) much water from the soil, thereby destroying the canopy because of a lack of water. This equilibrium condition is also simulated in ECHAM although much higher LAIs are required.

Annual cycles

The seasonal impact of LAI changes on surface variables differ considerably between ECHAM and EM (Fig. 4.20). The annual cycle of the transpiration difference shows a pronounced annual cycle in ECHAM which strongly decreases during late summer in EM. The high shortwave radiation input leads to an excessive drying of the EM-soil in April/May and reduces the available amount of water in such a way that the transpiration rate is reduced during late summer and autumn.

The change in surface runoff when increasing the LAI from 1 to 5, shows a fairly smooth evolution in ECHAM, whereas in EM, the surface runoff is higher for LAI = 5 during the first six months and decreases rapidly in June/July. The increase in surface runoff in June/July occurs simultaneously with the largely decreasing difference in transpiration (Fig. 4.20). This characteristic possibly reveals some deficiencies in the EM parameterization: It is unreasonable that the surface runoff increases when the soil moisture content
4.3.4 Sensitivity to the vegetation ratio

The impact of output variables on the vegetation ratio ($\sigma_{PLNT}$) is of great importance for climate change, i.e. vast regions of rain forests are being cut down, and deserts are expanding. Both LAI and $\sigma_{PLNT}$ can also change as a result of air pollution (e.g., through trees losing their leaves/needles). It is, essential, therefore, to understand the impact of variations in the vegetation ratio on the surface climate, and that the parameterizations represent the main feedback in a correct manner.

The vegetation ratio and the leaf area index represent similar parameters, since an increase in the LAI can lead to a similar increase in the vegetated area of a grid element. Thus, a number of results found in Section 4.3.3 can similarly be applied to the vegetation ratio.

It must be stressed that the EM includes no forest fraction whereas ECHAM distinguishes decreases.

The impact of the LAI on soil moisture differs considerably between both models (Fig. 4.20). The relative soil moisture in EM is reduced by 0.5 - 0.7 when increasing the LAI from 1 to 5, whereas the respective response in ECHAM is only 0.2 in late summer and autumn and negligible during winter/spring. The transpiration rate for dense canopies in EM is, presumably, too high and dries out the soil very effectively in early summer while the ECHAM algorithm for transpiration does not lead to such a pronounced deficiency of available soil water.

The phase shift of the change in ground heat flux (Fig. 4.20) is noteworthy: EM simulates the maximum change in June while the maximum change in ECHAM takes place in April. This shift is also simulated for latent and sensible heat flux as well as for surface temperature, but the phase shift period is about one month shorter.
between forested as well as non forested but vegetated areas. This means that EM contains no parameterization formulae which incorporates a distinction between forest and other vegetation (e.g. grass).

Discussion of parameterization formulae incorporating the vegetation ratio

The vegetation ratio affects the partition between the bare soil evaporation and the transpiration rate. The question arises as to which process extracts more water from the soil. This is, in ECHAM, dependent on different variables such as soil moisture, specific humidity, radiation and the leaf area index. Thus, it is difficult to decide which process is dominant.

In EM, the parameterization of transpiration and bare soil evaporation is less sophisticated than in ECHAM (see Eqs. 4.63 and 4.45). Figure 4.21 shows the functions \( \beta_E \) and \( \beta_B \) which control the bare soil evaporation and transpiration. The plot demonstrates that for very dry soils (below \( \sim 41\% \) relative soil moisture) and wet soil (above \( \sim 84\% \) relative soil moisture) bare soil evaporation is more efficient than transpiration in extracting water from the soil. Since, at the Cabauw site, wet soils predominate, it can be concluded that higher vegetation ratios generally yield lower sums of bare soil evaporation and transpiration. These calculations, based on the isolated parameterization equations, are confirmed in Fig. 4.22. The same characteristic is shown by the more sophisticated parameterization algorithm used in ECHAM.

The formula for the maximum skin reservoir content contains the vegetation ratio in both models. The derivative with respect to the vegetation ratio \( \sigma_{PLNT} \) is

\[
\frac{\partial W_{\text{Imx}}}{\partial \sigma_{PLNT}} = \begin{cases} 
W_{\text{Imx}}(LAI - 1) & \text{(ECHAM)} \\
5 \ W_{I, MB} & \text{(EM)}
\end{cases}
\] (4.68)
with

$$W_{I,MB} \quad \text{Constant set equal to 0.5 mm (EM)}$$

$$W_{\text{max}} \quad \text{Maximum amount of water held on one layer of leaf (ECHAM) [0.2 mm]}$$

$$W_{\text{imx}} \quad \text{Maximum content of skin reservoir [mm]}$$

Assuming LAI = 2, the sensitivity of the maximum water content to $\sigma_{PLNT}$ is distinctly higher in the EM. Nevertheless, the skin reservoir content is far more sensitive to vegetation changes in ECHAM. In both models, the water on leaves evaporates at the potential rate. In ECHAM, however, the water remains on the leaves for a much longer period than in EM, where the water is infiltrated very quickly into the soil from both vegetation and from the bare soil. Therefore, there is barely any change in the skin reservoir content with changing $\sigma_{PLNT}$. This is also applicable to the skin reservoir evaporation. In contrast to EM, the annual mean skin reservoir content in ECHAM is more than three times higher for fully vegetated regions than for desert conditions, and evaporation from the skin reservoir is nearly doubled (ECHAM).

The EM includes two further equations containing the vegetation ratio. The maximum infiltration rate (Eq. 4.22) is a function of $\sigma_{PLNT}$ for the range of $0.5 < \sigma_{PLNT} < 1.0$. The first term of Eq. 4.22 increases linearly with the vegetation ratio. This means that more water infiltrates into the soil, raising the soil moisture content and decreasing surface runoff. The sensitivity of the maximum infiltration rate ($N_{\text{max}}$) in relation to the vegetation ratio is

$$\frac{\partial N_{\text{max}}}{\partial \sigma_{PLNT}} = K_1(PV - W_1)/PV \quad (4.69)$$

**Figure 4.21**: Factor $\beta^2_B \cdot r_{WT}$ as used in Eq. 4.44 and $\beta^2_E$ specifying bare soil evaporation (Eq. 4.42). The function $\beta_{B,2} \cdot r_{WT,2}$ controls the transpiration from the lower soil layer whereas $\beta_{B,1} \cdot r_{WT,1}$ is related to the upper soil transpiration. Parameters $r_{WT,K}$ are calculated for a root depth equal to 0.7 m. Soil and root properties are as follows (typical values for sandy loam): $R_{WT,1} = 0.1$, $R_{WT,2} = 0.6$; air dryness point ADP = 3.0 Vol%, permanent wilting point PWP = 10 Vol%, field capacity FC = 26.0 Vol% and turgor loss point TLP = 17.2 Vol%.
for $0.5 < \sigma_{PLNT} \leq 1.0$ and zero for $0 \leq \sigma_{PLNT} < 0.5$. The symbols are as in Equation 4.22. The evaluation of Eq. 4.69 gives 0.0035 mm s$^{-1}$ (for very dry 'loamy sand') and 0.0055 mm s$^{-1}$ (for very wet 'loamy sand').

A further explicit dependence on $\sigma_{PLNT}$ can be observed in Equation 4.30, which describes the relationship between surface albedo and the vegetation ratio. The derivative of total albedo with respect to $\sigma_{PLNT}$ is:

$$\frac{\partial \alpha_{sf}}{\partial \sigma_{PLNT}} = \alpha_{PLNT} - \left( \alpha_{bas} - \frac{LK \cdot W_1}{KR} \right).$$ (4.70)

The symbols are as in Eq. 4.30. For the soil type of Cabauw, this equation reduces to $\frac{\partial \alpha_{sf}}{\partial \sigma_{PLNT}} = 4.25 W_1 - 0.15$. Assuming a saturated soil, this derivative amounts to approximately -0.06: The EM albedo decreases by 0.06 when the vegetation ratio increases from 0.0 to 1.0.

### Sensitivities derived from model experiments

In EM, the effect of the vegetation ratio on the surface albedo as determined with Eq. 4.70, is confirmed by sensitivity studies based on (off-line) model simulation (not shown). Thus, all additional feedbacks are minimal, or compensate each other, in EM.

The surface albedo, as generated by ECHAM, are almost independent of the vegetation ratio. A small impact on the surface albedo are found over snow covered regions, since changes in the vegetation ratio induce changes in the turbulent heat fluxes and, therefore, in the surface temperature.

The net shortwave radiation ($SW_{net}$) is modified according to the changes detected in the surface albedo. The sensitivity of $SW_{net}$ to $\sigma_{PLNT}$ equals 10 W m$^{-2}$ for low vegetation ratios and decreases to 7 W m$^{-2}$ in entirely vegetated areas. ECHAM shows a negligible change in net shortwave radiation due to the almost constant surface albedo. The impact of the vegetation ratio on water vapour fluxes differs in ECHAM and EM.

The sensitivity of transpiration (Fig. 4.23) is relatively constant for areas with less than 50% plant cover (~35 mm y$^{-1}$ for $\Delta \sigma_{PLNT} = 0.1$). The sensitivity rapidly increases with increasing vegetation ratio (EM), but remains fairly constant in ECHAM. The largest sensitivity is reached in fully vegetated areas for both models. It is evident that the sensitivities are all positive, since a larger vegetation ratio leads to more transpiration as long as soil water supply is sufficient.
The sensitivity of bare soil evaporation (Fig. 4.23) depicts a similar pattern as for transpiration, but with opposite signs. In ECHAM, the sensitivity is largest for no vegetation and fully vegetated areas, being at its minimum over moderately vegetated areas. The sum of annual transpiration and bare soil evaporation decreases with increasing vegetation ratio in both models, excluding almost vegetation-free areas (only EM). The impact of $\sigma_{PLNT}$ on the sum of transpiration and bare soil evaporation generally increases with increasing vegetation ratio (EM). In ECHAM, however, the sensitivity is significantly higher for low and high vegetation ratio while the minimum is associated with $\sigma_{PLNT} = \sim50\%$.

The results found in Fig. 4.21 suggest, for the soil type of loamy sand and a relative soil moisture between approximately 40% and 80%, that the sum of transpiration and bare soil evaporation increases with increasing vegetation ratio. For both wet and dry conditions, however, the opposite conditions apply. Since the annual mean soil moisture in EM is distinctly larger than 80%, a decrease in the sum of transpiration and bare soil evaporation with increasing vegetation ratio is expected. This has been confirmed in Fig. 4.23, i.e. the sensitivities for the sum of transpiration and bare soil evaporation are negative. The precipitation, intercepted in the skin reservoir, is lost for both transpiration and bare soil evaporation. This tends to further reduce the sum of transpiration and bare soil evaporation for high vegetation ratios in both models, and implies enhanced sensitivities for higher vegetation ratios. Note, that for $\sigma_{PLNT} > 0.5$, the sensitivities for the sum of these water vapour fluxes are similar in both models, whereas the individual components differ considerably. This, once again stresses the significant differences between the parameterization of the water vapour flux components. A rather insignificant effect may be
induced in EM in Eq. 4.69, which enhances the maximal infiltration rate with increasing vegetation ratio for \( \sigma_{\text{PLNT}} > 0.5 \). This process enhances soil moisture and depletes surface runoff and, thus, increases the latent heat flux. However, it cannot be excluded that this parameterization affects the distinct increase in the sensitivities in both bare soil evaporation and transpiration for vegetation ratios above \( \sim 0.5 \). A more detailed investigation regarding the infiltration processes and the frequency distribution of precipitation in EM would be necessary to prove such a relationship.

With the exclusion of very sparsely vegetated grid boxes, an increased vegetation fraction leads to less evapotranspiration, and higher sensitivities are associated with more vegetation (Fig. 4.24). The response of evapotranspiration to the vegetation ratio shows substantially lower sensitivities than the sum of transpiration and bare soil evaporation. This is related to the positive sensitivities of the skin reservoir evaporation to \( \sigma_{\text{PLNT}} \) due to an increasing amount of leaves with an increasing vegetation ratio.

The evaporation from the skin reservoir (not shown) is increasing with more vegetation in the ECHAM model. The annual sum is about 50 mm for \( \sigma_{\text{PLNT}} = 0.1 \) and double this for a fully vegetated grid box. In EM, the increase in the skin reservoir evaporation is more moderate (of the order of 1 mm per 10% increase in vegetation) because most of the skin reservoir water is infiltrated into the soil and is, therefore, lost to evaporation.

The model experiments show an increase in surface temperature in relation to more vegetation (Fig. 4.24) owing to decreasing evapotranspiration, with a lower albedo (EM) which enhances the absorbed solar radiation. The sensitivity of the surface temperature is considerably higher in EM than in ECHAM. This is due to the fact that both changes in radiation and turbulent heat fluxes play a considerable role in EM, whereas only changes in the latent and sensible heat flux are important in ECHAM. It should be noted that the lack of any canopy model in both models implies that the surface temperature in vegetated areas (especially forests) may be erroneous. However, at the Cabauw site with grass vegetation, this is of little importance.

Drainage and surface runoff generally increase with higher vegetation ratios, shown as positive sensitivities in Fig. 4.24. This is directly related to the lower latent heat fluxes with increasing \( \sigma_{\text{PLNT}} \). The result is opposite to that which is expected when reforesting landscapes with sparse vegetation: afforestation yields an increase in soil moisture content, higher evaporation rates and a reduced water discharge. This deficiency is presumably related to the missing canopy model in both the ECHAM and EM.

The ground heat flux is marginally influenced by vegetation changes in ECHAM but is rather strongly influenced in EM owing to the described deep soil temperature \( T_d \).

Fig. 4.25 gives an overview of the sensitivities for an almost fully vegetated grid element which is typically found in western Europe. The impact of vegetation on most surface variables predominates in EM, the main reason being differences in the parameterization of bare soil evaporation and transpiration. The less sophisticated parameterization in EM generally leads to greater sensitivities which is probably a result of neglecting some important feedbacks which are included in ECHAM. The significant impact of the vegetation ratio on the EM albedo also contributes to the generally higher sensitivity of the vegetation to the surface climate in EM when compared to ECHAM.
Fig. 4.24: As Fig. 4.23 but for surface temperature, sensible heat flux, drainage and surface runoff.

Fig. 4.25 illustrates that sensitivities of snow melt, snow depth and snow sublimation to $\sigma_{PLNT}$ are minimal. Including the interception of snow would likely enhance the effect of vegetation on snow variables. In ECHAM, the ground heat fluxes are slightly influenced by changes in the vegetation ratio whereas in EM, the impact is substantially larger. This is due to the prescribed deep soil temperature which cannot respond to possible variations in the surface temperature. The net shortwave radiation, as simulated in ECHAM, does not respond to changes in the vegetation ratio whereas in EM, the increase in soil moisture caused by an increase in $\sigma_{PLNT}$ from 0.8 to 1.0 leads to a substantial decrease in the surface albedo and thus, to enhanced net shortwave radiation.

Fig. 4.26 compares the ratios of sensitivities calculated for a low and high vegetation ratio. Note that for most output variables, e.g., turbulent heat fluxes, runoff, soil moisture and surface temperature, the sensitivity to $\sigma_{PLNT}$ is substantially reduced in sparsely vegetated areas. This is indicated by bar heights $> 1$. The increasing sensitivity with higher vegetation ratio is typical for both models and illustrates that the sensitivities can be highly nonlinear. For surface and ground temperatures (and thus, ground heat fluxes) the ratio is four times higher in ECHAM than in EM. Negative bars in Fig. 4.26 indicate an opposite sign in the sensitivity for $\sigma_{PLNT} = 0.1$ and $\sigma_{PLNT} = 0.9$. This applies to drainage, soil moisture and snow sublimation in EM, while in ECHAM, only surface runoff shows this characteristic. It is likely that these findings would differ significantly from results at locations with different climatic conditions (Pitman, 1994).
Annual cycles

The main features of the sensitivities can be captured by the annual means, as has been previously tested. However, in order to gather information on the seasonal response of the vegetation ratio to the surface climate, a number of surface variables are presented in Fig. 4.27.

Fig. 4.27 shows the effect of a change in the vegetation ratio from 0.4 to 1.0 on the annual cycles of some surface variables. The changes in surface runoff depict a pronounced annual cycle in EM with an amplitude of approximately 15 mm, while the corresponding differences in ECHAM remain below 5 mm throughout the whole year. This is likely due to the definition of surface runoff in EM, which includes the lateral waterflow from the uppermost soil layer. In ECHAM, however, non-zero surface runoff requires, particularly in more or less flat grid elements, the soil to be close to saturation.

The response of transpiration to an increase in the vegetation ratio by 0.6 exhibits a pronounced annual cycle in both models. The peak value for July in ECHAM is due to a warm and sunny mid-summer month. This enhances the photosynthetically active radiation (PAR) which substantially increases the transpiration rate. EM does not show this characteristic because PAR is not included in the parameterization of transpiration. Prior to April, the impact of an increasing vegetation fraction on bare soil evaporation is almost identical in both models but the sensitivity decreases strongly after April in ECHAM, while the difference remains high during summer in EM. The high negative response of bare soil evaporation in EM is compensated by the strong increase in the transpiration rate. Excluding the very sunny month of July, the difference in the sum of transpiration and bare soil evaporation compares fairly well in both models (Fig. 4.27). The sensitivity
of the skin reservoir evaporation with respect to the vegetation ratio is significantly higher in ECHAM since the water in the (EM-) skin reservoir rapidly infiltrates into soil and is, therefore, lost to evaporation. Preventing the skin reservoir water from infiltrating into soil would cause distinctly larger changes in the evaporation from the skin reservoir, since the maximal amount of water in the skin reservoir is substantially larger in EM than in ECHAM.

The large impact on transpiration due to changes in the vegetation cover implies the maximum sensitivity of soil moisture to \( \sigma_{PLNT} \) to occur as early as April and May in EM and ECHAM, respectively. The minimum difference in the soil moisture content is simulated in July and September for EM and ECHAM, respectively (Fig. 4.27). This phase shift is primarily caused by higher sensitivities of runoff and transpiration to the vegetation ratio in EM when compared to ECHAM.

The impact of an increase in the vegetation ratio from 0.4 to 1.0 on the surface albedo depicts a pronounced annual cycle in EM, showing an amplitude of approximately 0.04. The comparison of the simulated changes in the relative soil moisture content and the surface albedo reveals that, generally, the relative soil moisture does not follow the relative soil moisture of the uppermost soil layer which is included in the parameterization of surface albedo in EM (Eq. 1.30).

The impact of variations in the vegetated area on surface temperature (Fig. 4.27) is closely related to changes in evapotranspiration and absorption of solar energy. Since evaporation consumes energy, higher latent heat fluxes yield a negative temperature bias at the surface. Further, wetter soils tend to decrease the surface albedo and hence, lead to more absorption of solar radiation and thus, a warming of the surface. Note, that in three-dimensional model simulations, these conclusions are likely to be modified by various feedbacks.

\[ \text{FIGURE 4.26: Annual sensitivity ratios of } \sigma_{PLNT} = 0.9 \text{ to } \sigma_{PLNT} = 0.1. \text{ Black: EM; shaded: ECHAM.} \]
Comparison of day- and nighttime averages

Annual averages of day- and nighttime are compared in this section. Throughout this study, 'daytime' is defined as the period when the net shortwave radiation exceeds 2 Wm$^{-2}$. Table 4.3 lists a number of ratios of annual daytime means to annual nighttime means. The first two columns contain these ratios for $\sigma_{PLNT} = 0.9$ while the third and fourth columns contain the quotient of these ratios for vegetation ratios of 0.9 and 0.3, respectively. The values in the first two columns are detailed in Section 4.5 and are given in Table 4.3 for the sake of clarity.

A value equal to 1 (in the last two columns) implies that the ratio of the daytime to the nighttime annual averages is constant when increasing $\sigma_{PLNT}$ from 0.3 to 0.9. The tabulated ratios in Table 4.3 indicate that only slight changes in the ratios occur when changing the vegetated area from 30% to 90%, the sole exception being the sensible heat flux. A detailed investigation shows that the annual daytime mean of the sensible heat flux substantially increases with increasing vegetation ratios while at night, no significant changes are simulated.

Diurnal cycles

Some interesting results can be detected by plotting the mean diurnal cycle of individual months. Most variables show fairly smooth curves and the typical diurnal cycle expected
Figure 4.28: Monthly mean diurnal cycles of skin reservoir evaporation for a fully vegetated grid element. Solid: ECHAM, dashed: EM.

for radiation fluxes, surface temperature, transpiration, bare soil evaporation, latent and sensible heat flux, as well as ground heat fluxes (cf. Section 4.5). A surprising evolution is simulated for the diurnal cycles of skin reservoir evaporation (Figure 4.28) which strongly differs between EM and ECHAM. The evaporation rates are generally smaller during the day than at night, in accordance with higher radiation input and frequently higher wind speeds. Despite the same forcing for radiation and precipitation in both models, the evaporation from the skin reservoir hardly shows any similarities between both models. Closer investigation reveals that the diurnal cycles of the skin reservoir evaporation ($E_{\text{skin}}$) in ECHAM follow fairly closely the precipitation rates. $E_{\text{skin}}$ was found to be highest after precipitation events and concurrently high global radiation and wind speeds, as was to be expected. The variability in $E_{\text{skin}}$ is generally significantly larger than the corresponding variabilities in precipitation and global radiation. A correlation between time series of half-hourly rates of $E_{\text{skin}}$ with other surface variables demonstrates that the correlation is generally low. Highest correlation coefficients are computed during summer with respect to precipitation, global radiation and wind speed. In EM, the rapid infiltration of the water in the skin reservoir results in very low skin evaporation, which is poorly correlated with $E_{\text{skin}}$ in ECHAM.
Table 4.3: First two value columns: Ratio of annual mean during the day to the annual mean at night \( r_{dn} \). Last two columns: Ratio of \( r_{dn} \) for \( \sigma_{TPLNT} = 0.9 \) to \( r_{dn} \) for \( \sigma_{TPLNT} = 0.3 \).

<table>
<thead>
<tr>
<th>variable</th>
<th>ECHAM ( \sigma_{TPLNT} = 0.9 )</th>
<th>ECHAM ( \sigma_{TPLNT} = 0.3 )</th>
<th>EM ( \sigma_{TPLNT} = 0.9 )</th>
<th>EM ( \sigma_{TPLNT} = 0.3 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>8.3</td>
<td>10.2</td>
<td>1.13</td>
<td>1.06</td>
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<td>10.9</td>
<td>1.12</td>
<td>1.04</td>
</tr>
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<td>Sensible heat flux</td>
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<td>-2.1</td>
<td>1.51</td>
<td>2.30</td>
</tr>
<tr>
<td>Skin res. evap.</td>
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<td>0.97</td>
<td>1.03</td>
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<td>12.7</td>
<td>0.99</td>
<td>0.99</td>
</tr>
<tr>
<td>Bare soil evaporation</td>
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<td>14.5</td>
<td>0.87</td>
<td>1.29</td>
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<td>0.86</td>
<td>1.02</td>
<td>0.99</td>
</tr>
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<td>0.99</td>
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<td>1.00</td>
</tr>
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<td>1.13</td>
</tr>
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<td>Snow melt</td>
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<td>1.04</td>
</tr>
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<td>1.00</td>
<td>0.99</td>
</tr>
<tr>
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<td>0.54</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>Net radiation</td>
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<td>-3.4</td>
<td>0.99</td>
<td>1.04</td>
</tr>
<tr>
<td>Drainage</td>
<td>0.88</td>
<td>0.86</td>
<td>1.63</td>
<td>1.12</td>
</tr>
<tr>
<td>Surface runoff</td>
<td>0.84</td>
<td>0.77</td>
<td>0.98</td>
<td>1.15</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>0.96</td>
<td>0.92</td>
<td>1.00</td>
<td>0.96</td>
</tr>
</tbody>
</table>

4.3.5 Sensitivity to the background surface albedo

Various climate studies demonstrate that a change in surface albedo exhibits a large impact on both the surface climate and large-scale processes. A review on previous work is given in the general introduction (Section 1.3).

Sensitivities based on annual means

The sensitivity of net shortwave radiation \( (S_{net}) \) to the surface albedo is given by its definition. The simulated annual mean of \( S_{net} \) in both models is reduced by approximately 11 Wm\(^{-2}\) when increasing the surface albedo by 0.1 (Fig. 4.29).

The effect of surface albedo on the net radiation is, compared to the net shortwave radiation, slightly reduced due to a negative feedback: The higher the surface albedo, the lower the absorbed radiation by the Earth's surface. This leads to a lower surface temperature, which reduces the upward longwave radiation according to the Stefan-Boltzmann-Law:

\[
\frac{dLW}{dT_s} = 4\sigma T_s^3, \tag{4.71}
\]

with

\( LW \uparrow \) Upward directed longwave radiation [Wm\(^{-2}\)]
\( T_s \) Surface temperature [K]
\( \sigma \) \( 5.669 \times 10^{-8} \) Wm\(^{-2}\) K\(^{-4}\).

Eq. 4.71 yields a change in \( LW \uparrow \) of approximately 3.1 Wm\(^{-2}\)K\(^{-1}\) for \( T_s = 240 \) K and 6.8 Wm\(^{-2}\)K\(^{-1}\) for \( T_s = 310 \) K. For temperatures generally measured at Cabauw, \( LW \uparrow \) increases by approximately 1 Wm\(^{-2}\) for a temperature increase of 0.2°C. The substantial impact of the surface albedo on the surface temperature is mainly caused by the change in the energy balance. This response is equal to -0.2 °C for an increase in albedo of 0.1, which is similar in both models.
The impact of the background surface albedo $\alpha_{sb}$ on the turbulent heat fluxes requires a more detailed discussion. The parameterization of the turbulent heat fluxes contains the heat transfer coefficient $C_h$ (cf. Section 4.2.4) which is strongly dependent on the stability within the boundary layer. Since the temperature, the specific humidity as well as the wind speed on the first (atmospheric) model level are forced variables, the sensitivity of $C_h$ to surface albedo can be investigated using only the surface temperature. To calculate the impact of the surface temperature on the potential evaporation, the differentiation of the specific saturation humidity ($q_{sat}$) with respect to temperature is also necessary (Eq. 4.72).

$$
\frac{dE_p}{dT_s} = \rho \theta_h \left[ \frac{dC_h}{dT_s} (q_v - q_{sat}) + C_h \frac{dq_{sat}}{dT_s} \right],
$$

with all symbols as in Equation 4.39.

Fig. 4.30 displays, assuming reasonable parameters, the magnitudes for the derivatives of $C_h$ and potential evaporation with respect to the surface temperature. Fig. 4.30a shows the increase in $C_h$ with increasing surface temperature and a fixed temperature $T_1 = 280$ K. Since a warming at the surface destabilizes the boundary layer, Fig. 4.30a illustrates that $C_h$ increases with decreasing stability of the boundary layer. The differentiation of $C_h$ with respect to $T_s$ demonstrates that the impact of surface temperature on $C_h$ is low for a relatively stable situation and reaches a pronounced peak for neutral conditions for $T_s = T_1 = 280.3$ K. Note that the surface temperature required for neutrality is lower than $T_1$ due to wind shear stress.

The derivative of the saturation specific humidity with respect to the surface temperature ($dq_{sat}/dT_s$) is displayed in Fig. 4.30c. $dq_{sat}/dT_s$ increases slightly with $T_s$ and amounts to $2.3 \cdot 10^{-4}$ K$^{-1}$ and $7.7 \cdot 10^{-4}$ K$^{-1}$ for $T_s = 270$ K and 290 K, respectively. Fig. 4.30d
Figure 4.30: Differentiation of some variables with respect to surface temperature. a) $C_h$; b) Derivative of $C_h$ with respect to $T_s$; c) Derivative of specific humidity at saturation with respect to $T_s$; d) Derivative of potential evaporation with respect to $T_s$. Solid line: temperature dependency of $C_h$ included. Dashed line: Derived with the assumption that $C_h \neq C_h(T)$. All calculations are based on the equations used in the ECHAM model. The computation of specific humidity at saturation is based on the Magnus' formula. The following parameter set is presumed: $z_1 = 30$ m (height of first atmospheric model level), $v(z_1) = 2$ ms$^{-1}$, $T_1 = 280$ K (temperature on the first atmospheric model level) and relative humidity $q = 70\%$.

(solid line) shows the impact of changes in surface temperature on potential evaporation. Recognize that for very stable conditions, the derivative approaches zero which is physically reasonable. The sensitivity is largest for neutral conditions and amounts to 5 Wm$^{-2}$K$^{-1}$. An increase in the windspeed on the first model level from 2 ms$^{-1}$ to 5 ms$^{-1}$ would lead to a significantly lower peak value since $dC_h/dT_s$ is reduced as well. This is reasonable since a wind speed equal to 5 ms$^{-1}$ enhances the turbulence significantly and, hence, reduces the influence of any temperature changes close to the surface. In the unstable range, the sensitivity of the potential evaporation to the surface temperature slowly decreases, in accordance with Fig. 4.30b. The dashed line in Fig. 4.30d demonstrates that the derivative of the potential evaporation with respect to surface temperature substantially increases when omitting the temperature dependence of $C_h$. To summarize, the highest effect of surface temperature on the heat transfer coefficient and potential evaporation is close to neutrality. Further, the impact of $T_s$ on $C_h$ and, consequently, the potential evaporation,
can not be neglected.

The results found in the preceding section provide an estimate for the change in the latent heat flux due to albedo variations. It has been shown that a reduction of the surface albedo by 0.1 raises the surface temperature by \( \sim 0.2^\circ\text{C} \). This warming corresponds to an increase in the latent heat flux by approximately 1 W m\(^{-2}\) when assuming the same sensitivity as for potential evaporation, i.e. a 1°C warming of the surface increases the latent heat flux by 5 W m\(^{-2}\). The simulated value is approximately double the estimated value (Fig. 4.29), which is within the range of uncertainty. The negative impact of the surface albedo on the sensible heat flux is caused by the cold bias, due to less net solar radiation and enhanced stability, and thus, less turbulence.

The surface albedo hardly interacts with the ground heat flux in ECHAM (Fig. 4.29), since soil temperatures respond quickly to variations in the surface temperature. In EM, in contrast, the ground heat flux increases substantially with higher albedos. Its sensitivity is in excellent agreement with the value calculated by using simplified equation of the ground heat flux \( G_{MB} \) at a soil depth equal to \( \Delta z_h \) (cf. Section 4.2.1 and Eq. 4.10). A change of the annual background surface albedo by 0.05 yields a decrease of 0.1°C in the surface temperature (Fig. 4.29). Applying Equation 4.10, this leads to a change in the ground heat flux \( G_{MB} \) of -0.32 W m\(^{-2}\).

Transpiration strongly decreases with enhanced reflected solar radiation. In addition to the stability of the boundary layer, the transpiration parameterization in ECHAM is dependent on the photosynthetically active radiation (PAR), which causes significant variations in the stomatal resistance of the canopy. Furthermore, PAR and thus transpiration is reduced due to higher albedo values. The model comparison shows that the stomatal resistance is a key factor for transpiration: The sensitivity of transpiration to albedo in ECHAM is more than three times higher than in EM where the algorithm for transpiration does not include any dependence on PAR.

The sensitivity of annual bare soil evaporation to surface albedo has opposite signs in ECHAM and EM (Fig. 4.29). EM simulates the expected effect: Higher albedos lead to less energy available for evaporation and thus, less bare soil evaporation. The positive effect of surface albedo on bare soil evaporation in ECHAM is presumably related to the relative humidity \( h \), which greatly determines the bare soil evaporation in ECHAM (cf. Eq. 4.41). The impact of a weaker turbulence owing to higher stability is likely to be overcompensated by an increase in the relative humidity caused by a higher soil moisture content.

Increasing surface runoff and drainage compensate for the enhanced soil moisture caused by reduced evapotranspiration (cf. Fig. 4.29). The impact of surface albedos on total water discharge is similar in ECHAM and EM, although the partitioning into the single components differs strongly in the two models. While the EM removes the additional soil water mainly by drainage processes, surface runoff plays the major role in ECHAM. These differences are attributed to the parameterization of surface runoff including a fractional saturated area (cf. Eq. 4.16) which efficiently enhances surface runoff for soils close to saturation (Section 4.3.7). The noticeable increase of drainage in EM may be induced by a too efficient infiltration of water by gravity which prevents the water from draining off close to the surface (remember that the surface runoff in EM includes water discharge from the uppermost soil layer).

Annual sensitivity ratios of \( \alpha_{sb} = 0.18 \) to \( \alpha_{sb} = 0.32 \) are given in Fig. 4.32. The bar heights are generally larger than 1, implying increasing sensitivity with increasing surface albedos. However, the bar ranges do not significantly deviate from 1, indicating that, in a first approximation, the sensitivities on an annual basis are reasonably constant.
Annual cycles

The sensitivity of most output variables to surface albedo varies significantly throughout the year. This is demonstrated in Fig. 4.31 which displays the change in nine surface variables caused by an increase in the surface albedo from 0.15 to 0.35.

The sensitivity of bare soil evaporation ($E_b$) to surface albedo is negligible during the winter months (November through March) with saturated soils. The response is similar in ECHAM and EM. During summer and autumn, however, the effect on $E_b$ is of opposite sign and is generally larger in ECHAM. This emphasizes the different structure of the parameterization in the two land surface schemes. In EM, the bare soil evaporation decreases in spite of increasing soil moisture and seems to be primarily affected by decreasing turbulence. In ECHAM, the positive impact of the surface albedo on $E_b$ is likely related to wetter soils.

In both models, the response of the surface albedo on transpiration is similar from November through to June (cf. Fig. 4.31). However, during summer and autumn, EM is far less sensitive than ECHAM. The reason is probably due to the neglect of the stomata resistance in the EM parameterization, while ECHAM includes a strong dependence on the photosynthetically active radiation (PAR) and includes the stomata resistance in its parameterization.
As discussed above, the impact of changes in albedo on the two components of runoff (surface runoff/drainage) differs in both models. While the sensitivity of drainage is higher in EM during the first six months of the year, ECHAM shows a higher sensitivity of surface runoff from June through to March. In ECHAM, the substantial change in surface runoff is mainly due to fast drainage as the soil moisture is below 90% (threshold value for slow/fast drainage) for a surface albedo of 0.15, whereas it rises above 90% when using a background albedo of 0.35. Since the soil moisture in EM is partly above the field capacity during the period from January to May and the EM-parametrization very efficiently generates drainage from oversaturated soils (Eq. 4.26), the EM shows a higher impact of surface albedo to drainage during the months of January through to May.

It is surprising that the sensitivity of soil moisture to surface albedo (Fig. 4.31) during the winter months is much higher in EM than in ECHAM, which simulates a negligible impact of surface albedo on soil moisture. Obviously, in EM, the surplus soil water induced by positive albedo changes can only be partially compensated by higher runoff fluxes. In ECHAM, however, the ‘fast drainage’ is very efficient in maintaining a constant soil moisture level independent of albedo. In summer and autumn, however, the changes compare fairly well in both models.

In EM, snow water equivalent, snow melt and snow sublimation change by ~1% when increasing the surface albedo from 0.15 to 0.35. In ECHAM, the sensitivity is approximately a factor 3 higher in January. This may be due to the EM parameterization which allows for a redistribution of heat in both the snow pack and the uppermost soil layer. Note that, the surface albedo in EM responds stronger to changes in the background albedo than in ECHAM, in spite of smaller changes in the snow water equivalent. This apparent discrepancy is caused by the substantially higher derivation of the snow cover fraction with respect to snow water equivalent in EM (see Fig. 4.2).

In ECHAM, variations in the background surface albedo hardly alter the ground heat
fluxes (Fig. 4.31) since the impact of albedo changes on surface and soil temperatures, and thus, on ground heat fluxes, are similar. In contrast to ECHAM, the sensitivity of soil heat fluxes to $\alpha_{sb}$ shows a pronounced annual cycle, the maximum being in summer due to higher absolute changes in global radiation and, thus, in surface temperatures.

The impact of (background) surface albedo on the sensible heat flux (Fig. 4.31) is similar in both models, since the response on surface temperatures is also similar (not shown). The sensitivity of latent heat flux to the background albedo are also alike, excluding during the months of May and June. Sensitivities of the single water vapour components, however, differ strongly during summer and autumn, but generally compensate each other to a large extent.

**Day time and nighttime sensitivities**

Table 4.4 lists a number of ratios between annual daytime means to annual nighttime means ($r_{nd}$) for $\alpha_{sb} = 0.15$. In the last two columns the ratio of $r_{nd}$ for a surface albedo of 0.35 to 0.15 is presented. The high values of $r_{nd}(\alpha_{sb} = 0.15)$, regarding the water vapour fluxes, are caused by the missing energy supply at night. The impact of surface albedo on runoff, which is typically less than 1, is likely to be induced by the almost missing evapotranspiration at night which makes runoff the only efficient way of reducing excessive soil water. Most values in the third and fourth column are close to 1, indicating that $r_{nd}$ depends slightly on the surface albedo, excluding the sensible heat flux. This is caused by the strong decrease in sensible heat flux during the day with increasing surface albedo due to increasing stability. In EM, the annual daytime mean of the sensible heat flux amounts to approximately 35 Wm$^{-2}$ and 10 Wm$^{-2}$ using $\alpha_{sb} = 0.15$ and $\alpha_{sb} = 0.35$, respectively. ECHAM even simulates negative sensible heat fluxes (annual means) during the day for $\alpha_{sb} > 0.3$, whereas in EM, the respective albedo is 0.42. This means that for sufficiently high surface albedos, the stability of the boundary layer tends to involve more stable than unstable conditions even during the day. The change in stability significantly influences the Bowen Ratio, defined as the ratio between sensible and latent heat flux.

**Table 4.4:** Annual daytime means divided by annual nighttime means in ECHAM and EM, using $\alpha_{sb} = 0.15$ (first two value columns). Last two columns: quotients of the ratios of day- and nighttime annual averages for the model simulation using $\alpha_{sb} = 0.15$ and $\alpha_{sb} = 0.35$.

<table>
<thead>
<tr>
<th>parameter</th>
<th>ECHAM $\alpha_{sb} = 0.15$</th>
<th>EM $\alpha_{sb} = 0.15$</th>
<th>ECHAM $\alpha_{sb} = 0.35$ / $\alpha_{sb} = 0.15$</th>
<th>EM $\alpha_{sb} = 0.35$ / $\alpha_{sb} = 0.15$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>7.7</td>
<td>10.1</td>
<td>0.94</td>
<td>0.91</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>6.9</td>
<td>11.4</td>
<td>0.94</td>
<td>0.94</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>0.3</td>
<td>-0.8</td>
<td>0.24</td>
<td>0.31</td>
</tr>
<tr>
<td>skin reservoir evap.</td>
<td>3.1</td>
<td>1.6</td>
<td>1.00</td>
<td>0.88</td>
</tr>
<tr>
<td>Transpiration</td>
<td>14.0</td>
<td>15.9</td>
<td>0.92</td>
<td>0.98</td>
</tr>
<tr>
<td>Bare soil evaporation</td>
<td>7.3</td>
<td>11.8</td>
<td>1.06</td>
<td>0.84</td>
</tr>
<tr>
<td>Skin reservoir content</td>
<td>0.8</td>
<td>0.86</td>
<td>1.04</td>
<td>1.00</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>0.95</td>
<td>0.97</td>
<td>1.03</td>
<td>1.01</td>
</tr>
<tr>
<td>Surface ground heat flux</td>
<td>-0.91</td>
<td>-0.54</td>
<td>1.00</td>
<td>0.84</td>
</tr>
<tr>
<td>Snow melt</td>
<td>2.0</td>
<td>1.1</td>
<td>0.76</td>
<td>0.78</td>
</tr>
<tr>
<td>Snow sublimination</td>
<td>0.9</td>
<td>1.6</td>
<td>0.84</td>
<td>0.96</td>
</tr>
<tr>
<td>Snow depth</td>
<td>0.5</td>
<td>0.5</td>
<td>1.01</td>
<td>1.00</td>
</tr>
<tr>
<td>Drainage</td>
<td>0.84</td>
<td>1.00</td>
<td>0.96</td>
<td>1.02</td>
</tr>
<tr>
<td>Surface runoff</td>
<td>0.84</td>
<td>0.88</td>
<td>1.00</td>
<td>1.07</td>
</tr>
</tbody>
</table>
4.3.6 Sensitivity to the snow albedo

Snow cover in Cabauw occurs infrequently and by far the most frequent snow cover occurs in January (9 days with 1 - 2 cm closed snow cover and 6 days with a patchy snow cover). Therefore, the following results are limited to January. Since the first snowfall occurs on January 7 and the last on January 21, the snow water balance must be closed for January, i.e. snowfall is balanced by the sum of snow melt and snow sublimation.

Although there is only a thin snow cover at the Cabauw site, it must be emphasized that the January in 1987 represents a typical winter month for a large part of Western Europe, including the Swiss midland.

The parameterizations applied for the snow albedo $\alpha_s$ have been discussed in Section 4.2.3. The effect of the forest ratio on surface albedo does not play a role since the forest fraction equals zero in all off-line model simulations.

Sensitivities to snow albedo in January

The key difference between ECHAM and EM in computing $\alpha_s$ is associated with substantial differences in the computation of the snow cover fraction, with the largest differences obtained for thin snow packs. This is illustrated in Fig. 4.33 which gives the ratio between the derivatives of the total surface albedo $\alpha_{surf}$ with respect to $\alpha_s$ in ECHAM and EM, using the isolated model equations. Assuming a typical snow depth of 1 cm at Cabauw, corresponding to a water equivalent of $S_n = 2 - 3$ mm, the plot results in a ratio of approximately 0.3, i.e. the EM provides a three times higher sensitivity of $\alpha_{surf}$ to $\alpha_s$ than is found to be the case in ECHAM. The model experiments suggest that the different computation of the snow cover fraction is the main difference between the two models: $\alpha_{surf}$ increases in ECHAM from $\sim 0.21$ to $\sim 0.29$ using $\alpha_s = 0.5$ and $\alpha_s = 0.9$. The corresponding values in EM equal $\alpha_{surf} = 0.31$ and $\alpha_{surf} = 0.52$, respectively, implying the magnitude in the sensitivity in EM to be approximately three times higher than in ECHAM. This means that the simulated sensitivity can be explained by the sensitivity of the isolated model equations for $\alpha_{surf}$. This finding was further verified by incorporating Equation 4.31 (used in ECHAM) into EM which led to a close agreement of the sensitivities between both models.
Figure 4.34 illustrates percentage changes $\Delta$ (Eq. 4.57) of 15 output variables for an increase in the snow albedo from 0.6 to 0.7. Relative changes are shown since absolute changes are rather small owing to low shortwave radiation fluxes as well as to generally thin or vanishing snow packs at the Cabauw site in January. The sensitivities between both models differ largely. The sensitivities in EM are mostly higher, chiefly due to the differing calculation of the snow cover fraction, as outlined in the previous section. In general, the findings presented in Section 4.3.5 can be applied to snow albedo as well. It is obvious that during snow periods, an increase in the snow albedo increases the total surface albedo, thereby reducing solar absorption and cooling the surface. Net shortwave radiation decreases between approximately 0.5 Wm$^{-2}$ and 1.5 Wm$^{-2}$ when increasing $\alpha_s$ by 0.1 in both ECHAM and EM, respectively. In a three-dimensional model experiment, the impact of changing snow albedo would likely be significantly higher due to, e.g. the positive snow-albedo feedback (higher albedo induces lower temperatures and thus, more precipitation falling as snow, further enhancing the albedo) which is almost eliminated by the prescribed temperature on the first model level used as a forcing variable in all the simulations.

Transpiration is more sensitive to $\alpha_s$ in ECHAM (Fig. 4.34) than in EM due to the dependence of the transpiration algorithm in ECHAM on shortwave radiation. Model simulations indicate, for an increase in the photosynthetically active radiation ($PAR$) by 5 Wm$^{-2}$, a change in the transpiration rate of the order of 10%, while the soil moisture is not a limiting factor for transpiration in January. Changes in transpiration can also be caused by altered surface temperatures and thus, stability, i.e. the cooler the surface, the smaller the turbulence and, therefore, the water vapour fluxes. This explains the decrease in transpiration, bare soil evaporation and snow sublimation with increasing snow albedo. It is evident that cooler surfaces yield an enhanced downward component of the sensible heat flux due to increased stability.

The reduced evapotranspiration for enhanced snow albedos implies a higher soil moisture, leading to more runoff, applicable to both models. The distribution of the additional runoff into its components (surface runoff, drainage), however, differs strongly. Since EM allows for fast infiltration of water into the lower soil layers, additional soil water enhances drainage more than it does surface runoff. Surface runoff is, however, strongly influenced by changing soil moisture contents when soils are close to saturation.

The large differences in the sensitivity of ground heat fluxes in ECHAM and EM are mainly attributed to the prescribed deep soil temperature $T_{dl}$ in EM.

Since no snow is modeled at the beginning and at the end of January, the positive impact of snow albedo on the snow water equivalent due to lower surface temperatures is associated with a higher melting rate. Snow sublimation is reduced with increasing snow albedo primarily due to weaker turbulence.

**Diurnal cycles**

Fig. 4.35 illustrates the modifications in the mean diurnal cycles of January caused by changes in the surface albedo. The two thicker lines, representing the ECHAM variables, are almost identical, implying that the sensitivity is nearly independent of the snow albedo, the exception being snow melt where a phase shift of two hours is found in the late evening. The reason for this feature could not be ascertained but this might be a hint to a deficiency in the snow parameterization.

In EM, the changes in the diurnal cycles are generally higher than in ECHAM. In addition, the two thin lines in Fig. 4.35 often differ to a significant extent, i.e. the sensitivities to snow albedo are not constant. This primarily applies to surface temperature and
variables which are directly dependent on surface temperature, e.g. ground heat fluxes and evapotranspiration. Note that the changes in surface temperature and ground heat fluxes shift the phase when increasing the snow albedo on which the calculations are based. This, again, is related to the prescribed deep soil temperature $T_d$.

In EM, a higher $\alpha_s$ leads to more snow melt between 23 UT and 9 UT, whereas during the day, a decrease is found. The reason may be that during the day, a stronger cooling occurs for higher reflectivities whereas at night, temperatures are hardly influenced.

It is noteworthy, that the sensitivities at noon are often an order of magnitude larger than monthly averages, i.e. the latent heat flux changes by $6 \text{ Wm}^{-2}$ (EM) and the net shortwave radiation by about $15 \text{ Wm}^{-2}$ when altering $\alpha_s$ by 0.2.

### 4.3.7 Sensitivity to the maximum soil water content

The soil moisture content is crucial for the climate of a region as it determines, to a large extent, the development of vegetation as well as the water vapour fluxes and runoff. The partition between latent and sensible heat fluxes and thus, the Bowen ratio, is greatly influenced by the soil moisture. Enhanced evaporation due to higher soil moisture leads to a cooling of the surface. Furthermore, circulation and rainfall is substantially affected by the soil moisture (Walker and Rowntree, 1977; Schar et al., 1999). Soil moisture strongly changes the heat capacity of the soil due to the large heat capacity of water. An enhanced soil wetness reduces the surface albedo and consequently, results in a higher net shortwave radiation.

In the following, the expression 'maximum soil water content' will be explained in detail. In ECHAM, it is obvious to equate the maximum soil water content with $W_{\text{max}}$. However, the definition in EM allows two possibilities to define the maximum soil water content, namely, the field capacity ($FC$) or the soil porosity ($PV$). Since the field capacity in
EM can be exceeded during several days, the soil porosity was defined as the maximum soil water content. The relative soil moisture is, however, computed relative to the field capacity according to most authors, leading to relative soil moistures above 1. The other parameters needed in EM to describe the soil (permanent wilting point and air dryness point) are readapted to conserve equal ratios between the soil parameters as for the soil type 'loamy sand' which applies to the Cabauw site.

Discussion of sensitivities investigating the isolated parameterization equations

Whereas ECHAM includes only one layer for water (bucket model), the EM incorporates two or more different layers for water. This implies that in ECHAM, all the evaporation and runoff processes have to be parameterized using a single layer for soil moisture (cf. Section 4.2.2). This is physically incorrect because the evapotranspiration and runoff originate at different depths in the soil. The bare soil evaporation is determined by the soil moisture in a thin layer close to the surface (cf. Section 4.6) while drainage processes originate from lower soil layers. The transpiration rate, on the other hand, depends on the root depth of plants which draw water from the soil. Therefore, a more physically based
description of these processes is possible by introducing several soil layers for water (as in EM).

In the following section, the sensitivity of different variables to the soil moisture content is investigated. The parameterization equations generally include the relative soil moisture $W_{srel}$ rather than the absolute amount of water ($W_s$). The second addend in Eq. 4.15, parameterizing fast drainage, can be easily transformed into $c(W_{srel} - 0.9)d$, where $c$ is a constant. The water stress factor $F(W_s)$, as used in the control simulation, can be written as $F(W_s) = 3.33 W_{srel}^2 - 0.66$. The evaporation efficiency $E$ (Eq. 4.44) as well as the relative humidity at the surface $h$ (Eq. 4.41) are also functions of $W_{srel}$.

The sensitivity of surface runoff and drainage (ECHAM) to $W_{srel}$ increases strongly when approaching the field capacity (Figs. 4.36 and 4.37). By changing $W_{srel}$ from 90% to 100%, drainage and surface runoff increase by a factor of approximately 25 and 2, respectively. This means that the partition of total runoff into drainage and surface runoff is strongly expected to change when modifying the soil moisture. While the derivative of drainage with respect to the water content $W_{srel}$ increases by a factor of 3 when changing the relative soil moisture from 90% to 100% (Fig. 4.37 b), the corresponding factor for surface runoff amounts to ~7 (Fig. 4.36 b). The sensitivity of drainage to $W_{srel}$ flattens with increasing $W_{srel}$, while the sensitivity of surface runoff to $W_{srel}$ shows, in a first approximation, an exponential increase.

The relative humidity $h$ at the surface is represented by a cosine function of the relative soil water content (Eq. 4.40), leading to the highest sensitivity of bare soil evaporation for $W_{srel} = 50\%$ and decreasing sensitivities for drier or wetter soils. This statement applies only when neglecting the water vapour gradient within the boundary layer. The evaporation efficiency $E$ (Eq. 4.44) contains the water stress function $F(W_s)$ which linearly increases from the permanent wilting point to the critical value. Figure 4.38 shows that the evaporation efficiency increases between 20% (default value in the control simulation permanent wilting point) and 50% (critical value). The impact of $W_{srel}$ on $E$ decreases with increasing relative soil moisture content in the range $W_{srel} = [20\%, 50\%]$, but vanishes else outside this range.
Figure 4.37: Drainage and the derivative with respect to $W_{rel}$ (Eq. 4.15). All parameters as in the ECHAM control simulation.

Figure 4.38: Evaporation efficiency $E$ and its derivative with respect to $W_{rel}$. Parameters: $C_h = 0.005$, $v(30 \text{ m}) = 5 \text{ ms}^{-1}$, PAR (photosynthetically active radiation) = 77 Wm$^{-2}$ (mean of June at Cabauw). Other parameters as in the ECHAM control simulation.

Regarding the EM, the formulae containing the soil water content refer to either the soil moisture content of the upper or the lower soil layer (cf. Eq. 4.42). The bare soil evaporation increases as a quadratic function for soil water content in the uppermost soil layer. Thus, the sensitivity of the bare soil evaporation increases linearly with $W_s$ between the air dryness point and the field capacity. The transpiration rate in EM from both layers is also represented by a quadratic function of soil water (Eq. 4.45). However, the range
for non-zero transpiration and non-zero bare soil evaporation differ: Below the permanent wilting point, plants no longer transpirate, and above the turgor loss point, transpiration works at its optimum.

It is difficult to compute the sensitivities of runoff processes to \( W_s \) (in EM) by using the isolated equations since the Darcy equation (Eq. 4.27) is a differential equation. Moreover, the coefficients used in this equation are dependent on soil moisture.

Neglecting the water transport between the soil layers, the lateral runoff (in EM) out of the soil layers increases linearly with the soil moisture content between the field capacity and the soil porosity (Eq. 4.26) which yields a constant sensitivity of drainage to \( W_s \).

**Annual sensitivities to \( W_s \) using results from off-line simulation**

The maximum soil moisture content was varied in the ECHAM simulations by between 0.2 m and 0.4 m, with increment steps equal to 0.05 m. The same values were selected in EM for the soil porosity, maintaining identical ratios between all soil parameters as for loamy sand, the soil type used in the control simulation. Unfortunately, the different soil structure in ECHAM and EM complicates the comparison between both models.

In ECHAM and EM, the (annual) soil water content increases significantly with increasing maximum soil water content \( W_{smax} \). For brevity, the soil porosity is abbreviated to \( W_{smax} \) in the course of this chapter. The sensitivity of the soil moisture to \( W_{smax} \) is fairly constant over the entire (investigated) range and amounts to \( \sim 6 \) mm and \( \sim 8 \) mm in ECHAM and EM, respectively, when increasing \( W_{smax} \) by 1 cm. Note that the relative soil moisture remains almost constant. This is of large importance since a number of processes as transpiration, bare soil evaporation and runoff, rather contain \( W_{srel} \) than the absolute amount of soil moisture. However, since the global radiation and thus, the turbulent heat fluxes, show a pronounced annual cycle, it is inappropriate to conclude that slight sensitivities in the annual relative soil moisture content lead to adequate changes in evapotranspiration and runoff.

Analyzing annual cycles on a monthly basis (Fig. 4.39) reveals that, for both models, the relative soil moisture decreases for enhanced values of \( W_{smax} \) in winter and spring, whereas during summer and autumn, the opposite applies. This results in lower annual amplitudes of \( W_{srel} \), when \( W_{smax} \) is enhanced which is physically reasonable: The deeper the soil and the more water that can be stored, the smaller the relative changes in the soil moisture. Note that the apparent inconsistency between the differences in the absolute and relative soil moisture content is due to the different definition of these variables, as was outlined earlier: The maximum soil water content in EM refers to the soil porosity but \( W_{srel} \) is computed using the field capacity.

Drainage in summer is only slightly influenced by changing \( W_{smax} \) because in ECHAM, the relative soil moisture which is lower than the threshold of \( W_{srel} = 90\% \) which allows for 'fast drainage'. During the period where soils are close to saturation, the change in drainage is strongly dependent on the number of days with \( W_{srel} > 90\% \), where there is an occurrence of 'fast drainage' (ECHAM). This is clearly demonstrated in Fig. 4.39. Decreasing relative soil moisture is generally related to reduced drainage. The same applies, although less evidently, for surface runoff. In ECHAM, changes in surface runoff follow well the alterations in relative soil moisture since the parameterization of surface runoff does not include a discontinuity for \( W_{srel} = 90\% \), as has been incorporated in the parameterization for drainage.

It is surprising that, in EM, surface runoff decreases with increasing soil moisture content in summer and autumn, while the evapotranspiration is significantly enhanced (Fig. 4.39). Obviously, the additional water storage capacity favours evaporation over surface runoff.
This discrepancy can be clarified by analyzing the evolution of transpiration since changes in transpiration largely contribute to the change in evapotranspiration. The key parameter in the parameterization equation of transpiration (Eq. 4.45) is the factor $\beta_{B,K}$. The turgor loss point $TLP$ (which determines $\beta_{B,K}$) ranges, according to model experiments, between 70% - 75% and > 95% for low and high global radiation. Since transpiration increases quadratically with the soil moisture between the permanent wilting point and the turgor loss point, and $W_{srei}$ is generally lower than $TLP$ during summer and autumn, a positive bias in $W_{srei}$ greatly affects transpiration. The simple equation used to parameterize the transpiration allows the model to produce excessive transpiration, but substantially reduces surface runoff. In contrast to EM, the transpiration, simulated with ECHAM, hardly changes with soil moisture, the influence of the net shortwave radiation being more important for possible changes.

The partition into surface runoff and drainage (EM) may also be influenced by the following effect. Detailed model investigations have revealed that the water exchange between the soil layers is often not determined by the Darcy equation but rather by the empirical constraint as follows: the maximum water flux between the soil layers amounts to 10% of the soil porosity per timestep. This induces a more rapid water flux from the upper to the lower soil layer for enhanced soil porosity and therefore, enhances drainage (lateral outflow.

**Figure 4.39: Impact of an increase in $W_{smax}$ from 0.2 m to 0.3 m on surface variables.**

*Solid: ECHAM, dashed: EM.*
Figure 4.40: Relative and absolute soil water content for different maximum soil water contents (EM, monthly averages). a) Soil moisture content upper soil layer, b) relative amounts; c) Soil moisture content lower soil layer, d) relative amounts.

from the lower soil layer) and reduces surface runoff (which includes lateral outflow of the upper soil layer). This may play a major role in reducing surface runoff when increasing $W_{srel}$.

Annual bare soil evaporation depends primarily on the soil moisture conditions during the warmer season, while anomalies in $W_s$ during winter and spring with soils close to saturation and generally low solar radiation fluxes are of less importance. In both the ECHAM and EM, bare soil evaporation increases significantly with enhanced $W_{smax}$ in summer, the changes in EM being substantially smaller than in ECHAM.

Enhanced latent heat fluxes are related to higher values of $W_{smax}$ (Fig. 4.39). The larger sensitivity of transpiration to $W_{smax}$ in EM also implies higher impacts on the latent heat fluxes, while the effect of skin reservoir and bare soil evaporation is of less importance. The increased water vapour flux leads to a cooling of the surface and to a distinct decrease of the surface temperature with increasing $W_{smax}$. However, the lower albedo, caused by increased soil moisture, compensates partially for the cooling effect, but the model simulations show that the effect of evaporation clearly exceeds the albedo influence. The positive anomalies in the latent heat flux are accompanied by negative changes in the sensible heat flux (not shown) of almost the same amount.

Since in ECHAM, each grid element is assigned a prescribed background albedo, the surface albedos remains constant. The EM albedo, on the other hand, is influenced by the soil moisture of the uppermost soil layer and is reduced when the soil moisture increases. Thus, the albedo becomes lower during summer due to an enhanced soil moisture content. During winter, the impact of soil moisture on surface albedo can be neglected because soils
are saturated. During this season, the negative sensitivities of $e_{surf}$ to $W_{smax}$ are likely due to the slightly warmer temperatures and therefore smaller snow water equivalent.

In general, snow conditions in both models are only minimally influenced by changing soil properties. It may be of some importance that the sensitivity of snow depth (and thus, snow melt and snow sublimation) is more than a magnitude larger in EM than in ECHAM, the only reason being the different sensitivities in surface temperatures and thus, snowfall, which is confirmed in Fig. 4.39: ECHAM shows no sensitivity of surface temperature to $W_{smax}$ during winter.

Figure 4.40 demonstrates how the annual cycles of $W_s$ differ between the two soil layers in EM. The annual amplitude of the relative water content in the lower soil layer is distinctly larger than in the upper layer. Note the significant oversaturation of soils during the winter season. This emphasizes again that in EM, the soil moisture content is allowed to exceed the field capacity, in contrast to ECHAM. Observational data confirms that, in western Europe, oversaturated soils during the winter season occur only rarely.

4.4 Assessment of the surface climate in ECHAM and EM

This chapter compares a number of surface variables which are simulated using the ECHAM and EM control simulation (cf. Section 4.3.1). The comparison further assesses some interesting model characteristics related to the forced temperature and the height of the first atmospheric model level. A comparison with observational data is provided at the end of this chapter.

Comparison between ECHAM and EM

The annual means of 23 surface variables are listed in Table 4.5. It is obvious that some variables differ significantly between ECHAM and EM. The following differences in the water vapour fluxes were found: The annual evapotranspiration computed by the EM is approximately 44 mm or 8% higher than in ECHAM. However, the single components of evapotranspiration, such as bare soil evaporation, skin reservoir evaporation and transpiration, often deviate significantly. The most pronounced difference, however, is related to the skin reservoir evaporation: ECHAM computes a 2.5 times higher value than EM. This result is surprising since the maximum skin reservoir content ($W_{lmax}$) in EM (Eq. 4.21) equals 3 mm, whereas in ECHAM, $W_{lmax} = 0.04$ mm, assuming LAI = 2. Despite this imbalance, the ECHAM skin reservoir usually contains more water than the respective reservoir in EM, the reason being the very efficient infiltration into the soil, using Equation 4.23. It is physically not reasonable to compute the infiltration rate from the skin reservoir of vegetated area (leaves) in the same way as from bare soil. The rapid infiltration of the skin reservoir’s water in EM leads to a markedly larger evaporation rate from the skin reservoir in ECHAM, the difference being 58 mm/year. In comparison to ECHAM, the excessive infiltration in EM leads to distinctly higher annual transpiration and bare soil evaporation. The comparison on a monthly basis reveals that transpiration is slightly higher in the EM from September through March, whereas in spring time (April - June) the difference is markedly higher, being about 20 mm/month for this period (Fig. 4.41). This characteristic can be related to the factor $\beta$ in Eq. 4.45: since soil moisture is above the value where water stress is initialized (turgor loss point), $\beta$ equals 1. Furthermore, the EM parameterization neglects the impact of shortwave radiation on transpiration. Thus, transpiration is at its maximal rate in spring, whereas in ECHAM this evaporation flux is efficiently reduced by fairly low global radiation. In the episode from July through September, ECHAM simulates slightly higher transpiration rates than
EM, the main reason being the rapid decrease of the factor $\beta^2$ in Equation 4.45 when the soil moisture falls below the turgor loss point (mainly during the day). In addition, high incoming shortwave radiation favours enhanced transpiration rates when using the parameterization developed for ECHAM. This results in largest transpiration rates during July when global radiation is at its maximum. Bare soil evaporation ($E_b$) is of lesser importance for the total water vapour flux since the grid box which represents Cabauw, is essentially covered with vegetation. The evolution of bare soil evaporation substantially differs between both models. During winter and spring (November - April), when the soils in both models are generally saturated, the water vapour fluxes from bare soil are similar. In May, however, the evaporation in ECHAM is dramatically reduced. In late summer, the evaporation rate approaches values such as those simulated in winter time. The curve simulated with EM, on the other hand, depicts a pronounced annual cycle with a maximum during summer, as is to be expected over western Europe (Fig. 4.41). This results in a ratio between annual $E_b$ (EM) and annual $E_b$ (ECHAM) of 0.59. In late summer, this ratio is reduced to approximately 0.2. The annual mean of the sensible heat flux in EM is equal to 8.8 Wm$^{-2}$, which is considerably higher than in ECHAM which has a value of 2.9 Wm$^{-2}$. Largest differences are simulated in summer as well as for the period from February through to April (Figure 4.41).

Model simulations have revealed that the difference in the sensible heat fluxes, simulated with the off-line version of ECHAM and EM, are significantly reduced when forcing the two models with identical surface temperatures $T_s$ (Figure 4.42). This signifies that variations in the surface temperature affect the turbulent heat fluxes substantially.

Figure 4.42 illustrates that the deep soil temperature ($T_U$) also plays a major role in the magnitude of the sensible heat flux in EM. Changes in the sensible heat flux due to changed temperature conditions seem, however, to be less affected by the ground temperature than by the surface temperature. This is reasonable since the surface temperature is a key parameter when determining the stability conditions within the boundary layer. Simulated
Figure 4.41: Monthly means of transpiration, bare soil evaporation, sensible heat flux and of surface ground heat flux. Solid: ECHAM, dashed: EM.

Figure 4.42: Simulated monthly means of sensible heat flux. 1: Control simulation, ECHAM; 2: Control simulation, EM; 3: Simulation with EM, but surface temperature as in ECHAM; 4: Simulation with EM, but $T_d$ (deep soil temperature) as in ECHAM.

ground heat fluxes (Figure 4.41) also differ significantly in both models. While during the winter, differences are negligible, strong differences occur from April through to October. Whereas the ground heat flux simulated by ECHAM shows the expected evolution, with downward fluxes during summer and an upward directed flux in winter, the fluxes generated by EM are likely afflicted with errors: The monthly ground heat fluxes are usually directed into the ground, conflicting with the observations and the ECHAM simulation.
(Figure 4.43). This feature is attributed to an overestimation in the forced temperature $T_u$ during summer, which generates an increasing temperature with increasing soil depth during summer and which, in turn, produces an upward directed heat flux. Although the expressions for calculating the ground heat fluxes in ECHAM and EM are based on strictly different methods (Eqs. 4.6 and 4.7), the simulated ground heat fluxes are in nearly perfect agreement when forcing the two models with identical soil temperatures (Figure 4.43). Both the ECHAM and EM generate high variability in the monthly means of drainage and surface runoff. In some months, the simulated runoff agrees reasonably well in both models, while in other months large differences are generated. The analysis, on an annual basis, reveals both drainage and surface runoff to be approximately 20 mm/year lower in EM compared to ECHAM. The most pronounced differences occur in winter and spring, while during summer, the deviation between both models is generally small. The increased drainage in winter, compared to EM, is mainly a result of 'fast drainage' (Eq. 4.21), which strongly enhances the drainage in ECHAM for relative soil moistures above 90%.

The surface runoff in EM includes the runoff from the uppermost soil layer (cf. Section 4.2.2). Off-line model simulations show that the runoff from the uppermost soil layer contributes more than 99% to the total surface runoff, all other components being negligible. This feature is attributed to a very efficient infiltration of water into the soil. During winter, ECHAM typically simulates more surface runoff than EM. This may be related to the ECHAM parameterization which allows for a fractional saturated area before the soil is completely saturated (Equation 4.16), i.e. surface runoff is initialized before the soil moisture reaches its field capacity: For a relative soil moisture of 95% and 99%, the saturated fraction equals ~7% and 11%, respectively.

The soil moisture content (not shown) can not be directly compared, since ECHAM is based on a bucket model, while EM sustains several layers for soil moisture. Since EM allows for the soil moisture to exceed the field capacity, the relative soil moisture $W_{s,rel}$, i.e. the soil moisture divided by the field capacity (for a detailed discussion the reader is referred to Section 4.3.7) occasionally exceeds 1, predominantly in winter. Summer values

![Figure 4.43: Simulated monthly means of surface ground heat flux. 1: Control simulation, ECHAM; 2: Control simulation, EM; 3: Simulation with EM, but $T_u$ as in ECHAM.](image-url)
of $W_{s,rel}$ are typically close to 0.6 and 0.9 in ECHAM and EM, respectively, which signify that the summer drying of the soil generated by ECHAM is absent in the EM control simulation.

Radiation fluxes simulated by ECHAM deviate little from the fluxes generated by the EM experiment due to the atmospheric forcing of global radiation and downward longwave radiation ($LW_{\downarrow}$). Differences in $LW_{\uparrow}$ are related to different surface temperatures according to the Stefan-Boltzmann-Law: A 1°C increase in surface temperature yields $\Delta LW_{\uparrow} = 4\ Wm^{-2}$ and $\Delta LW_{\uparrow} = 6\ Wm^{-2}$ for $T_s = -10^\circ C$ and $T_s = 30^\circ C$, respectively. Annual $\Delta LW_{\uparrow}$ in EM exceeds the respective ECHAM flux by approximately $1.5\ Wm^{-2}$.

Figure 4.44: Simulated differences in turbulent and ground heat fluxes as well as surface temperature: ECHAM simulation with the lowest atmospheric level at a height of $z = 20\ m$ minus the control simulation ($z = \sim 30\ m$).

Net shortwave radiation is exclusively influenced by the surface albedo. During the snow free period (March to December), the modeled surface albedos hardly differ from one another, while during winter, substantial differences between the surface albedo simulated in both models, have been detected. In January, the mean albedo in ECHAM and EM equals approximately 0.27 and 0.4, respectively. This is due to (i) a larger grid fraction covered with snow in EM (Figure 4.2) and, less important, (ii) a slightly higher snow water equivalent in EM, caused by a lower surface temperature. The albedo difference, discussed above, leads to a substantial deviation in the net shortwave radiation in January of $5\ Wm^{-2}$ (or approximately 20%).

Comparison of simulated surface variables with observations

The following output variables are available as both simulated and observed values:

- surface temperature
- latent heat flux
- sensible heat flux
- ground heat flux at a depth of 6.5 cm
- net shortwave radiation
- net longwave radiation
- surface ground heat flux.
It should first be mentioned that the measurement height (20 m), on which the data at the Cabauw site are collected differs from the lowest atmospheric model level in ECHAM (~30 m), leading to noticeable deviations (Fig. 4.44).

The determination of surface temperatures is difficult, with regard to both modeling and observation. In ECHAM, it is assumed that, for snow free condition, the surface temperature and temperature of the uppermost soil layer of 6.5 cm depth are identical. Other studies suggest the introduction of a very thin uppermost soil layer to compute a more realistic skin temperature (e.g. Betts et al., 1993; Viterbo and Beljaars, 1995). In EM, the surface temperature and uppermost soil layer temperature generally differ. The measurement of surface temperature is also a problem. Studies using the Cabauw data usually identify the surface temperature with the temperature measured at a depth of 2 cm. The effective skin temperature can also be computed via the Stefan-Boltzmann-Law. A comparison of the simulated and observed temperatures, as well as the temperature derived from LW↑ is shown in Figure 4.45.

It is evident that the observed temperatures differ quite substantially, i.e. the 2 cm temperature is generally 2 - 3°C lower than the temperature derived from the observed LW↑ (T_{eff}). In addition to the different definition, uncertainties in the measurement of the downwelling longwave radiation might also contribute to this large bias. The curves of the simulated surface temperature evolve, for both the ECHAM and EM, within the range of the 2 cm-temperature and the effective skin temperature, and differ only slightly. T_{eff} is usually distinctly higher than the simulated temperatures in both the ECHAM and EM. This feature can be attributed to the discrepancy in ECHAM: that the simulated temperature represents both the soil temperature at a depth of 3.25 cm (corresponding to the middle of the uppermost soil layer), and the temperature at the Earth’s surface. It is, however, somewhat surprising that the simulated surface temperature in EM, despite its “correct” definition, provides no significant improvement. One could argue that the temperature forcing in EM is too strong but, unfortunately, EM requires, when run in an off-line mode, both T_7 and T_{30m} as forcing variables. However, since the simulated (monthly) surface temperatures generally range between the two observations, they might be reasonable.
The comparison of simulated and observed monthly values of sensible and latent heat flux, net shortwave radiation and ground heat flux (Fig. 4.46) yields the following results: the simulated latent and sensible heat flux in ECHAM are in better agreement with the observations. The sensible heat flux is for the most part overestimated in both models. The discrepancies are greater during summer and autumn due to (i) higher global radiation and the related increase of thermal instable situations, (ii) more pronounced differences in the surface temperature, and (iii) drier soils, which enhance the sensitivity (compared to winter conditions with essentially saturated soils). How the turbulent heat fluxes vary when $T_u$ in EM is forced to follow the deep-soil temperature generated in ECHAM has been investigated. This experiment reveals that the agreement of the simulated sensible heat flux (EM) with the observations is substantially improved. The improvement in the latent heat flux is, however, less significant. This reflects that in EM, the sensible heat flux is more sensitive to changes in the surface temperature than it is in the latent heat flux. The pronounced overestimation in the sensible heat flux in EM, from April through to May (Fig. 4.46), is mainly caused by an overestimation in the respective transpiration rate. The oversimplified parameterization of transpiration in EM, neglecting the global radiation, is obviously not applicable for simulating realistic transpiration rates.

The ground heat flux in ECHAM is simulated with reasonable skill, while in EM, the upward component is typically overestimated by a large degree, the main reason being the described temperature $T_u$. The ground heat fluxes compare well when forcing the two models with identical temperature conditions ($T_s$ and $T_U$) the whole year round (Fig. 4.46). This allows the conclusion, that the two algorithms used in ECHAM and EM to compute

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**Figure 4.46**: Monthly means of simulated and observed turbulent and ground heat fluxes as well as net shortwave radiation. Obs: Observation (Cabauw); CTRL$_{ECH}$: control simulation, ECHAM; CTRL$_{EM}$: control simulation, EM; EM$_{mod}$: EM, but forced with $T_s$ and $T_U$ from ECHAM.
soil heat fluxes provide nearly the same results for identical temperature conditions, even though they are substantially different.

The differences in the simulated shortwave radiation are small, but the net radiation is significantly overestimated (Fig. 4.46) due to an underestimated surface albedo in the models (assumed to be $\alpha = 0.15$). Radiation data from the Cabauw site suggest a surface albedo of approximately 0.3 (in the absence of snow), which is unrealistically high for grassland (see e.g., Roesch et al., 1997).

### 4.5 Diurnal cycles of surface variables in the ECHAM and EM control experiments

**Annual ratios between 'day' and 'night'**

In addition to the results derived on a monthly basis, the question arises how the simulated surface variables compare to each other on shorter time scales. More information can be obtained by comparing either diurnal cycles, or the components averaged over the day and night. The distinction between 'day' and 'night' has been assessed using the global radiation ($SW_\downarrow$), the threshold value being defined as 2 Wm$^{-2}$.

The annual ratios, computed as the daytime means divided by the nighttime means, of a number of simulated surface variables, are compared in Table 4.6. We refer to this ratio as $r_{dn}$. Thus, $r_{dn} = 1$ indicates that the annual means for day and night are identical, while a negative value indicates that the flux direction during the day is opposite to that at night.

<table>
<thead>
<tr>
<th>parameter</th>
<th>$r_{dn}$ (ECHAM)</th>
<th>$r_{dn}$ (EM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>9.0</td>
<td>11.1</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>8.0</td>
<td>12.1</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>-1.6</td>
<td>-2.6</td>
</tr>
<tr>
<td>Evaporation from the skin reservoir</td>
<td>3.4</td>
<td>1.8</td>
</tr>
<tr>
<td>Transpiration</td>
<td>15.8</td>
<td>16.3</td>
</tr>
<tr>
<td>Bare soil evaporation</td>
<td>7.8</td>
<td>14.0</td>
</tr>
<tr>
<td>Skin reservoir content</td>
<td>0.77</td>
<td>0.86</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>0.93</td>
<td>0.96</td>
</tr>
<tr>
<td>Surface ground heat flux</td>
<td>-0.90</td>
<td>-0.64</td>
</tr>
<tr>
<td>Ground heat flux, depth 6.5 cm</td>
<td>-0.91</td>
<td>-0.47</td>
</tr>
<tr>
<td>Snow melt</td>
<td>2.2</td>
<td>1.5</td>
</tr>
<tr>
<td>Snow sublimation</td>
<td>1.1</td>
<td>1.6</td>
</tr>
<tr>
<td>net shortwave radiation</td>
<td>288</td>
<td>240</td>
</tr>
<tr>
<td>net radiation</td>
<td>-3.1</td>
<td>-3.4</td>
</tr>
<tr>
<td>runoff due to drainage</td>
<td>0.88</td>
<td>0.97</td>
</tr>
<tr>
<td>surface runoff</td>
<td>0.84</td>
<td>0.82</td>
</tr>
<tr>
<td>total precipitation</td>
<td>0.90</td>
<td>0.90</td>
</tr>
<tr>
<td>albedo</td>
<td>0.96</td>
<td>0.92</td>
</tr>
</tbody>
</table>

It is evident that the evaporation fluxes predicted during the day are much higher than the respective fluxes at night (Fig. 4.47). The nighttime evaporation rate is in the order of ten times smaller than the rate during the day, mainly due to the pronounced diurnal cycle of global radiation which essentially controls the transpiration rate. A minor contributing factor is also the diurnal cycle of wind speed. Note, that the accumulated evaporation depends on the length of the 'day' and 'night': during summer, the day covers a longer period than the night, during winter this characteristic is reversed. However, the main
characteristics presented in this chapter are the same when converting the means to equal time periods.

\( r_{nd} \) for bare soil evaporation is distinctly larger in EM than in ECHAM, i.e. \( r_{nd} \) is 7.8 in ECHAM and 14.0 in EM, indicating that the evaporation rate simulated during the day is, in EM, distinctly higher than in ECHAM (bare soil evaporation at night compares very well in both models). This is due to (i) the excessive soil dryness in ECHAM in late summer, which strongly reduces the evaporation rate, and (ii) a higher transfer coefficient \( C_h \) in ECHAM compared to EM, caused by differences in the von-Kármán-constant \( k \) (cf. Section 4.2.5).

\( r_{nd} \) for annual transpiration is close to 16 and has the same magnitude in both model simulations, implying that transpiration comes to an end when the global radiation approaches zero. It can, therefore, be argued that the simple algorithm used to compute the transpiration rates in EM, is efficient enough to capture correctly the ratio between transpiration during both day and night. \( r_{nd} \) for the evaporation from the skin reservoir significantly differs between ECHAM and EM. This is probably on account of the efficient infiltration of the skin reservoir water in EM. ECHAM, on the other hand, allows for a substantial part of the water to remain on leaves from where it evaporates at the potential rate. Analyzing \( r_{nd} \) on a monthly basis reveals that, in October, \( r_{nd} \) becomes less than 1, due to unevenly distributed precipitation: 80 mm fell at night and 15 mm during the day.

The soil water content does not follow the short-term fluctuations of precipitation and evapotranspiration and, thus, the differences between day- and nighttime averages are small. The relative difference becomes larger when restricting the investigation to the uppermost soil layer in EM (to a depth of 10 cm) and equals approximately 5%, corresponding to a \( \sim 1 \) mm water column. \( r_{nd} \) of the skin reservoir deviates more noticeably

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**Figure 4.47:** Simulated monthly mean day- and nighttime averages of latent and sensible heat flux, ground heat flux and snow melt.
from unity due to (i) the small amount of water stored in the skin reservoir (annual average is some tenths of a millimeter) and (ii) the lower potential evaporation at night.

The monthly means (averaged over daytime) of the sensible heat flux are directed upward from March through to September whereas during late autumn and winter the flux direction is reversed (Fig. 4.47). At night, simulated monthly averages of the sensible heat flux are directed to the ground throughout the year. This is in agreement with the evolution of the net radiation which has, at night, a downward component of between 30 Wm$^{-2}$ and 50 Wm$^{-2}$ (monthly basis), leading to a cooling of the surface and the evolution of a stably stratified boundary layer. During the day, in contrast, the positive net radiation (related to the Earth) leads to primarily unstable and convective conditions. Absolute $\tau_{nd}$, based on annual means, is distinctly larger in EM, due mainly to significant differences in the simulated heat fluxes during the day (Fig. 4.47). Monthly means, averaged during the nighttime periods, show almost no annual cycle, whereas the averages related to the day episodes depict a marked annual cycle which follows the cycle of net shortwave radiation.

The direction of the ground heat flux is also reversed between day- and nighttime (Figure 4.47). Maximal monthly mean fluxes during the day are simulated in April, whereas the maximal (absolute) value at night is generated in July. While ECHAM generally computes higher mean values during the day, the opposite applies to the (absolute) fluxes at night. From April through September, the differences between ECHAM and EM are more pronounced while in winter the fluxes compare well. The direction of the ground heat flux can be easily related to the soil temperatures: the simplified flux equation in EM (Eq. 4.10) computes, for an increasing temperature with soil depth, an upward directed ground heat flux. These conditions typically occur during periods with a negative net ra-

![Figure 4.48: Comparison of monthly mean diurnal cycles of observed and simulated surface temperature.](image)
Radiation balance (in winter or at night), leading to an upward transport of the heat stored in the ground.

Significant differences between ECHAM and EM are also found in the simulated monthly snow melt rates. While in ECHAM, most snow melts during the day in January (with a day length of \(~9\) hours), the EM, on the other hand, generates more snow melt at night. This demonstrates that the EM parameterization incorporates a more efficient melting of snow at night. This may be due to snow melt at the lower boundary of the snow deck which may be triggered through heating by a ground heat flux which is generally directed upward at night.

Runoff processes are dependent on the soil moisture content as well as on the precipitation rate and evapotranspiration. The ratio between surface runoff and precipitation increases with soil moisture, since wetter soils enhance the fraction of saturated area (cf. Eq. 4.16) and thus, surface runoff. Hence, \(r_{nd}\) of surface runoff is closely related to the corresponding ratio in the total precipitation (Table 4.6), or, in other words, precipitation and surface runoff correlate positively, the correlation coefficient on a daily and monthly basis being 0.76 and 0.79, respectively. It is surprising, that \(r_{nd}\) of surface runoff compares very well in both models despite completely different approaches in the parameterizations.

\(r_{nd}\) of drainage (not shown) differs quite considerably between ECHAM and EM and is close to unity for EM (annual basis). Drainage, as parameterized in ECHAM, is mainly dependent on the soil moisture: 'Fast drainage' is initialized when the relative soil moisture exceeds 90% (Eq. 4.15). Thus, drainage is essentially zero during the relatively dry period from May to September, and drainage and total precipitation are little correlated. The
annual cycle of drainage (monthly 'day' and 'night' means) show less scatter than the respective surface runoff due to a large inertia of the relative soil moisture content but the (temporally) highly variable precipitation.

Diurnal cycles

More insight into the characteristics of surface variables can be obtained by analyzing diurnal cycles. In this section, the simulated and observed diurnal cycles of the turbulent heat fluxes, ground heat flux and surface temperature are discussed.

The surface temperature in ECHAM is representative for the middle of the uppermost soil layer, which is 6.5 cm deep.

Fig. 4.48 illustrates the observed and modeled diurnal cycle of the surface temperature. The temperature curves in both the ECHAM and EM are generally within the range of temperature measured at a depth of 2 cm ($T_{s2}$) and the skin temperature $T_{eff}$. All temperature curves increase with increasing radiation input in the morning. The models, however, do not follow the rapid rise of the observed skin temperature $T_{eff}$, due to the “thermal inertia” of the uppermost soil layer and a time-truncation problem in ECHAM (Schulz et al., 1996; Betts et al., 1993). Thus, the simulated surface temperatures are, in the morning, lower than $T_{eff}$. In the cooling phase during the afternoon, the processes are reversed, which leads to a slower decrease of the modeled temperature compared to $T_{eff}$. This characteristic yields a significantly lower amplitude of the generated surface temperature than $T_{eff}$. Since the surface temperature $T_s$ in ECHAM represents the temperature in a depth of 3.25 cm, it is expected that the amplitude in $T_{s2}$ is higher than in $T_s$, which was disproved in Fig. 4.48. The reason for this discrepancy is given in Schulz et al. (1996):
Figure 4.51: As Fig. 4.49 but for latent heat flux.

Since \( T_s < T_{eff} \) generally applies in the morning, the turbulent heat fluxes are underestimated during this period (Figs. 4.50 and 4.51). The residual heat flux must flow into the ground, which leads to a substantially enhanced ground heat flux when compared to the observation in the morning (Fig. 4.49). The monthly mean peak value in April reaches more than 120 Wm\(^{-2}\), compared to approximately 25 Wm\(^{-2}\) in the observation. The relative differences in the ground heat flux are distinctly higher in the transition seasons than in summer. As a consequence of the overestimated ground heat flux, the amplitude of the simulated soil temperature is overestimated. EM simulates similar surface temperatures to ECHAM and, thus, ground heat fluxes do not significantly differ. However, the dramatic overestimation in the amplitude of the ground heat flux is somewhat dampened, using the Extended Force Restore-method (EFR-method, see Section 4.2.1).

To further investigate the influence of the thickness of the uppermost soil layer in ECHAM, an experiment with halved thickness of that soil layer was conducted (Fig. 4.52). A distinct improvement is seen in the simulated ground heat flux when using a thinner uppermost soil layer, which leads to a lower thermal inertia. In April, the amplitude in the ground heat flux is reduced by \( \sim 30 \text{ Wm}^{-2} \). This indicates that the amplitude of the ground heat flux and \( Z_1 \) are likely to be positively correlated. Model experiments have shown that this could lead to numerical instabilities and would, therefore, need a reformulation of the numerical scheme. The comparison of the simulated diurnal cycles of the turbulent heat fluxes (Figs. 4.50 and 4.51) reveals distinct differences between ECHAM and EM. The fluxes, as simulated in ECHAM, generally compare better with the observations. In spring, represented by the month of April, the peak value in the latent heat flux is distinctly higher than in ECHAM. In addition, the rise in the morning, due to solar heating, sets in approximately an hour
earlier in EM. This phase lag, which is also simulated in other months, is mainly caused by a different evolution in the diurnal transpiration rate (not shown), induced by the likely oversimplified approach for transpiration in EM. Further investigations have revealed that the evaporation efficiency $E$ as used in ECHAM, plays a major role in the delay in the onset of transpiration. The time of the onset is better captured in EM. However, it has to be emphasized that this feature likely comes from an error cancellation: The delayed rise in the surface (skin) temperature, as previously outlined, yields a delayed (and too slow) increase of the latent heat flux, while the oversimplified transpiration formulae initializes a premature onset of transpiration. This error cancellation effect is less pronounced in ECHAM.

Fig. 4.53 demonstrates that the diurnal cycles of the surface runoff, as simulated in ECHAM, are in phase with the precipitation rate. As outlined previously, this is primarily related to the Arno scheme in ECHAM, which allows for regions with saturated soils, even if the soil moisture averaged over the entire grid square is below its maximum value. In EM (not shown), this relationship is hardly detectable because EM requires a soil moisture above field capacity for initialization of lateral outflow (remember that surface runoff in EM is defined to contain also the runoff from the uppermost soil layer which contributes, at the Cabauw site, to over 99% of the total surface runoff). Drainage is essentially zero from April through to September due to a soil moisture below the (empirical) threshold value above which 'fast' drainage (Eq. 4.15) is initialized. The smooth diurnal cycle and the low amplitudes in the simulated drainage is closely related to the nearly vanishing diurnal cycle in the soil moisture.

**Figure 4.52**: Comparison of the monthly mean diurnal cycles of the ground heat flux for modified thickness ($z_1$) of the uppermost soil layer.
Phase shifts

In this section, it is investigated how the phases of surface temperature, total surface energy flux and soil heat fluxes are related to each other. The problem is tackled both theoretically and experimentally.

Fig. 4.54 demonstrates that the daily maximum of the surface temperature occurs approximately three hours earlier than the peak value of the surface ground heat flux. The corresponding interval related to the ground heat flux at a depth of 6.5 cm, is 1 - 2 hours. The total surface energy flux is in phase with the surface ground heat flux in both models (displayed for ECHAM in Fig. 4.54). While in EM the cycles of the surface ground heat flux and the total surface energy flux are in perfect agreement, the corresponding amplitudes differ considerably in ECHAM.

These results can be confirmed using the one-dimensional heat conduction equation for a homogeneous soil:

\[
\frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2}, \tag{4.73}
\]

where

- \( k \) Thermal diffusivity \([m^2 s^{-1} K^{-1}]\)
- \( t \) Time \([s]\)
- \( T \) Temperature \([K]\)
- \( z \) Soil depth \([m]\).
This simplified equation, together with
\[
T(0, t) = T_0 + \Delta T_0 \sin(\omega t)
\]
\[
T(\infty, t) = T_0
\]
leads to the following solution for \( z > 0 \):
\[
T(z, t) = T_0 + \Delta T_0 \exp(-z\sqrt{\frac{\omega}{2k}}) \sin \left( \omega t - z\sqrt{\frac{\omega}{2k}} \right),
\]
where
\[
T(z, t) \quad \text{Temperature at depth } z \text{ and time } t
\]
\[
\omega \quad \text{Frequency of the surface temperature.}
\]
The thermal diffusivity is defined by \( k = \frac{\lambda}{\rho c} \), where \( \lambda \) represents the thermal conductivity, \( c \) the heat capacity and \( \rho \) the density of the soil.
The maximum temperature at a certain depth can be determined by setting the first derivative with respect to \( z \) equal to 0. This leads to
\[
t_{\text{max}} = \frac{1}{\omega} \left( \frac{\pi}{2} + z\sqrt{\frac{\omega}{2k}} \right).
\]
This equation describes the phase delay of the temperature signal (and thus ground heat flux) with increasing soil depth. Assuming typical values for the Cabauw site,

**Figure 4.54:** Simulated monthly mean diurnal cycles in EM. 1: Surface temperature; 2: surface ground heat flux; 3: ground heat flux at a depth of 6.5 cm.
\[ \lambda = 1.4 \text{ Wm}^{-1}\text{K}^{-1} \]
\[ \rho c = 2.43 \cdot 10^6 \text{ Jm}^{-3}\text{K}^{-1} \text{ (loamy sand)} \]

yields a phase delay of approximately 2 hours between the surface temperature and the temperature at a soil depth of 6.5 cm, which compares well with the simulated phase shift between the surface ground heat flux and the heat flux at a depth of 6.5 cm.

The ground heat flux \( G(z, t) \) at the depth \( z \) in a homogeneous soil can be expressed as

\[ G(z, t) = -kpc\frac{dT}{dz}. \]  

Inserting Eq. 4.75 in Eq. 4.77 leads to the time \( t_{max} = \frac{\pi}{4\omega} \) at which the ground heat flux is at its maximum. The time difference between the maximal surface temperature and the peak value in the surface ground heat flux amounts to

\[ \Delta t_{max} = t_{max(\text{temp.})} - t_{max(\text{heat flux})} = \frac{\pi}{2\omega} - \frac{\pi}{4\omega} = \frac{\pi}{4\omega} = P/8, \]  

where \( P \) is the period of surface temperature (\( P=24 \text{ hours} \)). This means that the maximum surface soil heat flux occurs 3 hours earlier than the maximum surface temperature. This theoretically derived time lag is confirmed in both models. The respective time difference in the observational data at Cabauw compares well with both simulations and theory; the observed \( \Delta t_{max} \) seems to be slightly smaller.

The above results suggest that both models capture the phase shifts with considerable skill. Further, it suggests that the EFR-method implemented in EM is a beneficial tool to correctly capture the soil temperatures and ground heat fluxes as well as the phase shifts between those variables.
4.6 Comparison of the ECHAM- and EM-parameterizations with other parameterization formulae

It is useful to investigate the impact of different parameterization sets on simulated surface variables. The objective of this particular study is to show how the parameterization type for bare soil evaporation and the stability function might affect the surface climate.

4.6.1 Bare soil evaporation

Evaporation over bare soil involves very complex mechanisms. Hence, numerous parameterizations have been developed in the past. The simulation of evaporation requires a good resolution close to the soil-atmosphere interface in order to ensure the continuity of water fluxes. However, most of the more sophisticated models are unsuitable for use within large scale models because they require numerous prescribed soil parameters for each grid box. The following discussion is, therefore, restricted to simple approaches. The available parameterizations can be divided into
- bulk aerodynamic methods and
- threshold formulations.

Mahfouf and Noilhan (1991) (see references therein) compare several parameterizations for bare soil evaporation with in situ data collected in Montfavet, France.

Bulk aerodynamic methods

The bulk aerodynamic method provides an explicit relation between the bare soil evaporation $E_b$ and the near-surface water content $w_g$ by using a parameterization for the surface specific humidity. The bulk-aerodynamic formulation can be divided into the so-called 'α-methods' and 'β-methods'. The parameterizations used here are briefly summarized.

The α-method requires the computation of the factor $a$ in the formula

$$E_b = \frac{p}{R_a}[\alpha q_{sat} - q_a],$$

(4.79)

where

- $E_b$ Bare soil evaporation [mms$^{-1}$]
- $q_a$ Specific humidity of air [kgkg$^{-1}$]
- $q_{sat}$ Saturated specific humidity at the surface temperature [kgkg$^{-1}$]
- $R_a$ Aerodynamic resistance [sm$^{-1}$]
- $\alpha$ Relative humidity of air at the land surface [mm$^{-1}$].

The factor $\alpha$ is computed as

$$\alpha = \min \left( 1, \frac{1.8w_g}{w_g + 0.3} \right) \tag{Barton, 1979}$$

$$\alpha = \min \left( 1, \frac{0.7w_g + 0.3}{1.8w_g} \right) \tag{Yasuda and Toya, 1981}$$

$$\alpha = \frac{1}{2} \left[ 1 - \cos \left( \frac{\pi w_g}{FC} \right) \right] \tag{Noilhan and Planton, 1989},$$

(4.80)

where

- $FC$ Field capacity
- $w_g$ Near-surface water content.
The $\beta$-formulation is based on

$$E_b = \frac{\rho}{R_a} \beta[q_{sat} - q_a]$$  \hspace{1cm} (4.81)

with

$\ h$ Relative humidity of the air adjacent to the water
$\ \beta$ Moisture availability parameter.

The factor $\beta$ is computed as

$$\beta = \min \left( 1, \frac{w}{0.75PV} \right) \quad \text{(Deardorff, 1978)}$$

$$\beta = \frac{R_a}{R_a + R_{soil}} \quad \text{(Sun, 1982)}$$  \hspace{1cm} (4.82)

and

$$R_{soil} = 3.5 \left( \frac{PV}{w_g} \right)^{2.3} + 33.5.$$  \hspace{1cm} (4.83)

$R_{soil}$ is the resistance to water diffusion in the large soil pores.

**Threshold methods**

The threshold formulations are based on

$$E_b = \min \left[ \rho_w E_t, \frac{\rho}{R_a} (q_{sat} - q_a) \right],$$  \hspace{1cm} (4.84)

with

$\ \rho_w$ density of water [kgm$^{-3}$],

and the water flux

$$E_i = 2D_W \frac{w_g - PWP}{v_1} - K_W \quad \text{(Mahrt and Pan, 1984)}$$

$$E_t = \frac{D_W w_g \pi^2}{4d_1} \quad \text{(Abramopoulos et al., 1988)},$$  \hspace{1cm} (4.85)

with

$\ D_W$ Near-surface hydraulic diffusivity [m$^2$s$^{-1}$]
$\ v_1$ Depth of the top soil layer [m]
$\ K_W$ Near-surface hydraulic conductivity [ms$^{-1}$]
$\ PWP$ Permanent wilting point.

**Model results**

Figure 4.56 shows the evolution of $\alpha$ (thin lines), $\beta$ (thick lines) and $E_t$ for the soil type of loamy sand which is the most frequent soil type in Central Europe. The aerodynamic resistance $R_a$ was set to 50 sm$^{-1}$. The models of Barton (1979) and Yasuda and Toya (1981) compare well, the first-mentioned giving slightly higher values for $\alpha$.

The model of Noilhan and Planton (1989) generates a distinctly lower $\alpha$ when soils are dry, but give enhanced values for soil moisture contents typically observed over Central Europe. According to this model, evaporation is at its potential rate for soil moistures above field capacity. The models, following Barton (1979) and Yasuda and Toya (1981), however, require soil moistures close to the soil porosity to achieve potential evaporation.
The model of Noilhan and Planton (1989) which incorporates a cosine-function, is similar to Eq. 4.41 used in ECHAM to define the relative humidity at the surface. The $\beta$ factor of the two models selected differs for fairly wet soils quite strongly, while small differences are associated with drier soils. The model developed by Sun (1982) does not allow the soil to potentially evaporate.

The threshold formulation models incorporate a linear increase of $E_t$ with increasing soil moisture content, but the model of Mahrt and Pan (1984) assumes lower threshold values (Fig. 4.56b). In the following, it is investigated how the above-discussed parameterization operate within the model framework. Since most models require the near-surface water content ($w_g$) rather than the average soil moisture, it is more reasonable to test the impact of the presented expressions on the surface climate by using the EM (Fig. 4.57). The model is run in an off-line mode using the atmospheric forcing at Cabauw (cf. Section 4.3.1). The evaluations demonstrate that large deviations in (bare soil) evaporation fluxes ($E_b$) are simulated by means of different parameterizations. The control simulation generates monthly evaporation rates which compare well with the maximum values simulated by a bulk aerodynamic formulation. The threshold formulations generally give higher evaporation rates than methods based on bulk aerodynamic expressions, with evaporation being close to potential evaporation from October through to June likely giving an unrealistic result.

Limited confidence is associated with the models of Barton (1979) and Yasuda and Toya (1981) due to negative $E_b$ in winter, which indicates formation of dew (remember that monthly averages are taken into consideration). Both above-mentioned models, which are

![Figure 4.56: a) Factor $\alpha$ and $\beta$ as defined in Eqs. 4.80 - 4.82. Parameter values: $FC = 0.26$, $PV = 0.445$, $R_e = 50 \text{ sm}^{-1}$. b) $E_t$ as defined in Eq. 4.85. Parameter values: $D_w = 5.078 \cdot 10^6 \text{ m}^2\text{s}^{-1}$, $K_w = 8.6 \cdot 10^6 \text{ ms}^{-1}$, $PWP = 0.1$. Definition of $D_w$ and $K_w$ in Eq. 4.28. Soil type for all simulations is loamy sand.](image-url)
based on the application of $\alpha$, generate untrustworthy low evaporation rates, even during summer, and are, therefore, not qualified for use in GCMs.

The ECHAM parameterization produces lower evaporation rates than the EM formulation, the largest difference being found in July. The ECHAM algorithm for $E_b$ is particularly sensitive to warm and sunny weather with high net shortwave radiation which was typically found during July 1981.

Mahfouf and Noilhan (1991) found, that the method of Noilhan and Planton (1989) (Eq. 4.80) fits the observations best, while the threshold methods strongly underestimate surface evaporation. The close agreement of $E_b$, simulated by the ECHAM parameterization and Noilhan and Planton (1989), suggests that the ECHAM parameterization is better adapted for computing the bare soil evaporation.

![Figure 4.57: Simulated monthly means of bare soil evaporation using different parameterizations for $E_b$. All parameterization algorithms are incorporated in EM. 'ECHAM' indicates that the ECHAM parameterization for $E_b$ is used.]

4.6.2 Stability function

The stability function $\Phi_h$ plays a major role in the computation of the turbulent heat fluxes. It is, therefore, relevant to determine this stability function as accurately as possible. Observations in Greenland at the ETH camp during 1991 (A. Ohmura 1997, personal communication) suggest that, in the case of a stably stratified boundary layer, Webb’s formulation of the stability function (Webb, 1970) should be preferred to the expression suggested by Louis (1979) which is used in ECHAM and EM.

The stability functions for stable conditions can be expressed as

$$\Phi_{h,L} = \left(1 + \frac{15R_i}{\sqrt{1+5R_i}}\right)^{1/2}$$  \hspace{1cm} (Louis, 1979) \hspace{1cm} (4.86)

$$\Phi_{h,W} = \left(1 - 5.2R_i\right)^{-1}$$  \hspace{1cm} (Webb, 1970),
where $Ri$ is the Richardson number. The stability functions $\Phi_{h,L}$ and $\Phi_{h,W}$ are displayed in Fig. 4.58. $\Phi_{h,L}$ depicts a slow near-linear increase with increasing stability (i.e. increasing Richardson number) while $\Phi_{h,W}$ increases at a noticeably faster rate.

\[ \Phi_h = \left[ \Phi_m \Phi_h \right]^{-1}, \]

(4.87)

where the index $h$ and $m$ refer to heat/moisture and momentum, respectively. Thus, Webb's correction function $f_h$ approaches zero at a much faster rate than Louis' formulation, implying that the differences in $f_h$ increase with stability. $f_h$ is proportional to the transfer coefficient $C_h$ (Eq. 4.47). Potential evaporation, as well as the sensible heat flux, is parameterized linearly in $C_h$. Hence, the difference in the turbulent heat flux between Webb (1970) and Louis (1979) increases with increasing stability. In Webb (1970), less turbulence is produced than in Louis (1979), assuming same stratification. This leads, as seen in off-line EM simulations, to less evaporation and thus, a reduced drying of the ground. This, again, implies a higher soil moisture content during summer which yields enhanced evaporation during the warmer season (Fig. 4.59). The significant change in the latent and sensible heat flux leads to a substantial change in the Bowen ratio, which is the ratio between the sensible and latent heat flux. The simulated differences in the relative soil moisture content are considerable and amount to approximately 5% in winter and 7% in summer with reversed signs (not shown). It is noteworthy that similar changes in the turbulent heat fluxes are produced by changing the leaf area index from 1 to 5. It is also evident that the simulated surface temperatures are strongly dependent on the choice of the functional form for the stability function $\Phi_h$: In winter, the reduced turbulence, when applying Webb's equation, leads to negative temperature biases of up to 0.5°C, compared to Louis (1979). These results emphasize the significant influence of the stability function on the surface climate.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure4_58.png}
\caption{Stability function $\Phi_h$ of Webb (1970) and Louis (1979) as a function of the Richardson number.}
\end{figure}
4.7 Comparison of the ECHAM3 and ECHAM4 land surface schemes

The land surface schemes of ECHAM3 and ECHAM4 are essentially identical. Some minor differences are discussed in the following section.

Modifications in the land surface scheme

The relative humidity $h$ at the surface (Eq. 4.41) has been reformulated in ECHAM4 to avoid unrealistically large evaporation rates at grid points with large field capacities $W_{\text{smax}}$:

$$ h = \begin{cases} \frac{1 - \cos \left( \pi \frac{W_s - (W_{\text{smax}} - W_{\text{top}})}{W_{\text{top}}} \right)}{2} & \text{if } W_s > W_{\text{smax}} - W_{\text{top}} \\ 0 & \text{if } W_s \leq W_{\text{smax}} - W_{\text{top}} \end{cases} \tag{4.88} $$

This parameterization allows for evaporation from the upper reservoir $W_{\text{top}}$ which is

$$ W_{\text{top}} = \begin{cases} 0.1 \text{ m} & \text{if } W_{\text{smax}} \geq 0.1 \text{ m} \\ W_{\text{smax}} & \text{if } W_{\text{smax}} < 0.1 \text{ m}. \end{cases} \tag{4.89} $$

Figure 4.61 shows the substantial differences of $h$ for $W_{\text{smax}} = 0.2$ m, the value used in the ECHAM3 control simulation, and $W_{\text{smax}} = 0.364$ m which has been estimated by Henderson-Sellers et al. (1993) for the Cabauw site.
A second modification is applied to the parameters which control transpiration rates, i.e. the critical value ($W_{cr}$) and the permanent wilting point ($W_{pwp}$). Both parameters are significantly higher in ECHAM4 than in ECHAM3: $W_{cr}$ is raised from 50% to 75% and $W_{pwp}$ is increased by 15% from 20% to 35%. The Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) suggests, for the Cabauw site, even higher values than those used in ECHAM4: $W_{cr} = 90\%$ and $W_{pwp} = 59\%$. Figure 4.61 illustrates the relationship between relative soil moisture and the water stress factor $F(W_s)$, which controls transpiration via the evaporation efficiency $E$ (Eq. 4.44) for different sets of $W_{cr}$ and $W_{pwp}$.
Figure 4.62: Monthly means of drainage, surface runoff, relative soil moisture. 1: Observation; 2: ECHAM3; 3: ECHAM4, Control; 4: ECHAM4 with PILPS parameters for Cabauw.

Comparison of ECHAM3 and ECHAM4 model simulations

These modifications, discussed above, have been implemented in the ECHAM3 framework and a number of off-line experiments have been conducted to investigate their impact on the surface climate. Some simulated (and, if existent, then observed) surface variables are displayed in Figures 4.62 and 4.63. Results are firstly discussed for the control simulations conducted with ECHAM3 (hereinafter called ECH3) and ECHAM4 (hereinafter called ECH4). In the following section, the surface climate simulated with ECHAM4 but based on the PILPS values for Cabauw, as listed in Table 4.7 (hereinafter called ECHAPILP), are compared with ECH4. The modified parameterization for $h$ (Eq. 4.88) leads to a dramatic decrease in the evaporation from bare soil compared to the parameterization applied in ECH3 (Fig. 4.63). The unrealistically high maximum in the ECH3 bare soil evaporation in April is clearly reduced in the simulation ECH4. The increased values for the permanent wilting point and the threshold value, where the water stress of plants is

Table 4.7: List of parameters used for Cabauw in PILPS.

<table>
<thead>
<tr>
<th>parameter</th>
<th>value</th>
</tr>
</thead>
<tbody>
<tr>
<td>leaf area index</td>
<td>1.3</td>
</tr>
<tr>
<td>vegetation ratio</td>
<td>0.956</td>
</tr>
<tr>
<td>field capacity</td>
<td>0.364 m</td>
</tr>
<tr>
<td>wilting point</td>
<td>0.59 x $W_{smor}$</td>
</tr>
<tr>
<td>critical value</td>
<td>0.90 x $W_{smor}$</td>
</tr>
</tbody>
</table>
initialized ($W_{str}$ in $ECH4$), yield a reduced transpiration rate. The decrease in the water flux components lead to a significantly reduced latent heat flux. A comparison with observed data reveals that the simulation in $ECH3$ compares better with the measurements than in $ECH4$ (Fig. 4.63). The reduced latent heat flux in $ECH4$ implies that from May through November there occurs a substantial decrease in surface runoff. During the period where soils are close to saturation (i.e. with considerable drainage caused by 'fast drainage'), runoff due to drainage is considerably enhanced in $ECH4$. The comparison shows that the reduced latent heat flux in $ECH4$ cannot be compensated by enhanced runoff rates, and leads to a distinct increase in the relative soil water content $W_{srel}$ (Fig. 4.62). During summer and autumn, $W_{srel}$ increases by approximately 15%.

Finally, it is shown how the ECHAM4 model results are modified when using the PILPS parameters for Cabauw ($ECH4PILP$).

The assumption that water stress in plants for $ECH4PILP$ is initialized when the relative soil moisture falls below 90%, leads to a significantly reduced transpiration when compared to $ECH4$. This difference in transpiration is larger than the difference between the experiments $ECH3$ and $ECH4$. A small part of the detected differences is caused by the lower leaf area index ($LAI = 2$ in $ECH4$, and $LAI = 1.3$ in $ECH4PILP$). Although the bare soil evaporation is expected to be higher in $ECH4PILP$ than $ECH4$ due to (i) a thicker bucket layer and (ii) enhanced relative soil moisture due to reduced transpiration, the bare soil evaporation is lower (Fig. 4.63). The likely reason is the different shape of the relative humidity at the surface ($h$) which is essentially zero for $W_{srel} < 75\%$.

**Figure 4.63:** Monthly means of latent and sensible heat flux, transpiration and bare soil evaporation. 1: Observation; 2: ECHAM3; 3: ECHAM4, Control; 4: ECHAM4 with PILPS parameters for Cabauw.
in ECHAPILP (Fig. 4.61). The lower evaporation rates lead to a distinct decrease in the latent heat flux. As seen in Fig. 4.63, the agreement of the simulated and observed turbulent heat fluxes is even worse than in ECH4.

The lower evaporation rates in ECHAPILP lead to a substantial increase in the runoff. Surface runoff is enhanced by the pronounced increase in the relative soil water content and thus, by the fractional saturated area which allows for non-zero surface runoff. Drainage is greater since the higher \( W_{srel} \) induces more 'fast drainage' occurrences. In April, for example, \( W_{srel} \) is higher than 90% in ECHAPILP (monthly mean), whereas in ECH3 and ECH4 it is distinctly lower than 90%, the threshold value required for 'fast drainage'. This leads to considerable drainage in April, while in ECH3 and ECH4 drainage approaches zero. In contrast to ECH3 with rather dry soils during summer, the soil, as simulated in ECHAPILP, is mostly moist the whole year round. This is consistent with the results from Beljaars and Bosveld (1997). It can, therefore, be concluded that a better simulation of the soil water content does not conclusively lead to a better agreement between simulated and observed turbulent heat fluxes.

4.8 Comparison between off-line model simulation with Russian sites

The following results are based on observational data from Russian sites, described in Section 2.2 and Robock et al. (1995). The focus is placed on (i) the snow-rain temperature criterion and (ii) the delayed snowmelt in spring.

Several studies have validated land surface schemes using the that Russian dataset (Robock et al., 1995; Douville et al., 1995b; Yang et al., 1997; Slater et al., 1998).

Forcing data

All the simulations in this section are based on the data observed at Russian stations in 1978 (Section 2.2). Since the meteorological data were recorded at three-hourly intervals, the observations had to be interpolated to 30 minute intervals (corresponding to one time step). For air temperature and the radiation fluxes, a cubic spline interpolation was applied while the precipitation rate, the wind speed and the specific humidity was assumed to be constant during the entire three-hourly interval. The transition from snowfall to rainfall was assumed to occur for \( T = 0^\circ\text{C} \) since the observations do not specify whether precipitation is rain or snow.

ECHAM requires downward longwave radiation (\( LW_\downarrow \)) but the original dataset did not include this variable which had to be calculated separately. Slater et al. (1998) describe several methods to determine \( LW_\downarrow \). Yang et al. (1997) and Slater et al. (1998) both report that the following approach provides the best estimate:

\[
LW_\downarrow = \epsilon_a C \sigma T_a^4, \tag{4.90}
\]

where \( \epsilon_a \) is the clear-sky emissivity of the atmosphere and \( \sigma \) is the Stefan-Boltzmann constant. \( C \) represents the increase in clear sky emissivity caused by clouds:

\[
C = 1 + 0.2 \cdot (C_L + C_M)^2 + 0.04 \cdot C_H^2, \tag{4.91}
\]

where \( C_L \) is the lower cloud fraction as provided in the observed data set. \( C_M \) and \( C_H \) are the middle and high level cloud cover fractions and are determined as in Slater et al. (1998):

\[
C_M = C_H = (C_T - C_L)/2, \tag{4.92}
\]
where $C_T$ is the total cloud cover fraction.

$\epsilon_a$ is assumed to be

$$
\epsilon_a = 1.08 \cdot \left[ 1 - \exp(-e_a T_a^{1/2016}) \right],
$$

which provides good estimates in cold regions. $\epsilon_a$ and $T_a$ represent the vapour pressure (hPa) and air temperature (K) at standard level of measurement, respectively.

**Experimental design**

The experimental design is strictly the same as in Roesch et al. (1997) (cf. Section 4.3.1). Only the atmospheric forcing data and the parameters which characterize the land surface differ. All simulations using the Russian data are based on the first year (1978) of observed data. The model was run for five years applying the one-year forcing data repeatedly to reach equilibrium. The evaluation is based on the last year of this 5-year simulation.

The list of parameters which has to be prescribed in ECHAM4 is given in Table 4.8 and is taken from Yang et al. (1997).

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Proportion of area covered by vegetation</td>
<td></td>
<td>0.8</td>
</tr>
<tr>
<td>Fractional forest area</td>
<td></td>
<td>0.0</td>
</tr>
<tr>
<td>Surface roughness</td>
<td>m</td>
<td>0.2</td>
</tr>
<tr>
<td>Maximum soil moisture content</td>
<td>m</td>
<td>0.45</td>
</tr>
<tr>
<td>Permanent wilting point</td>
<td>m</td>
<td>0.135</td>
</tr>
<tr>
<td>Critical value for water stress</td>
<td>m</td>
<td>0.25</td>
</tr>
<tr>
<td>Leaf area index (winter)</td>
<td>m</td>
<td>0.5</td>
</tr>
<tr>
<td>Background surface albedo</td>
<td></td>
<td>0.2</td>
</tr>
<tr>
<td>Minimum stomatal resistance</td>
<td>s m$^{-1}$</td>
<td>200</td>
</tr>
</tbody>
</table>

**Validation Data**

Numerous observational data describing snow characteristics are available for the sites used in this study. Snow based measurements, taken every ten days, include snow water equivalent, snow density, fractional snow cover and snow depth near the station.

**Validation of snow variables**

In the 3-D ECHAM simulations, the delayed snowmelt implies excessive snow amount in spring (cf. Section 3.4). This section should clarify if this model deficiency vanishes when using off-line simulations with an atmospheric forcing.

Fig. 4.64 shows that the snow water equivalent is generally overestimated in the model which is in line with Roesch et al. (1999). The inception of the snow season is captured rather well by ECHAM. In addition, the snow accumulation in January and February is simulated with reasonable skill at Uralsk and Kostroma while at Yershow, snow water equivalent decreases slightly during the last three weeks in January. Topographic complications or high wind speeds during snowfall may be responsible for these differences (Slater et al., 1998).

The results from off-line model simulations for the sites Yershow and Kostroma illustrate that the spring snowmelt is delayed by approximately 10 - 15 days. Unfortunately, the resolution of the observed data for determining the end of the snow season is not ideal,
Figure 4.64: Off-line model simulations for three Russian sites. Year 1978. The observations (shown by crosses connected with solid lines) of snow water equivalent, snow cover fraction and surface albedo were made every ten days during the winter season. The model results (dashed lines) have the same time resolution.

but where observations exist, ECHAM tends to overpredict snow water equivalent towards the end of the snow season. These results indicate that the delayed snowmelt in spring is probably not only due to incorrect temperature distributions and precipitation rates in the 3-D ECHAM experiments but also point out model deficiencies in the parameterization of snow melt. This may be caused by the following approach applied in the ECHAM parameterization: snow melt is only initialized when both the snow temperature and the temperature of the uppermost soil layer reach 0°C (cf. Section 4.2.3). One can speculate that a multi-layer snow model would, allowing the snow to melt when only the uppermost snow layer temperature reaches 0°C, lead to a faster snow melt in spring. In addition, snow melt would be initialized earlier when reducing the thickness of the uppermost soil layer (currently set to 6.5 cm). However, the validation of the control simulation with globally observed fields of temperature and precipitation revealed that in April, in large parts of the former Soviet Union and North America north of 50°N, excessive precipitation is simulated (cf. Section 3.4). This yields excessive snow fall and thus, abundant snow. Furthermore, the link between snow and albedo provides a positive feedback enhancing the initial difference. The cold temperature bias in April, which has been found in the control simulation in the above-mentioned regions, confirms this correlation. An additional source
of error could be erroneous representation of the downwelling longwave radiation. Slater et al. (1998) found, using off-line simulations with BASE (Best Approximation of Surface Exchange), that the snowpack ablation is very sensitive to downward longwave radiation. The observed snow cover fraction equals 1 during almost the whole winter, excluding the inception of the snow season as well as the period of snow melt (Fig. 4.64). These measurements are probably afflicted with errors: Despite a shallow snowpack in Uralsk and Kostroma during November, $f_s$ is equal to 1 (with one exception). Slater et al. (1998) also report that the observed fractional snow cover correlate poorly with both the observed snow depth and the observed albedo. Therefore, it is inappropriate to compare these observations with the simulated snow cover fraction. The model generally underestimates the observed surface albedo. Since the surface albedo is largely influenced by the snow cover fraction, it is likely that much of the deviation in surface albedo is attributable to differences in the fractional snow coverage rather than to errors in the actual snow albedo.

**Sensitivity to the snow-rain temperature criterion**

The description of the snow-rain temperature criterion is detailed in Section 5.8. It has been shown that the fully developed three-dimensional model version of ECHAM4 is sensitive to the value of the temperature when snow flakes' melt in an atmospheric model layer is initialized. This section should clarify the importance of the snow-rain temperature $T_{rs}$ using off-line simulation with atmospheric forcing. Fig. 4.65 displays the response of the snow conditions to a rise of $T_{rs}$ from 0.0°C up to 2.0°C. The snow water equivalent is significantly reduced, with increasing deficit from mid-winter to the end of the snow season. The relative decrease in mid-winter is approximately 20%. Towards the end of the snow season, this percentage increases to ~80% at the sites of Yershow and Kostroma. It is evident that the marked decrease in snow water equivalent is related to a substantial decrease in the snow cover fraction and thus, surface albedo. This leads to an enhanced net shortwave radiation, thereby producing significant warm temperature biases which have similar magnitudes during the inception of the snow season and the ablation period. The corresponding peak values in the $f_s$-differences, however, differ strongly at Uralsk and Yershow. This apparent discrepancy is due to markedly larger global radiation in March (ablation period) than in late November (inception of snow period).

Increasing the snow-rain temperature $T_{rs}$ leads to better agreement of the simulated snow water equivalent with the observation at Yershow and Uralsk, but reduces the agreement at Kostroma. It can thus be assumed, that a readjustment of $T_{rs}$ is not sufficient to reach perfect agreement between the observed and simulated snow water equivalent at all the Russian sites. Nevertheless, it should be mentioned that the most accurate estimate for the Russian sites is $T_{rs} = 0.75°C$, following Yang et al. (1997) who used observational data for the period 1936 - 1985.

**4.9 Summary and conclusions**

The focus of this work is set on a comparison of the land surface schemes used in ECHAM and EM. After a description of both land surface schemes, a detailed study of the sensitivities to a number of key surface parameters is given. The analysis of variations in the surface climate, primarily comprising of the surface energy and water budget, are based on off-line model simulations with atmospheric forcing described on the lowest atmospheric model level. A further goal of the sensitivity study is, to compare simulated sensitivities with sensitivities derived from isolated parameterization equations.
In the second part, annual and diurnal cycles of various surface climate variables simulated by ECHAM and EM have been compared to each other and to observations. Towards the end of this report, the impact on the surface climate by substituting a specific parameterization by a new algorithm was investigated. Further, a discussion of the differences between ECHAM3 and ECHAM4 is included.

**General remarks**

The results of this study stress the importance of the correct specification of key surface parameters in a GCM. However, it has been shown, that the differences in the surface climate, using either the land surface schemes of EM or ECHAM, are typically larger than the difference generated by a significant variation of a surface parameter. The same applies to the introduction of a new parameterization algorithm for a specific process, e.g. bare soil evaporation. The study shows that the comparison should not be restricted to annual and monthly means but should be extended to diurnal cycles to capture the significant differences between daytime and nighttime. Moreover, it is demonstrated that it is undue to assume constant sensitivities over the entire parameter range due to highly nonlinear model structures. The findings illustrate that it is often efficient to compute the impact of a surface parameter on a surface (output) variable by using the isolated parameterization
equation. Individual components of an output variable generally show larger sensitivities and larger variations between the two investigated land surface schemes than the variable consisting of all components (e.g. latent heat flux versus bare soil evaporation)

**Differences between the sensitivities to key surface parameters in ECHAM and EM**

The key land surface parameters tested are: surface albedo (background albedo and snow albedo), vegetation ratio, leaf area index, roughness length and the maximum soil water content. The main differences between ECHAM and EM are briefly summarized.

1. **Maximum soil water content**: The assessment of the impact of maximum soil water content on the surface climate is afflicted with problems due to the different structure of the soil model with regard to water transport (bucket model versus multi-layer model with vertical water transport). However, both land surface schemes are highly sensitive to the value of $W_{\text{max}}$, caused mainly by the principal role of $W_{\text{max}}$ in processes such as transpiration, bare soil evaporation, surface runoff and drainage. These variables are typically parameterized using the relative soil moisture content rather than its absolute value. The model simulations reveal that changing $W_{\text{max}}$ also affects its relative value. The EM land surface scheme usually generates a higher sensitivity to $W_{\text{max}}$. The sensitivity to surface climate parameters generally levels off when $W_{\text{max}}$ attains a certain value due to a “saturation effect”: If $W_{\text{max}}$ is sufficiently high, the soil water becomes no longer a limiting factor for evapotranspiration. The impact of surface climate variables on $W_{\text{max}}$ is in summer generally significantly larger than in winter, since, during winter, sufficient water is available most of the time, independent of the maximum water storage capacity. Since the response of latent heat flux on $W_{\text{max}}$ is considerable, the Bowen ratio also changes considerably. Since the EM soil allows for different soil depths in which different processes (e.g., transpiration or bare soil evaporation) originate, the EM parameterization is favoured above the ECHAM.

2. **Vegetation ratio ($\sigma_{\text{PLNT}}$)**: The impact of vegetation is, in both models, poorly parameterized. Only the transpiration rate and processes related to the skin reservoir are dependent on $\sigma_{\text{PLNT}}$. EM further incorporates the surface albedo and the maximum infiltration rate so that it becomes explicitly dependent on the vegetation ratio. The response of variation in the vegetation ratio on surface climate variables is in EM typically more than doubled than in ECHAM (on an annual basis) due to the more frequent explicit incorporation of $\sigma_{\text{PLNT}}$ in parameterization equations. The sensitivity generally increases with increasing vegetation ratio. However, there are some exceptions where highest sensitivity for $\sigma_{\text{PLNT}} = 0.5$ is obtained, likely related to the parameterization of the maximum infiltration rate in EM.

3. **Leaf area index (LAI)**: Since the operationally used EM model does not allow for any dependence on the LAI, all the EM results are based on the Dickinson parameterization. Transpiration, evaporation from the skin reservoir and the latent heat flux generally increase with increasing leaf area index, whereas the reverse applies to runoff, soil moisture content, bare soil evaporation and surface temperature. Most sensitivities of surface climate variables decrease with denser foliage, due to the following saturation effect: assuming an exponential decrease of incoming shortwave radiation with canopy depth (which is in both land surface schemes represented by the LAI) implies, for high LAI values, a low response in the transpiration rate on the LAI.

4. **Surface albedo**: A higher albedo leads to a smaller net shortwave radiation and yields a temperature decrease, thereby increasing the stability and reducing the turbulent heat fluxes, which subsequently enhance runoff and soil moisture. The sensitivity to albedo is
generally larger in ECHAM than in EM primarily due to the parameterization of transpiration, which is strongly affected by the net shortwave radiation in ECHAM. The sensitivity of total surface albedo to the snow cover is significantly larger in EM than in ECHAM, primarily due to different relationships between the snow water equivalent and snow cover fraction.

(5) Roughness length: Turbulent heat fluxes become larger with increasing $z_0$ since turbulence is enhanced over rough surfaces. The wind speed is part of the atmospheric forcing which was used to drive the schemes and is, therefore, identical in all the conducted model simulations. The sensitivity of most surface variables decreases with increasing roughness length, which is related to the lower derivative of the heat transfer coefficient $C_h$ with respect to $z_0$ over rough surfaces compared to smooth surfaces.

Two general remarks concerning sensitivities should be added:

- The impact of variation in surface parameters on ground heat fluxes is significantly higher in EM than in ECHAM due to the prescribed deep soil temperature.
- The sensitivity of the skin reservoir and its evaporation to most key surface parameters is approximately an order of magnitude higher in ECHAM than in EM, being a result of the rapid infiltration of the skin reservoir water into the soil.

Differences between the modeled surface climate in the ECHAM and EM control simulations and comparison with observations

- The simulated surface albedo is, during winter, significantly larger in EM than in ECHAM. This is the result of different approaches in ascertaining the relationship between snow water equivalent and snow cover fraction. More information concerning the parameterization best qualified for use in GCMs, is given in Sections 5.1 and 5.2.
- The annual amplitude in the surface temperature is larger in EM than in ECHAM, caused by warmer summer temperatures but colder temperatures in winter. This result is associated with the lower simulated net shortwave radiation during winter in EM, as well as the prescribed deep soil temperature. The comparison with observation reveals that the simulated surface temperature generally lies between the observed skin temperature and the temperature at a soil depth of 2 cm.
- The ground heat flux simulated in ECHAM compares rather well with the observations (on a monthly basis), whereas the downward component in EM is significantly overestimated. This deficiency is related to the prescribed deep soil temperature ($T_D$) in EM. Forcing both models with identical surface temperature and identical $T_D$ demonstrates that both the explicit EFR method (applied in EM) and the implicit solution of the heat conduction equation (applied in ECHAM) generate almost identical ground heat fluxes. The diurnal amplitude of the ground heat flux is strongly overestimated in both models due to the thermal inertia of the uppermost soil layer, which leads to a wrong diurnal cycle of the skin surface temperature.
- Water vapour fluxes: During winter and spring, the transpiration rate is higher in EM than in ECHAM. The lacking dependence on the photosynthetically active radiation (PAR) on the stomata closure is likely to be the main reason. The bare soil evaporation, simulated in ECHAM, reaches its peak value already in April and, subsequently, decreases at a rapid rate to reach a July mean which is typically simulated during winter due to the dry soil. The EM simulation of bare soil evaporation, in contrast, shows the expected annual cycle reaching its maximum in summertime. The skin reservoir content and evaporation are generally more than twice as high in ECHAM than in EM owing to the fast infiltration of skin reservoir water into the soil.
Total annual runoff is larger in ECHAM than in EM to fulfill the water balance since the soil water is evaporated more efficiently in EM. Substantial differences are mainly simulated during the winter due to the process of 'fast drainage' in ECHAM and the parameterization of surface runoff which allows the water to run off when soils are not fully saturated. Significant differences are also found in the partition of total runoff into surface runoff and drainage, caused primarily by the capability of EM to simulate water fluxes between the soil layers.

Model development

Although it might be risky to suggest model improvements purely based on the experience collected in the course of this work, an attempt to establish a basis for a "better" land surface scheme is presented in the following section.

An improved land surface scheme should ideally include the following components:

a) from ECHAM

- Several soil layers for temperature (to avoid artificial boundary conditions) and the determination of soil heat fluxes
- The parameterization of the transfer coefficients $C_h$ for heat and moisture (essentially the same in both land surface schemes)
- Transpiration (the respective parameterization in EM is oversimplified). ECHAM lacks the transpiration to originate from different soil depths.
- Correct representation of the albedo of snow-covered vegetation (forests) and a polynomial relationship between snow albedo and temperature (the implementation of an aging function would likely produce more reasonable results).

b) from EM

- Several soil layers for water transport (for bare soil evaporation, it may be preferable to introduce a very thin uppermost soil layer)
- Surface runoff/drainage (runoff from different soil layers)
- Infiltration rate
- Response of soil moisture and vegetation ratio on surface albedo
- Snow melt (melt from lower and upper boundary of snow pack, redistribution of heat).

The study has shown that both land surface schemes primarily need improvements regarding the following processes:

- Introduction of a simple physically based canopy model for a correct description of radiation and temperature within the vegetation, which would lead to a more realistic snow distribution and surface albedo
- Incorporation of a very thin uppermost soil layer (for temperature and water) in order to improve the simulated surface temperature and the evaporation rate from bare soil. However, it must be guaranteed that the implemented numerical method does not lead to instabilities.
• Physically based parameterization of albedo for snow covered conditions.

• A reformulation of the stability function for vertical diffusion in stable conditions.
5. 3-dimensional ECHAM4 simulations

The surface albedo is one of the most important land surface parameters in GCMs. However, in ECHAM, most of the processes determining the albedo are either very poorly parameterized or neglected.

For this reason, various ECHAM4/T42 experiments have been set up to investigate the influence of more sophisticated schemes for the surface albedo. In each climate simulation, a climatological annual cycle of sea surface temperature (SST) and sea ice coverage are prescribed as boundary forcing while interannual SST variability has been neglected. Daily values were computed using a linear interpolation between the monthly means of SST and sea ice coverage from the atmospheric model intercomparison project (AMIP) (Gates, 1992).

All experiments have been initialized with the same set of atmospheric data (ECMWF analysis for October 1, 1983). Each simulation covers a period of eleven years and three months. Since these simulations need a spin-up time of about a year in order to settle to their own climate, the first 15 months were discarded. Therefore, all climatological means refer to a ten-year period.

The simulations were carried out on a NEC SX4 installed at the Swiss Scientific Supercomputing Center (CSCS). This powerful computer with four processors allowed several long-term integrations at T42 resolution within a few months.

It is not worthwhile to integrate the ECHAM4 in the higher resolution T106 since the main features and results appear to be similar in both resolutions. The higher resolution of T106 (1.1° x 1.1°) needs more CPU time than T42 (2.8° x 2.8°) by a factor of 11. The reason for this dramatic increase of computational requirements is not only the higher horizontal resolution in T106 but also the shorter timestep (t = 24 min and t = 12 min for T42 and T106, respectively) which is necessary to ensure numerical stability. In addition, most of the former model studies have been performed with T42 resolution and have shown the T42 resolution to be sufficient for climate studies (K. Arpe, pers. comm.). However, studies of local phenomena, e.g. orographic precipitation at the slopes of the Himalayas or the Andes, and comparisons with site measurements are more difficult when using the coarser resolution. Furthermore, processes which are largely influenced by the orography, e.g. atmospheric blocking, should be investigated with a high resolution model (Tracton, 1990).

**Overview of the experiments**

The work includes ten 10-year simulations which have been conducted to assess the influence and problems inherent to surface albedo and snow cover. All the simulations are conducted with the 3-D ECHAM4/T42 GCM. A review of the experiments is given in the following.

EXP1: The modification of the snow cover fraction parameterization with a $\tanh$-expression. This expression has been developed using satellite-derived measurements of the snow cover extent and the global snow depth climatology compiled by the U.S. Air Force Environmental Technical Application Center (USAF/ETAC).

EXP2: The implementation of the snow cover fraction expression as in the latest version of the Météo-France climate model. This expression includes an orography effect in the computation of the snow cover fraction using the standard deviation of the subgrid orography.
EXP3: The implementation of a polynomial temperature dependence of snow albedo between -10°C and 0°C, based on measurements at six Russian sites over six years. The range of the surface albedo over snow covered forests had been reduced by 0.1.

EXP4: The division of total surface albedo into the near-infrared (NIR) and visible (VIS) components. This simulation was further modified by the introduction of a zenith angle dependence of the water albedo.

EXP5: The implementation of the parameterization set for snow albedo as used in the Biosphere Atmosphere Transfer Model (BATS) which was originally designed for use in the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM).

EXP6: The annual cycle of the leaf area index (LAI) has been extracted from the International Satellite Land Surface Climatology Project (ISLSCP) dataset.

EXP7: The implementation of a simple canopy model, as used in CLASS (Canadian Land Surface Schemes for GCMs), to clarify the effect of snow covered vegetation on surface albedo. Moreover, a new parameterization to account for the influence of wind and temperature on intercepted snow was introduced.

EXP8: The ECHAM GCM assumes the melting of snow-flakes when the mean temperature of the corresponding atmospheric layer exceeds 2°C. This experiment is based on a threshold temperature of 1°C.

EXP9: The introduction of a surface albedo dependent on soil moisture.

EXP10: The effect of subgrid topography is explicitly incorporated for an improved representation of the area with snowfall and rain, respectively. For the snow pack, two regimes (above and below 0°C) in each grid cell are maintained.

Structure of the sections on the 3-D model simulations

In the Sections 5.1 - 5.10, a detailed description of the model experiments is given. The structure of each section is generally as follows.

1) For clarification, after an introduction, the model modifications are shortly summarized. The focal point is placed on the distinction between the part which has been developed in the course of this work ("the scientific contribution") and the part adapted from literature. This part anticipates thus the results which result from point 2) and 3).

2) A short overview of other work done in the corresponding field.

3) Most discussions focuses on the derivation of the implemented algorithms which is intended to be one of the major goals of this work. This often includes off-line studies and/or comparisons with observational data.

4) Results from the 10-year integration of the 3-D ECHAM4/T42 GCM. In most model experiments, the effect of the modification will be discussed on an annual and monthly basis for the Northern Hemisphere (NH) or the NH land area which is usually snow covered in winter. In addition, one or two regions of special interest are selected and the response of the (surface) climate is detailed.

5) Finally, a short summary of the main results is given.

5.1 EXP1: Snow cover fraction deduced from global datasets

Snow cover fraction is crucial for the computation of surface albedo during the winter season and literature presents many parameterizations in this regard. However, most
GCMs treat the snow cover fraction in an oversimplified way. It is reasonable that most models incorporate the snow water equivalent or the snow depth in the parameterization of the snow cover fraction. Several approaches include a parameter accounting for varying vegetation roughness. Only a few expressions take into account the reduction of surface albedo due to subgrid-scale orography, e.g., over rough mountainous regions. The following section gives an overview of some expressions for snow cover fraction used in snow models of GCMs.

5.1.1 Overview of modifications

In this experiment, Eq. 4.32 will be replaced by a new relationship between the snow cover fraction $f_s$ and snow water equivalent $S_n$. This approach suggests that the functional between form between $f_s$ and $S_n$ can be approximated by a tanh-function (cf. Eq. 5.12). The function will be derived from a ground-based and satellite derived dataset and should only be applied to more or less flat and non-forested grid elements.

5.1.2 Parameterization of snow cover fraction

The simplest formula for snow cover fraction assumes a constant value for the snow cover fraction. The CCM1, the first version of CCM, developed at the National Center for Atmospheric Research (NCAR) initially used a snow cover fraction of 0.5 wherever snow occurred (Williamson et al., 1987).

The snow covered grid fraction $f_s$ in the ECHAM GCM is calculated according to Equation 5.1.

$$ f_s = \frac{S_n}{S_n + S_n^*}, $$

where $S_n$ is the water equivalent of snow in meters and $S_n^*$ the critical snow depth ($= 0.01$ m). Note that no distinction between different vegetation and soil types is provided. The above approach is used for all resolution (T21, T42 and T106) despite a significant effect of the grid resolution on the formulation of snow cover fraction.

Some other snow models use similar expressions to those used in ECHAM. The Météo-France climate model with the ISBA0 (Interaction between Soil, Biosphere and Atmosphere) land surface scheme uses the same empirical expression.

Several parameterizations incorporate, as second variable, the roughness length of the vegetation as an estimate of the masking effect: The higher the roughness, the higher the masking effect.

In the CCM2, the 2nd version of the NCAR climate model, the snow cover fraction is computed following Eqs. 5.2 and 5.3 (Marshall et al., 1994):

$$ f_s = \frac{f_{so}}{1.0 + f_{so}} $$

$$ f_{so} = \frac{0.1S_n}{0.2\varepsilon_0}, $$

where $S_n$ is the snow water equivalent (in kgm$^{-2}$). The term $S_n/0.2$ represents the normalized snow water equivalent, assuming a constant snow density of 200 kgm$^{-3}$. The roughness length is normalized with 0.1 since the roughness length is approximately 10% of the geometric height of the vegetation.

In the Biosphere Atmosphere Transfer Model (BATS), the fraction of vegetation covered by snow (Dickinson et al., 1993), includes the snow depth $h_s$ and the roughness length for
The fraction of bare soil covered by snow is inferred according to the formula

\[ f_s = \frac{h_s}{h_s + 10z_{0b}}, \]  

where \( z_{0b} = 0.01 \) m is roughness length for bare soil.

Another functional form for snow cover fraction is proposed by Yang et al. (1997):

\[ f_s = \tanh \left[ \frac{h_s}{2.5z_0} \right], \]  

with the abbreviations as above. This expression is based on observations by Baker et al. (1991) who discussed the snow depth required to mask the underlying surface, based on the measurements of the daily mean surface albedo and snow depth. Their results showed that there are two distinct stages in the surface albedo/snow depth relationship. During the first stage, the surface albedo increases rapidly as snow depth increases before it reaches a critical depth. During the second stage, on the contrary, the surface albedo slowly increases as snow depth increases.

The form of Eq. 5.6 may be the most appropriate one for grass and agricultural land, which commonly experience vegetation slumping due to snow burdening (Yang et al., 1997). For spatially uncorrelated stands such as forests, however, a more exponential form can be expected. This is confirmed by Barker and Davies (1989) and Otterman (1984). The latter treated the forest as thin, vertical cylinders which are randomly distributed.

The Europa-modell (EM), developed at the DWD (German Weather Service), suggests an exponential form for all vegetation types:

\[ f_s = \max (0.01, 1.0 - e^{-S_n/0.0025m}). \]  

It should be stressed here that the approach as used in EM is valid for resolution between approximately 15 km and 50 km. The resolution is thus distinctly lower than in ECHAM4/T42 (~280 km).

SiB2 (Simple Biosphere Model), described in detail in Sellers et al. (1996a), incorporates for bare soil a linear variation of \( f_s \) in respect of the snow water equivalent:

\[ f_s = a_s S_n, \]  

where \( a = 0.132 \) m\(^{-1}\). The snow covered canopy fraction \( f_{sc} \) is determined according to

\[ f_{sc} = 0.5 \frac{S_{ni}}{S_{ni,max}}, \]  

where \( S_{ni} \) is the intercepted snow on the canopy in [m]. \( S_{ni,max} \) stands for the canopy storage limit which increases by \( 10^{-4} \) m as the LAI increases by one unit.

A rather recent formulation for \( f_s \), which includes as further variable the standard deviation of the subgrid orography (\( \sigma_z \)) in metres, has been proposed by Douville et al. (1995b):

\[ f_s = \frac{S_n}{S_n + S_0} \sqrt{\frac{S_n}{S_n + \max(1.0m, 0.15\sigma_z)}}, \]  

where \( S_0 = 0.01 \) m. The first part of the equation is the same as in ECHAM (cf. Eq. 5.1). The second part accounts for a reduction of the surface albedo due to the irregular distribution of snow cover in mountainous areas.
Some of the proposed relationships between snow depth and the fraction of surface covered by snow are displayed in Fig. 5.1. Note the large differences in the snow cover fraction for a fixed snow depth. The curve using the \( \tanh \)-function typically evolves between those of ECHAM and EM. CCM2 computes unduly low snow cover fraction, e.g. for a snow depth of 15 cm, the grid element is supposed to be only half snow covered. To summarize, all

\[
\text{ECHAM4} \quad \text{EM} \quad \tanh(h_s/(2.5 \ z_0)) \quad \text{CCM2} \quad \text{SIB2}
\]

**Figure 5.1:** Relationship between snow depth and snow cover fraction assuming \( z_0 = 2 \text{ cm (grass)} \) and a snow density \( \rho_s = 300 \text{ kg m}^{-3} \) using various parameterizations.

the above described formulae are simple expressions in respect of variables such as snow water equivalent, surface roughness, leaf area index or a measure of subgrid orography. However, an exact form, suitable for a grid square should be established with remotely sensed observations of snow cover data (see Section 5.1.3).

5.1.3 Derivation of a snow cover expression using satellite data

An expression for snow cover fraction suitable for large areas (e.g., grid elements in GCMs), can best be derived from satellite based measurements of snow cover fraction and surface snow depth observation. For the global snow depth climatology, the dataset compiled by the U.S. Air Force Environmental Technical Application Center (USAF/ETAC) was used, while for the weekly snow cover, the global dataset compiled by the National Oceanic and Atmospheric Administration (NOAA) was utilized (see Section 2.1).

In order to compile a global snow cover climatology, the weekly snow cover information from 1973 - 1996 was analyzed. This data contains information regarding the existence (1) or non-existence (0) of snow cover. The "frequency", or empirical probability, of snow cover is the ratio of occurrence of 1's to the total count of samples in each gridbox. To give an idea of the annual variability, the annual cycles of snow covered area of North America and Eurasia are displayed in Fig. 5.2. The annual cycle is about one decimal order-of-magnitude larger than the interannual variability. The growth of the snow cover is about 1.5 times faster than that of the decay in spring. The extremes of the snow cover extent (dashed lines) indicate that snow cover patterns differ considerably from year to year. The parameterization of monthly snow cover fractions using monthly snow water
Figure 5.2: Annual cycles of total snow covered area over North America, Eurasia and the Northern Hemisphere. Thin solid lines: standard deviation; dashed lines: monthly maxima and minima during 1973 - 1996. All curves are based on the satellite-derived NOAA snow cover data.

Figure 5.3: Comparison of simulated mean monthly snow cover fraction for January and March. Data from the ECHAM4 control run. Abscissa: based on high temporal resolution data (24 min); ordinate: based on monthly means of snow water equivalent. Parameterization of snow water equivalent according to Eq. 5.1. Solid line: Eq. 5.11.

equivalents should be carefully investigated. If the snow depth varies significantly during the month (snow melt season), this may yield substantial errors. One would expect very different monthly mean snow cover fractions for $S_n = 1$ cm on all 30 days, as opposed to
10 cm of snow on 3 days with the remaining 27 days snow free. To account for this problem, the monthly snow cover fraction are computed benefiting from the high time resolution in the T42 simulations (24 min). The difference in the monthly snow cover fraction computed with temporally low and high resolved snow water equivalent data, could, therefore, be investigated in greater detail. Fig. 5.3 shows the difference between the two averaging methods for January and March. Note that the largest differences occur for half snow covered grid boxes. To account for this averaging problem, a correction factor was applied to snow cover fractions which are derived from monthly means of its associated snow water equivalent. This correction factor was derived from a simple fitting function to the data (Eq. 5.11).

\[ f_{sy} = \sqrt{5 - (f_{sx} - 4)} - 1, \]

where \( f_{sx} \) and \( f_{sy} \) denote the snow cover fractions as computed with snow water equivalent data of high and low temporal resolution, respectively.

A comparison between the remotely sensed observation and the snow cover fraction computed by ECHAM4, reveals a severe deficiency in the ECHAM4 parameterization of the snow cover extent (Eq. 5.1). The USAF snow depth climatology, combined with the

![Figure 5.4: Snow cover intercomparison for Eurasia of results from ECHAM/T42 (T42), NOAA (1973 - 1996) (NOAA), the USAF snow depth climatology (USAF) and the reanalysis (Reanalysis). Note that for the USAF and the re-analysis data, snow water equivalent was transformed to snow cover fraction using Eq. 5.1. USAF1: USAF data but transformed with the tanh-function (cf. Eq. 5.12). Additionally, a correction term to account for the low temporal resolution was applied.

ECHAM transformation of snow water to snow cover fraction, substantially underestimates the satellite-derived snow covered area (Fig. 5.4). Considerable differences are detected during winter and spring. ECHAM4/T42 (control run) is close to the NOAA estimate for March and April. However, this contradicts the expected feature since the snow water equivalent is strongly overestimated in the ECHAM4/T42 simulation during the spring snow melt period (cf. Section 3.4). These results give a clear hint as to the failure of the model's Equation 5.1. The evaluation for North America leads basically to the same findings.
Different approaches were tried in order to yield better estimated snow covered area. The best method to show which expression fits the observation most effectively, is to plot the (observed) snow water equivalent versus the (remotely sensed) snow cover fraction (Fig. 5.5). The USAF snow depth climatology and the NOAA snow cover data are generally thought to be the best datasets. Since satellite-based measurements of snow cover over forested areas have limited confidence, all grid elements with more than 10% forested area were excluded. This is the reason why the new snow cover expression can only be applied to the non-forested component of the grid box. In addition, the expression will be inappropriate for orographically very rough areas. Note that the high resolution of T106

![Graphs showing snow cover fraction as a function of snow water equivalent](image)

**Figure 5.5:** Snow cover fraction as a function of snow water equivalent. Snow water equivalent from USAF snow depth climatology; snow cover fraction: NOAA data (1973 - 1996). Both data sets were interpolated onto the T106 grid. Only grid elements with less than 10% forest fraction, $a_z < 300$ m as well as $S_n \geq 0.1$ mm were considered. Long-dashed: EM (Eq. 5.7); dash-dotted: ECHAM (Eq. 5.1).

was used to achieve a better statistical performance. The $tanh$-function, proposed by Yang et al. (1997), captures the functional form best. Therefore, the following approach is suggested

$$f_s = b \cdot \text{tanh}(a S_n) \quad \text{with} \quad a = 1 \, \text{cm}^{-1}; \quad b = 0.95,$$

(5.12)

The CURVEFIT function in IDL was used to obtain the best fit. This function uses a gradient-expansion algorithm to compute a non-linear least-squares fit to a given function.
with an arbitrary number of parameters (Marquardt, 1963). The coefficients $a$ and $b$ in Eq. 5.12 have been calculated separately for each month from November to April, yielding $a = 1 \text{ cm}^{-1}$ and $b = 0.95$. From Fig. 5.5, it is evident that the original parameterization in ECHAM substantially underestimates the snow cover fraction, while the expression used in the Europa-modell (Eq. 5.7) represents nearly the upper envelope of the point cloud displayed in Fig. 5.5. The exponential form appears to be inappropriate for describing the relationship between snow cover and snow water. The new relationship is well confirmed by the Russian observational data (cf. Section 2.2) (not shown). Applying the $tanh$-function on the USAF climatology to compute the snow covered area over Eurasia (see thick dash-dotted line in Fig. 5.4) is in better agreement with the values derived from NOAA. Further evaluations have demonstrated that the results may not be improved by introducing the surface roughness (of vegetation) as a further independent variable. Additionally, its value may be afflicted with substantial uncertainties which led to the surface roughness being rejected.

5.1.4 Results from the 3-D experiment

At first, it should be emphasized that Eq. 5.12 was applied to both the forested and non-forested part of all land grid boxes. This is in contrast to the restrictions which have been imposed for deriving the formula (less than 10% forest fraction and fairly flat land). However, to facilitate interpretation, this inconsistency was taken into account. The main purpose of this section is to elucidate possible consequences of the above-suggested parameterization for the snow cover fraction.

The calculation of the snow cover fraction using the $tanh$-function in EXP1 (Eq. 5.12) substantially increases the snow cover fraction compared to the 10-year control experiment. This is shown in Fig. 5.6 for the winter season (DJF) in the Northern Hemisphere. The notable increase in snow cover fraction is mainly limited to regions with relatively thin snow decks, while the snow cover fraction $f_s$ of most regions in Siberia and northern Canada with a mean (modeled) $f_s$ above 80% during winter, differs only slightly between the control simulation CTRL and EXP1. This is consistent with the fact that the largest differences in the snow cover fraction, using either Eqs. 5.12 or 5.1, occur for thin snow packs. The maximum difference in $f_s$ is associated with $S_n = 1.6 \text{ cm}$ and amounts to more than $\Delta f_s = 25\%$. For thicker snow decks, the difference between the two expressions decreases rapidly and is negligible for $S_n > 10 \text{ cm}$ (cf. Fig. 5.5). The simulated increase in $f_s$ during winter is considerably higher than the annual variability which is represented in Fig. 5.6c. The differences between the snow cover fraction in CTRL and EXP1 are statistically significant, using the t-statistic test on the 95% level.

10-year annual means

Before going into detail, it is worthwhile to analyze annual (land) means. A number of simulated long-term means, averaged over all grid boxes in Eurasia and North America with $S_n > 0.1 \text{ cm}$ in February, are provided in Table 5.1.

Annual differences of over 5% are found for snow cover, surface albedo and reflected short wave radiation. The large percentage computed for the sensible heat flux is only due to annual means close to zero. Some conclusions can be drawn analyzing the above tabulated figures. The increased snow cover fraction and the related increase in reflected short wave radiation leads to less heating of the ground, which implies lower surface temperatures. Lower surface temperatures lead to a higher fraction of snow in the total precipitation and less snow melt which again increases the snow cover fraction (positive feedback).
The lower amount of available radiation near the ground may also reduce the magnitude of the hydrological cycle which is supported by a decrease in total precipitation. The decreased precipitation might be also associated with enhanced stability, which yields less turbulence and thus a decrease in the turbulent heat fluxes. This has been confirmed in the current experiment. Reducing the averaging area to Eurasia enhances most of the above differences between the control simulation and EXP1. This is probably due to the larger land mass of Eurasia compared to North America which favours, e.g., the occurrence of positive feedbacks between snow amount and temperature.

The impact of increasing surface albedo on land evapotranspiration $E$ is negative, which is in line with other sensitivity studies with 3-dimensional GCMs (Garratt, 1993). In the current experiment, $E$ decreases by 0.1 mm/day for an increase in surface albedo of 0.1, which compares well with the value determined by Mylne and Rowntree (1991).
However, the numerous studies quoted in Garratt (1993) rather suggest a distinctly higher sensitivity of evapotranspiration to surface albedo. The sensitivity of total precipitation to surface albedo is approximately 1.7 times larger than $\Delta E/\Delta \alpha$, in good agreement with the sensitivity studies reviewed in Garratt (1993). The response of evaporation to changes in surface albedo is consistent with the findings in off-line experiments using atmospheric forcing from the Cabauw site (cf. Section 4.3.5).

**Annual cycles**

More insight into some characteristics of EXP1 can be gained by investigating annual cycles averaged over the 10-year simulation period. Annual cycles of nine different variables (or combinations of them) are plotted in Fig. 5.7. It is evident that the annual amplitude difference in snow cover fraction (and directly related components) between CTRL and EXP1 increases when restricting the area, during winter, to predominantly snow covered regions in Eurasia and North America. The effect of the new parameterization of the snow cover fraction on net radiation fluxes, turbulent heat fluxes as well as on surface temperature are fairly patchy.

**Figure 5.7:** Differences between a number of surface variables of EXP1 and the control simulation for two specified regions. Monthly means are based on 10-year means. Solid: average over all land points excluding ice covered grid boxes; dashed: average over Eurasia and North America where $S_n > 0.1$ cm in February (based on the USAF snow depth climatology).
Table 5.1: Comparison of 10-year-means between EXP1 and the control climate (EXP1 - CTRL). Figures refer to averages over all land points where in February, $S_n > 0.1$ cm (according to the USAF snow depth climatology), excluding ice covered grid boxes. The last column depicts the percentage differences. Sign convention: downward fluxes are counted positively.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP1</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover</td>
<td>%</td>
<td>39.6</td>
<td>33.6</td>
<td>6.0</td>
<td>17.9%</td>
</tr>
<tr>
<td>Snow water eq.</td>
<td>cm</td>
<td>3.27</td>
<td>3.23</td>
<td>0.04</td>
<td>1.3%</td>
</tr>
<tr>
<td>Surface albedo</td>
<td></td>
<td>0.34</td>
<td>0.31</td>
<td>0.03</td>
<td>8.4%</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>Wm$^{-2}$</td>
<td>94.3</td>
<td>95.8</td>
<td>-1.5</td>
<td>-1.6%</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm$^{-2}$</td>
<td>128.1</td>
<td>127.2</td>
<td>0.9</td>
<td>0.7%</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>Wm$^{-2}$</td>
<td>-33.8</td>
<td>-31.4</td>
<td>-2.4</td>
<td>7.6%</td>
</tr>
<tr>
<td>Net LW, surface</td>
<td>Wm$^{-2}$</td>
<td>-51.0</td>
<td>-51.2</td>
<td>0.2</td>
<td>-0.4%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>272.68</td>
<td>272.95</td>
<td>-0.26</td>
<td></td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm$^{-2}$</td>
<td>-31.0</td>
<td>-31.7</td>
<td>0.8</td>
<td>-2.4%</td>
</tr>
<tr>
<td>Evapotranspiration</td>
<td>mm/day</td>
<td>1.04</td>
<td>1.07</td>
<td>-0.025</td>
<td>-2.4%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm$^{-2}$</td>
<td>-1.3</td>
<td>-1.6</td>
<td>0.37</td>
<td>-23.0%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>73.8</td>
<td>73.4</td>
<td>0.4</td>
<td>0.5%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>1.65</td>
<td>1.69</td>
<td>-0.04</td>
<td>-2.5%</td>
</tr>
<tr>
<td>10 m windspeed</td>
<td>ms$^{-1}$</td>
<td>3.58</td>
<td>3.62</td>
<td>-0.04</td>
<td>-0.9%</td>
</tr>
</tbody>
</table>

Fig. 5.7 illustrates that differences in the 2-m temperature and the sum of total shortwave radiation and downward directed longwave radiation flux, are well correlated. This is physically reasonable since this radiation flux provides the energy for heating the ground. The maximum impact on the reflected shortwave radiation occurs in March, due to the rapidly increasing insolation in spring and still considerably high differences in the snow cover fraction. The main contribution to the decreasing net radiation is attributed to the shortwave spectrum which is directly influenced by albedo changes. The decreased longwave upward radiation, related to lower surface temperatures, is balanced by enhanced shortwave downward radiation due to a reduced cloud water amount within the atmosphere (not shown).

**Zonal means of surface pressure**

The surface parameter which is most useful to investigate changes in the global circulation pattern, is the surface pressure. Figure 5.8 illustrates the difference in the zonal averages of surface pressure over the Northern Hemisphere (10-year averages). The large positive peak in winter south of the 60° parallel is mainly related to an intensifying of the extensive Asiatic high pressure system which develops due to the stronger continental cooling during winter. South of the winter snow line and over polar regions, only slight changes in the zonally averaged surface pressure can be detected. In spring and autumn, the pressure deviations are very similar except that during spring the peak is shifted about 8° south compared to the SON season. Note that the pressure difference in the transition seasons is opposite to what is expected when assuming that a cooling of the atmosphere leads to a higher surface pressure (as has been stated for the winter season). Analysing the global pressure patterns (not shown) reveals, however, that the main contribution to this negative bias in spring and summer is attributed to changes over sea and not over the continents. The most significant depression of pressure is found east of Eurasia and North America where the north Atlantic low pressure system and the north Pacific low develop during the (Northern Hemisphere) winter. During summer when nearly all snow has melted in North America and Eurasia, zonal pressure is only slightly affected by the change in the parameterization of the snow cover fraction. This may imply that the changes in the radiation balance and circulation patterns due to increased snow cover fraction have
“no memory” which is reflected in the small differences in the pressure pattern. This hypothesis is supported by substantially smaller changes in the 10-m wind fields (not shown) in summer than in spring or autumn. However, this result may be questioned owing to a substantially higher interannual variability of the surface pressure during spring and autumn than during summer.

Representativity of 10-year-means

Experiment EXP1 has been selected to show that ten-year means produce predominantly the same results as when the conclusions would be based on the first or second five years of the model simulations. Figure 5.9 displays the seasonal zonal differences of snow cover fraction between CTRL and EXP1. The zonal deviations differ only slightly when replacing the 10-year averages by 5-year averages. Largest differences between the various time periods occur where the interannual variability is significant (Fig. 5.6c). The same procedure, but applied to monthly means instead of seasonal averages, reveals similar results. However, for some evaluations it may happen that the 10-year-means are not sufficient for calculating, from a statistical point of view, accurate averages. Problems particularly arise when investigating local processes and temporally highly variable components.

Summary

A new parameterization of the snow cover fraction is derived from global datasets of ground-based snow water equivalent and remotely sensed snow cover fraction. The developed relationship between snow cover fraction and snow water equivalent includes the \textit{tanh}-function and is applicable only to mainly flat grid-elements. Most significant differences in the snow cover fraction between EXP1 and the control climate are inferred in areas with relatively thin snow packs. The relationship has been turned out to be applicable to both the T42 and T106 horizontal resolution.
5.2 EXP2: Snow cover fraction formula as implemented in the Météo-France climate model

It has been demonstrated in numerous simulations and observational studies that the snow cover fraction plays an important role in modifying regional and possibly remote climate through changes in the surface energy balance (e.g., Yeh et al., 1983; Barnett et al., 1989). It is therefore appropriate to more closely investigate a further parameterization of snow cover fraction $f_s$.

As is shown later, the winter surface albedo over mountainous regions is likely to be overestimated in ECHAM. Therefore, in order to account for the reduction in $f_s$ over mountainous regions, a new parameterization, which is applied in the Météo-France climate model, has been tested in detail. Douville et al. (1995b) suggest the following expression which is based on the standard deviation of the subgrid orography ($\sigma_z$ in [m]):

$$f_s = \frac{S_n}{S_n + S_n^*} \sqrt{\frac{S_n}{S_n + \text{max}(1.0 \text{ m}, 0.15\sigma_z)}},$$

(5.13)

where $S_n^* = 0.01$ m. The computation of $\sigma_z$ is based on the global high-resolution topographical dataset compiled by the U.S. Geological Survey’s EROS data center, as detailed in Section 5.10. Eq. 5.13 accounts for the process that, during the melting period, snow
patches remain on the northern slopes of the mountains, whereas the sides exposed to solar radiation are snow free. In addition, in steep mountains, the bare rock can be exposed due to snow gliding even during snow rich winters. This may substantially lower snow cover and, therefore albedo, in some regions. The reduction in snow cover extent is substantial (Fig. 5.10a). Assuming $S_n = 0.2$ m over sufficiently steep mountain areas ($\sigma_z \geq 1000$ m), the snow cover fraction is about 70%, compared to more than 90% over flat land areas. In Figure 5.10b and 5.10d, the distribution of $\sigma_z$ of all land grid points and the Himalayan area is displayed. More than 50% of that huge mountain ridge is covered with grid boxes with $\sigma_z \geq 500$ m. Note that approximately 25% of the Earth's land surface has a standard deviation of the subgrid orography larger than 300 m, thereby pointing out the relevance of incorporating $\sigma_z$. In addition, notice that these regions are mostly linked to mountainous (and, therefore, higher/colder) areas that are likely to be snow covered during winter-time. It should, again, be stressed that all the above calculated values are based on the global height distribution with a resolution of approximately 1 km.

Figure 5.10c gives some reasoning for the inclusion of the new parameterization. The surface albedo of the Himalayan region, using different approaches, is displayed. The marked difference between the observed (SRB) and modeled mean surface albedos is evident. While the ECHAM/T42 experiment of the current climate simulates a pronounced annual cycle with an amplitude of about 0.2, the SRB data varies only slightly. The albedo evolution labeled 'USAF1' has been computed using the USAF snow depth climatology, but all other conditions are as in ECHAM4. These include the parameterization of the snow albedo, the calculation of the background albedos and surface temperatures, the transformation equation to compute the snow cover fraction as well as the forest fraction. The albedo, derived from 'USAF1', is about 0.05 lower in winter compared to the control simulation. Since the computation is based on monthly values, minor differences may be due to averaging problems. The curve 'USAF2' is based on the same assumption as 'USAF1' except that the snow cover fraction is calculated using the new parameterization (Eq. 5.13). The marked contrast between the two curves reflects the strong sensitivity of the albedo with respect to the subgrid orography. Note that the application of the new expression, using the observed snow depth and the other land surface properties as in ECHAM4, generates a surface albedo which is in good agreement with the SRB observation.

During summer, where the USAF climatology provides snow free conditions, the SRB albedo is significantly higher than in 'USAF1' and 'USAF2'. This may be caused by the problem that the measurements sites, used to compile the snow depth climatology, do not represent the very high mountain regions with permanent snow decks.

Summarized, it can be stated, that it is strongly recommended to introduce the subgrid scale orography for the description of the snow cover fraction, especially over steep mountainous areas.

5.2.1 Overview of modifications

In this experiment, Eq. 4.32 is replaced by the formula as used in the Arpège climate model and described in Douville et al. (1995b). In the following, it is shown that Eq. 5.13 provides substantially better surface albedos over snow covered mountainous regions.

5.2.2 Results from the 3-D experiment

At first, the snow cover anomalies as well as the large-scale changes between the control run and EXP2 are briefly discussed. Later on, the large-scale differences of some variables
Figure 5.10: a): Snow cover fraction as implemented in the Météo-France climate model (Eq. 5.13). b): Distribution of \( \sigma_z \) for all land points. The classes '1' to '7' embrace the following ranges for \( \sigma_z \): class 1: \( \sigma_z < 100 \) m; class 2: \( 100 \) m \( \leq \sigma_z < 200 \) m; class 3: \( 200 \) m \( \leq \sigma_z < 300 \) m; class 4: \( 300 \) m \( \leq \sigma_z < 500 \) m; class 5: \( 500 \) m \( \leq \sigma_z < 700 \) m; class 6: \( 700 \) m \( \leq \sigma_z < 1000 \) m; class 7: \( \sigma_z \geq 1000 \) m. c): Monthly mean surface albedos for the Himalayan region. 'SRB': remotely sensed surface albedo of the SRB Project (1984-1990); 'ECHAM4': simulation of the current climate with ECHAM4/T42; 'USAF1' and 'USAF2': modified albedos based on the USAF snow depth climatology (see text). d): As (b), but limited to the Himalayan area.

are presented. Since this experiment is especially suited to investigate the anomalies associated with a changed snow cover fraction in the Himalayas, the monsoon anomalies over India are discussed within the framework of this experiment.

As shown in the previous section, substantial differences in the surface albedo are expected over rough mountains, which is confirmed in Fig. 5.11. The plot displays the surface albedo in the control simulation (CTRL) during winter (DJF), its difference, and the normalized difference between CTRL and EXP2. The normalized difference between CTRL and any modified model simulation EXPX (X = 1,...,10) is

\[
\Delta = \frac{EXPX - CTRL}{\sigma},
\]

(5.14)

where \( \sigma \) is the monthly standard deviation computed from the 10-year control run. The measure \( \Delta \) is closely related to the t-statistic generally used to test model significance (e.g., Chervin and Schneider, 1976).
It is evident that the (orographically) rough mountain areas, as the Himalayas, show a
significant positive bias in the surface albedo which is also found in smaller mountains such
as the Alps, the Caucasus or the Scandinavian mountain chain. Over the Himalayas, the
mean albedo is in better agreement with the observation when allowing the reduction of
the snow cover fraction due to the subgrid scale orography. This is illustrated in Fig. 5.12.

Statistically significant albedo differences are generally found over the Himalayas and large
parts of northern Siberia as well as Alaska and the Canadian Archipelago in Fig. 5.11c. One
of the main reasons for the statistically significant albedo bias of the northern territories
is the low interannual variability of the snow water equivalent and thus, the snow cover
fraction.

The correlation between the albedo anomalies and the standard deviation of the sub-
grid scale orography ($\sigma_z$) is not excellent. The correlation coefficient between the albedo

FIGURE 5.11: Albedo anomalies over the Northern Hemisphere for winter (DJF), a)
Control run (CTRL); b) albedo difference between EXP2 and CTRL; c) as (b), but
normalized differences according to Equation 5.14.
difference and $\sigma_z$ equals 0.65 for the three winter months (DJF) for Eurasia and North America. A key factor which determines the quantitative albedo bias is, in addition to $\sigma_z$, the snow water equivalent. The sensitivities of the snow cover fraction to $\sigma_z$ are presented in Fig. 5.13. The effect on the snow cover fraction decreases with increasing $\sigma_z$. The sensitivity is considerably higher for thin snow packs. The discontinuity for $\sigma_z = 666.6 \text{ m}$ is rather arbitrary and could be avoided by allowing the factor $\max(1.0, 0.15 \sigma_z)$ to be larger than 1.

**Figure 5.13:** Derivative of the snow cover fraction $f_s$ with respect to the standard deviation of subgrid orography. Calculations are based on Eq. 5.13. Curves are given for snow water equivalent of $S_n = 1 \text{ m}, 0.5 \text{ m}, 0.1 \text{ m}$ and 0.01 m. $f_s$ shows no sensitivity to $\sigma_z$ when $\sigma_z > 666.6 \text{ m}$.

**Large-scale means**

It is unreasonable to investigate differences between the modified simulation and the control run on a global basis as climate anomalies over large areas which are snow free throughout the year are small and of limited importance. The annual means in Table 5.2 are thus averages over the same areas as in Table 5.1, i.e. Eurasia and North America where in February, at least a thin snow deck is established. Comparing the results of EXP1 and
EXP2 (Tables 5.1 and 5.2) reveals that the differences are very similar but have opposite signs. Thus, over the Northern Hemisphere, the increase of the snow cover due to the introduction of the tanh-function for snow cover fraction (Eq. 5.12) is compensated by the decrease of the snow cover fraction over mountainous regions. Thus, it can be concluded that the interpretation of the results in EXP1 can be adopted for EXP2, but with inverse signs: the decreased snow cover leads to a lower albedo and stronger heating of the surface due to increased available radiation. The higher surface temperature and higher wind speeds lead to enhanced turbulent heat fluxes and thus precipitation. This may be related to an intensification of the hydrological cycle.

However, the regions, where most pronounced changes occur, differ significantly between EXP1 and EXP2. Note that Equation 5.12 was applied to all land points, which contradicts the conditions which have been required for the derivation of the parameterization. Hence, limiting the application of Eq. 5.12 to more or less flat and to forest free grid boxes, would certainly result in distinctly lower anomalies in EXP1.

Table 5.2: Comparison of 10-year-means between the control climate and EXP2. Figures refer to averages over all land points where in February $S_n > 0.1$ cm (according to the USAF snow depth climatology), excluding ice covered grid boxes. The last column depicts the percentage differences. Sign convention: downward fluxes are counted positively.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP2</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover</td>
<td>%</td>
<td>28.6</td>
<td>33.6</td>
<td>-5.0</td>
<td>-14.9%</td>
</tr>
<tr>
<td>Snow water eq.</td>
<td>cm</td>
<td>3.14</td>
<td>3.23</td>
<td>-0.09</td>
<td>-2.9%</td>
</tr>
<tr>
<td>Surface albedo</td>
<td></td>
<td>0.29</td>
<td>0.31</td>
<td>-0.02</td>
<td>-6.0%</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>Wm$^{-2}$</td>
<td>97.5</td>
<td>95.8</td>
<td>1.7</td>
<td>1.8%</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm$^{-2}$</td>
<td>126.3</td>
<td>127.2</td>
<td>-0.9</td>
<td>-0.7%</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>Wm$^{-2}$</td>
<td>-28.7</td>
<td>-31.4</td>
<td>2.7</td>
<td>8.5%</td>
</tr>
<tr>
<td>Net LW, surface</td>
<td>Wm$^{-2}$</td>
<td>-51.5</td>
<td>-51.2</td>
<td>-0.3</td>
<td>0.6%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>273.23</td>
<td>272.95</td>
<td>0.28</td>
<td>0.8%</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm$^{-2}$</td>
<td>-32.4</td>
<td>-31.7</td>
<td>0.6</td>
<td>2.0%</td>
</tr>
<tr>
<td>Evapetrapiration</td>
<td>mm/day</td>
<td>1.09</td>
<td>1.07</td>
<td>-0.02</td>
<td>-1.9%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm$^{-2}$</td>
<td>-2.4</td>
<td>-1.6</td>
<td>-0.76</td>
<td>46.7%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>73.3</td>
<td>73.1</td>
<td>-0.09</td>
<td>-0.1%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>1.70</td>
<td>1.69</td>
<td>0.01</td>
<td>0.8%</td>
</tr>
<tr>
<td>10 m windspeed</td>
<td>m/s$^{-1}$</td>
<td>3.65</td>
<td>3.62</td>
<td>0.03</td>
<td>0.8%</td>
</tr>
</tbody>
</table>

The anomalies in the total surface radiation budget are mainly due to the positive bias in the shortwave spectrum while in the longwave range, the deviation in the upward and downward directed fluxes nearly compensate each other. This feature is also maintained on a monthly basis (not shown). The deviations are more pronounced during winter whereas during the snow free period, the differences between the CTRL and EXP2 are minimal. This means, that the atmosphere has a "short memory" concerning snow cover fraction anomalies.

Some considerations of the monsoon circulation

It may be somewhat hazardous to investigate the changes of the monsoon circulation between the control simulation and EXP2 since the ten-year simulations may be too short for such an investigation. In addition, interannual variability of the key climate variables are generally large in South East Asia which partially hinders the anomalies from being statistically significant. Furthermore, the sea surface temperature (SST) are prescribed in both experiments. SST anomalies are also known to be important for the strength of the monsoon circulation (Elliott and Angel, 1987). Nevertheless, in the following paragraphs some interesting results are outlined. At first, a brief introduction into the monsoon process...
is presented. Then the model results are shown, and finally the discussion with respect to a possible change in the monsoon circulation, focused on the Indian subcontinent, is provided.

a) Introduction

The Asian monsoon, which affects the Indian subcontinent and southeast Asia, is a result of several factors, mainly:

- The land-sea temperature contrast, particularly after snow melt, initiates the surface circulation from sea to land which leads to the summer monsoon pattern.
- Released latent heat (through condensation) by deep convection is a key factor in the stabilisation and conservation of the depression over India/Pakistan and hence the high pressure area over the Tibetan Plateau.
- the influence of the Tibetan Plateau: during winter, the Tibetan Plateau acts as a tropospheric cooling element when it is covered by snow, and as a heating element when it is free of snow. Cooling above the Tibetan Plateau initiates a cold anticyclone below 600 hPa that causes dry, stable conditions. When the plateau is heated during the snow free period, a high pressure system develops in the upper troposphere.
- the location of the ITCZ to southeast Asia. From its winter position of 15°S, it moves to the north in March and reaches the 25°N when the summer monsoon is fully developed (in mid-July).

Typical for the Asian monsoon is the development of the Tropical Easterly Jet (TEJ). Due to the shift of the ITCZ to the north, the humid low-level south-westerlies penetrate into India. The increased continental convection returns the upper air flow to the south, which is then deflected to the west by the Coriolis force to produce a strong Tropical Easterly Jet (TEJ) located at about 10° - 15°N.

b) Model results

The effect of a new parameterization for the snow cover fraction on a number of variables is shown in Fig. 5.14. The large negative snow cover bias in the Himalayas during winter yields a corresponding lower surface albedo of approximately 10%. This enhances the absorbed solar radiation at the surface, which increases the surface temperatures up to more than 1°C in January. The higher available radiation increases the (absolute) magnitude of the turbulent heat fluxes by approximately 10 Wm⁻² in winter which compensates the positive bias in the net surface radiation. The change in the integrated cloud water and the precipitation shows only slight variations throughout the year and is almost negligible. The higher latent heat flux provides the main contribution to the reduction of snow depth. The impact on snow melt and snowfall (not shown) are negligible.

The Indian subcontinent (dashed line in Fig. 5.12), in contrast, shows a completely different anomaly pattern. Snow cover does not occur over India and thus, there is no effect on the snow cover fraction, the surface albedo and the snow water equivalent. In spite of the unchanged surface albedo, a significant increase in net shortwave radiation and hence, total surface radiation in summer has been revealed. The reason for this positive radiation bias is a substantial decrease in the vertically integrated cloud water during the two months of June/July. In June, the cloud water amount decreases by 32% and in July by 18%. The deficiency in cloud water leads to a reduced absorption of shortwave radiation which lowers global radiation by 20.7 Wm⁻² and 22.0 Wm⁻² in June and July, respectively. It is evident that the reduction in cloud water implies less precipitation (-16% in June) and hence, reduced runoff (close to -30%). The enhanced temperatures favour more instable
conditions and therefore, an intensified sensible heat flux. The total turbulent heat flux changes by $-11.3 \text{ Wm}^{-2}$ and $-13.6 \text{ Wm}^{-2}$ in June and July, respectively. The averaged horizontal mean wind speed (June/July) in India decreases from 4.5 m/s to 4 m/s which corresponds to more than 10%.

For the monsoon, it is important to know the condition in the entire atmosphere since surface variables are not sufficient to characterize the monsoon. One of the quantities which is best qualified to capture the structure of the atmosphere, is the geopotential. The latitude-height section of the differences in the geopotential is shown in Fig. 5.15. The geopotentials are means from 70°W to 100°W. A positive bias of about 5 m is established close to the surface at 30°N, while above 400 hPa, a strong negative anomaly develops. This negative bias attains its maximum at 150 hPa, which corresponds to a height of approximately 13 km, i.e. the anticyclone in the high troposphere is too weak.

c) Discussion

Over India the monsoon rain onset occurs at the beginning of June in southeast India while in northwest India this date shifts to the end of June. All the changes found in the previous section fit well the assumption that EXP2 simulates a weaker (summer) monsoon than the
control simulation. A "weak" monsoon can be characterized by a weakened circulation and lower precipitation rates.

These changes during the summer season include:
- decrease in wind speed
- decrease in cloud water and precipitation
- enhanced radiation input and higher surface temperatures.

In Figure 5.16, the mean monthly precipitation rates over the Indian subcontinent are plotted. The observations are taken from the Global Precipitation Climatology Project (GPCP) (Rudolf et al., 1996 and Section 2.1). The control simulation and EXP2 show similar monthly precipitation amounts, excluding the monsoon season from June to August. In spring, before the summer monsoon onset, both model experiments significantly overestimate the precipitation. The control simulation significantly overestimates rainfall during the monsoon season in June/July, whereas EXP2 captures the monsoon precipitation surprisingly well. After the monsoon withdrawal over India in October, ECHAM
substantially underestimates the observed precipitation. The anomalies in the geopotential height (Fig. 5.15) are in line with the common theory for a "weak" summer monsoon over the Indian subcontinent. A strongly developed monsoon is generally related to a strong depression over India/Pakistan and to a well developed anticyclone in the upper troposphere over the Tibetan Plateau which generates an intense Tropical Easterly Jet (TEJ). However, the simulation indicates that the anticyclone, as well as the low, weaken which implies an attenuation of the monsoon circulation. 

The reason for the attenuation of the monsoon circulation, however, is opposite to the common view regarding the summer monsoon: A diminuation or delay of the normal warming of the land surface yields a delayed or weakened monsoon circulation. In EXP2, the snow cover fraction was reduced over the Himalayas compared to the control run which led to an excessive radiative heating and increasing surface temperatures. This would, according to the common theory, yield a stronger monsoon.

An explanation of this unexpected result obtained with the simulation EXP2 may be found in Barnett et al. (1989) who investigated the relation between the Eurasian snow cover and the monsoon development. They suggest two hypotheses which may be important for the development of a "good" or "poor" monsoon: The albedo feedback and the hydro-meteorological feedback (variation in snow depth and its subsequent melting and evaporation). They rejected the albedo feedback but found the hydro-meteorological feedback to be significant. In the current experiment the snow depths are not changed; only the snow cover fraction is modified and thus, the surface albedo.

To summarize, the present study indicates that a decreased snow cover fraction (but a similar snow depth) over the Himalayas yields a "poor" monsoon with a drier and hotter climate over India. Most of the variables, including the stratification of the entire atmosphere, demonstrate a weakening of the (summer) monsoon circulation over the Indian subcontinent. The findings do not contradict the results of Barnett et al. (1989).

5.3 EXP3: Polynomial temperature dependence of snow albedo

Most of the GCMs and snow models allow for a parameterization of the snow albedo $\alpha_s$ which is dependent on the temperature, e.g. the rate of snow aging. However, relatively few of these models explicitly incorporate the temperature dependence. A short review of some available relationships between snow albedo and temperature dependence is provided in this section.

In ECHAM, the snow albedo increases linearly from the melting point ($\alpha_s = 0.4$) to 0.8 for a temperature of $-10^\circ$C.

Following Cogley and Henderson-Sellers (1990), the albedo in the visible and near infrared spectrum is parameterized as:

$$\alpha_{s, VIS} = 0.9 - 0.2T_m^3$$
$$\alpha_{s, NIR} = 0.8 - 0.16T_m^3$$

where $T_m$, the temperature dependent metamorphism factor, is given by

$$T_m = \begin{cases} 
0 & T_s \leq 263.16 \, K \\
0.1(T_s - 263.16) & 263.16 < T_s < 273.16 \\
1.0 & T_s \geq 273.16 
\end{cases}$$

and $T_s$ denotes the snow temperature. The resultant albedo is shown in Fig. 5.17, assuming the incident solar radiation to be equal in the near infrared ($\lambda > 0.7 \, \mu m$) and visible.
waveband. This algorithm has been adapted in the land surface model BASE (Best Approximation of Surface Exchange, described in Slater et al. (1998)) which is a full GCM land surface scheme that can be applied to all types of land surfaces.

The second version of the Simple Biosphere model (SiB2), described in Sellers et al. (1996a), assumes that the snow reflectance is reduced by around 60% as the snow melts. Therefore,

\[
\begin{align*}
\alpha_{S, VIS} &= 0.8T_f \\
\alpha_{S, NIR} &= 0.4T_f,
\end{align*}
\tag{5.17}
\]

where

\[
T_f = 1.0 - 0.04(T_s - 263.15) ; 0.6 \leq T_f \leq 1.
\]

Only a few models neglect completely the temperature dependence. In the Europa-modell (EM), for example, the snow albedo is represented by a constant equal to 0.7. Fig. 5.17 shows the large discrepancies in different formulations describing the relationship between snow albedo and temperature. The most pronounced sensitivity with respect to temperature produces the parameterization applied in ECHAM. In-situ and remotely sensed observations strongly point out that the upper value (\(\alpha_s = 0.6\)) used in SiB2 is too low, whereas the maximum snow albedo used in ECHAM (\(\alpha_s = 0.8\)) appears to be a reasonable assumption. The model of Cogley and Henderson-Sellers (1990) is the only model, discussed here, which proposes a non-linear relationship between snow albedo and temperature.

Several snow models implicitly account for the temperature effect of the snow albedo, using a snow age factor which usually is different for melting and non-melting conditions (see Section 5.5.1).

### 5.3.1 Overview of modifications

In this model run, the linear relationship between snow albedo and surface temperature (cf. Section 4.2.3), as currently used in ECHAM4, is replaced by a polynomial relationship (cf. Eq. 5.18). This polynomial relationship is derived from a Russian dataset which provides 3-hourly measurements of both climatological and hydrological variables at 6
stations during 1978 - 1983. The formula strictly applies only to non-forested areas. In EXP3, however, the snow covered forest fraction is computed the same way.

### 5.3.2 Polynomial relationship between temperature and snow albedo

The proposal for a new formulation of the temperature dependence is mainly based on the comprehensive dataset of six Russian stations covering the period 1978 - 1983 (see Section 2.2). The snow albedo is displayed as a function of the screen temperature, measured approximately two meters above the ground (Fig. 5.18). To ensure a closed snow deck, only those measurements with a snow depth thicker than 10 cm are considered. Moreover, to avoid incorrect albedo measurements, observations with a global radiation lower than 20 Wm$^{-2}$ were omitted.

![Graphs showing snow albedo as a function of temperature for 4 selected Russian stations.](image)

**Figure 5.18:** Snow albedo as a function of surface temperature for 4 selected Russian stations. Dashed: linear relationship proposed in the ECHAM GCM; solid: new parameterization equation (Eq. 5.18). The plots are based on 3-hourly observations during 1978 - 1983 with snow depth thicker than 10 cm.

From the measurements, it is evident that a linear relationship, as used in ECHAM, does not represent a good choice for the desired relationship. It is more reasonable to assume that the sensitivity of snow albedo with respect to surface temperature increases when approaching freezing point which applies to the following expression

$$\alpha_s = 0.5 + a_1 T + a_2 T^2 + a_3 T^3 + a_4 T^4,$$

(5.18)
where \( a_1 = -0.07582627 \, K^{-1} \), \( a_2 = -5.5360168 \times 10^{-3} \, K^{-2} \), \( a_3 = 5.2966269 \times 10^{-5} \, K^{-3} \) and \( a_4 = 4.2372742 \times 10^{-6} \, K^{-4} \). This relationship was derived with the CRAMER-function implemented in IDL which solves an \( n \) by \( n \) linear system of equations using Cramer's rule. This function was applied with the following snow albedos: \( \alpha_s(-10^\circ C) = 0.8 \), \( \alpha_s(-8^\circ C) = 0.79 \), \( \alpha_s(-5^\circ C) = 0.75 \) and \( \alpha_s(0^\circ C) = 0.5 \). Note that the value at the melting point was increased from 0.4 (ECHAM) up to 0.5, according to literature (Briegleb and Ramanathan, 1982; Robinson and Kukla, 1984; Verseghy et al., 1993 and Walland and Simmonds, 1996) and Russian observations. Eq. 5.18 is displayed as solid lines in Figure 5.18. Note, that the curves were slightly adapted to account for the restriction that surface temperature in ECHAM is not allowed to exceed \( T_s = 0^\circ C \) for snow covered conditions. It should be stressed that the ECHAM parameterization is based on the surface temperature whereas the observations are made on a screen height of about 1.3 m above the ground. This reveals a severe problem deriving a relationship between temperature and snow albedo. Most temperature measurements refer to a height of approximately two meters above the ground, but the state of the snow deck’s surface depends particularly on the snow surface temperature which is rarely measured during field campaigns. Further problems arise due to the restriction in ECHAM of not allowing for surface temperatures above 0°C for grid boxes covered with snow: During snow melt, snow free spots coexist with snow covered areas. In addition, over snow free regions the surface temperature may locally exceed the freezing point and this may lead to a mean grid box surface temperature of over 0°C. Further, the composition of the snow deck is not instantaneously adapted to actual temperature since metamorphism processes lead to some inertia.

Baker et al. (1990) found in a long-term dataset of daily albedos that temperature is not a good predictor of snow albedo since temperature varies quickly and its effect can be cumulative. However, the Russian data have shown that it is more appropriate to choose a polynomial rather than a linear relationship between surface temperature and snow albedo. This is confirmed by using radiation measurements performed at the ETH-camp on the Greenland ice sheet during 1990/1991 (see Konzelmann, 1994). However, from Fig. 5.19 it is obvious that the relationship between \( \alpha_s \) and the temperature can not be reproduced sufficiently well using the same polynomial function for every condition.
Albedo of snow covered forests

The polynomial albedo-temperature relation is also adapted to snow covered forests. In addition to the change in the functional form, the minimum and maximum albedo of snow covered forests are reduced by 0.1. Thus, the range in the surface albedo, which is between 0.3 and 0.4 in ECHAM4, is reduced to 0.2 - 0.3. This modification is supported by various in-situ measurements (e.g., Betts and Ball, 1997; Robinson and Kukla, 1984) which have revealed very low mean surface albedos for snow covered boreal forests in the order of 0.2. These studies all suggest mean snow covered forest albedos of 0.4 to be untrustworthy even after heavy snowfalls.

5.3.3 Results from the 3-D experiment

The 10-year 3-D experiment shows that the effect of a polynomial relationship between snow albedo and temperature on the surface climate does not statistically differ from the control climate. In the following, Eq. 5.18 and the albedos of snow covered grid boxes with varying forest fraction are discussed. Further, the effect of Eq. 5.18 on the surface albedo is assessed in the framework of a long-term 3-D model simulation.

Modification in the albedo pattern

Fig. 5.20 illustrates the albedo differences between EXP3 and the control simulation in winter (DJF). The dipole pattern of the surface albedo anomalies over the snow covered regions is clearly visible in Fig. 5.20b. The boundary line between the region with a negative and positive albedo bias, respectively, can be approached by the contour line for $a_f = 0.4$ (forest fraction = 40%), drawn as a thick solid line in Fig. 5.20b. The map shows that cold winter climate in densely forested areas (e.g., Taiga) yields negative albedo anomalies while more moderate climates to the south with less forests lead to positive differences. The negative albedo differences are generally statistically significant, i.e. absolute normalized differences in Fig. 5.20c are distinctly larger than 1, whereas in the regions with a moderate climate, the interannual variability in the snow height affects the snow albedo more than the inclusion of Eq. 5.18.

A detailed analysis of Equation 5.18 provides more insight into the albedo changes discussed in the previous paragraph (Fig. 5.21). For forest free grid elements, the suggested polynomial relationship between snow albedo and temperature yields enhanced reflectivities for the entire temperature range of -10°C $\leq T \leq 0$°C. The highest absolute difference is obtained for $T = -3.9$°C. Increasing the forest fraction causes the two curves to approach each other. A forest fraction $a_f = 0.25$ implies higher albedos only for temperatures below -9°C, following the ECHAM parameterization. The corresponding temperature increases to -7.9°C for $a_f = 0.5$. For grid boxes with more than 70% forests, the new relationship yields lower reflectances for the entire temperature range below freezing point. This is due to the reduction of the albedos of snow covered forests from the range between 0.3 and 0.4 to 0.2 and 0.3. The preceding results explain the features in Fig. 5.20b: The northern parts of Eurasia and Canada with negative albedo biases are characterized by mean temperatures below -10°C and mostly forested landscapes. The regions adjacent to the south, however, are moderately forested and generally have mean winter temperatures above -10°C.
Normalized differences

This section demonstrates that the changes in the monthly means between the control simulation and EXP3 are generally negligible on both large and local scales. This supports the hypothesis that local modifications in the albedo pattern, and, hence the radiation balance, are efficiently compensated by the atmospheric circulation. This is illustrated in Table 5.3 which contains the mean difference in a number of surface variables normalized by the standard deviation of the control run (cf. Eq. 5.14). A value of 1 implies a mean difference which has the same magnitude as the standard deviation in the control simulation. Values above 1 indicate that, for the specified region, the change is statistically significant. All values are averages over that area where the difference in the surface albedo between EXP3 and the control simulation exceeds 0.01 during winter (DJF). That area corresponds to the regions with darker colours in Fig. 5.20b.

The values given in Table 5.3, reflect that the deviations in EXP3 from the control run
Figure 5.21: Comparison of grid-averaged snow albedo for the ECHAM4 parameterization. Dashed line: ECHAM4 parameterization (Eqs. 4.31 and 4.33, Section 4.2.3); solid line: Eq. 5.18. Forest fractions: a) $a_f = 0.0$; b) $a_f = 0.25$; c) $a_f = 0.5$; d) $a_f = 1.0$.

Table 5.3: Comparison of 10-year-means between the control climate and EXP3. Figures refer to averages over all land points where in winter (DJF), $\Delta \alpha > 0.01$ with $\Delta$ indicating the difference between the control simulation and EXP3. The second and third columns depict the differences $\Delta = \text{EXP3} - \text{CTRL}$, normalized by the standard deviation ($\sigma$) computed from the 10-year control run (cf. Eq. 5.14) for winter (DJF) and the entire year, respectively. Sign convention: downward fluxes are counted positively.

<table>
<thead>
<tr>
<th>parameter</th>
<th>Control (annual mean)</th>
<th>$\Delta / \sigma$ (DJF)</th>
<th>$\Delta / \sigma$ (annual mean)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover</td>
<td>20.3%</td>
<td>0.25</td>
<td>0.18</td>
</tr>
<tr>
<td>Snow water eq.</td>
<td>1.5 cm</td>
<td>0.20</td>
<td>0.19</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>0.26</td>
<td>0.49</td>
<td>0.45</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>36.2 Wm$^{-2}$</td>
<td>0.27</td>
<td>0.12</td>
</tr>
<tr>
<td>Global radiation</td>
<td>59.0 Wm$^{-2}$</td>
<td>-0.08</td>
<td>-0.03</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>-34.5 Wm$^{-2}$</td>
<td>-0.44</td>
<td>-0.40</td>
</tr>
<tr>
<td>Net LW, surface</td>
<td>-61.8 Wm$^{-2}$</td>
<td>0.13</td>
<td>0.09</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>277.9 K</td>
<td>-0.09</td>
<td>-0.05</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>1.6 mm/day</td>
<td>0.06</td>
<td>0.04</td>
</tr>
</tbody>
</table>

are minimal, i.e. all surface variables, including surface albedo, alter insignificantly when introducing the polynomial relationship between snow albedo and temperature. Applying the t-statistic test to the variables also indicates that the changes are, with a few isolated local exceptions, statistically insignificant. It is evident that the normalized difference is largest for the surface albedo since this quantity is directly influenced by Eq. 5.18.
5.4 EXP4: Splitting of the total surface albedo into its VIS- and NIR-component

Most land surface schemes do not distinguish between the snow albedo in the visible (VIS) and near-infrared (NIR) waveband or assume the total broad-band albedo to be the arithmetic average of the VIS-albedo ($\alpha_{\text{VIS}}$) and NIR-albedo ($\alpha_{\text{NIR}}$). $\alpha_{\text{NIR}}$ is the mean albedo for near-infrared radiation (0.7 - 4.0 $\mu$m) and $\alpha_{\text{VIS}}$ is the same for visible radiation (0.25 - 0.7 $\mu$m). In this chapter it is demonstrated that substantial errors in the albedo pattern occur when assuming identical radiation fluxes in the VIS and NIR spectrum.

The main problem which had to be solved, comprised the compilation a global dataset on the T42-grid for both the VIS- and NIR-albedo, thereby distinguishing between snow free and snow covered conditions.

5.4.1 Overview of modifications

The aim of this experiment is to distinguish between the radiation flux in the VIS and NIR spectrum. Hence, the background surface albedo field was replaced by global datasets comprising of the albedo in the visible and NIR spectra. The spectral background albedos are computed with the SiB2 model (Sellers et al., 1996a) on a regular 1° x 1°-grid. The fields are then interpolated on the T42-grid, using area-weighed interpolation. The dependence of the water albedo on the zenith angle is adopted from Briegleb and Ramanathan (1982).

The own contribution to the incorporated modifications primarily consists of the stipulation of minimum and maximum values for $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$ over snow- and ice-covered regions (cf. Table 5.4). Moreover, a comparison between spectral background albedos based on two different methods is given.

Compilation of global snow free spectral albedo fields

The compilation of a global dataset for $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$ under snow free conditions can be achieved by two methods which are described in the following section.

Method 1 is based on an ecotype classification and an allocation of “typical” albedos: The application of Method 1 requires the allocation of “typical” spectral albedos in the VIS and NIR range to each vegetation type. In the present study, the ecotype classification of Olson et al. (1983) was used. Unfortunately, no dataset for allocating each Olson vegetation type a value for $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$, is available. One of the best available allocations of spectral albedos to vegetation types has been developed for BATS (Dickinson et al., 1993), where 18 land cover/vegetation types are distinguished. Therefore, a further allocation between the Olson and BATS vegetation types is required. This can be accomplished using results presented in Claussen et al. (1994).

The main disadvantage of this method is that each grid cell is associated with one dominant vegetation type. The assessment of “typical” values of $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$ for all the vegetation types causes further uncertainties. Additional problems arise due to the high reflectivity of some deserts, which are not correctly captured when applying the procedure of Method 1. Furthermore, this method is not suited to reproduce the annual cycle of the snow free surface albedo but provides only annual means. It has thus been decided to compile the global dataset for the spectral albedos, as described in the next section.

Method 2 based on SiB2: The SiB2-model, described in detail in Sellers et al. (1996a), provides an excellent tool for compiling a global dataset for both the $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$. 


D. Dazlich from the Colorado State University kindly provided the global fields of the albedos on a 1° x 1° mesh in both shortwave bands. His calculations are based on an offline version of the radiation model in SiB2. This sophisticated canopy radiation transfer model includes, for the albedo computation, the influence of the leaf optics, leaf angles as well as the dead leaf index, green leaf index and stem-area index (for definitions see Sellers et al., 1996a). The soil albedo over deserts are based on ERBE data. The problems inherent to Method 1 are responsible for preferring Method 2 to Method 1. Hence, all the following results are based on surface albedos derived with Method 2.

**Global distribution of spectral albedos**

The albedo maps are shown in Figure 5.22. Simple area-averaging interpolation is used to generate the global fields on the T42-mesh. The bright deserts over the Sahara and the Arabian peninsula have high spectral albedos in both the visible and NIR range. The albedo in the NIR are higher for all land points. For green vegetation, this is due to the sharp increase of leaf reflectance from approximately 0.1 up to 0.4 at the wavelength $\lambda = 0.7 \mu m$. This justifies the partition between visible and NIR radiation at $\lambda = 0.7 \mu m$. The albedo of bare soil does not exhibit a sharp increase at $\lambda = 700$ nm, but rises gradually from about 0.1 for $\lambda = 400$ nm via 0.2 for $\lambda = 700$ nm to 0.35 for $\lambda = 1600$ nm for a dry soil (Goudriaan, 1977). These characteristics lead to differences between the two spectral albedos of close to 0.2 for extended regions in South America, Africa and South East Asia (rain forests) as well as western Europe and main parts of North America. The differences are somewhat smaller over the boreal forests with their chiefly needleleaf trees (about 0.12). This is mainly due to the lower leaf reflectance in the NIR for needleleaf trees compared to deciduous trees. Sellers et al. (1996b) assume a value equal to 0.45 and 0.35 for the leaf reflectance of broadleaf and needleleaf trees, respectively. The broad band with low $\alpha_{NIR}$ between 0.15 and 0.20 (Fig. 5.22b) is primarily covered by boreal forests. The two spectral albedos of bright deserts differ by approximately 0.08. The VIS-albedo over the Sahara is typically about 0.38 while in the NIR, only approximately 30% of the incoming radiation is reflected.

The differences between the global albedo distributions using Methods 1 and 2, as described above, are generally small (not shown) and rather patchy. Over desert areas, $\alpha_{VIS}$ is considerably higher when Method 2 is applied while in the remaining areas, the visible surface albedo is approximately 0.03 less when using Method 2. The difference pattern for $\alpha_{NIR}$ is more patchy. Over extended areas of Russia, albedos derived with SiB2 (Method 2) are lower than in Method 1, whereas the opposite applies for western Europe, the main part of South America (excluding the desert) as well as the zone close to the equator in Africa.

**Annual cycles**

In Fig. 5.23, annual cycles of $\alpha_{VIS}$ and $\alpha_{NIR}$ for nine selected regions are displayed. All graphs are based on results using the SiB2 canopy radiation transfer model (Method 2). The regions are the same as in Fig. 3.12.

The monthly albedos generally show a distinct annual cycle. For example, over Eurasia (region R1 in Fig. 5.23), the annual amplitude amounts to close to 0.04, which greatly influences the radiation budget. The summer minimum of $\alpha_{VIS}$ is consistent with the results in Sellers (1985) who predicted, using a two-stream approximation model to simulate the radiative transfer, a decreasing $\alpha_{VIS}$ with increasing foliage. This is largely due to the scattering properties of the vegetation ensemble (radiation trapping). An additional
Figure 5.22: Global distribution of visible and near-infrared albedo for snow-free conditions (annual means) as derived from the radiation model used in SiB2 (Simple Biosphere Model). The water albedo is not calculated in SiB2. The albedo over ice-covered area is specified assuming ice-free conditions.

effect may arise due to the increasing albedo for decreasing solar angles. Sun rays from near-horizon are more likely to be reflected than is radiation from a high standing sun which is subject to increased radiation trapping. This feature may be even more important than the annual cycle of the LAI as the spectral albedo over the boreal forests (Taiga, region R4) depicts a strong annual cycle with an amplitude of approximately 0.05 despite a nearly constant LAI of the evergreen needleleaf trees. However, other studies (e.g., Verseghy et al., 1993) suggest that the zenith angle dependence of forest albedos is minimal.

The evolution of $\alpha_{NIR}$ during the year is somewhat more fuzzy. A lag of about three months for most vegetated areas excluding the Taiga can be revealed (Fig. 5.23). All these regions show the maximum in late autumn (October/November), with the minimum occurring in late winter/early spring (February to April). A reason for this evolution may
be the distinction between the dead and live leaf reflectances in SiB2. The very low leaf transmittance of dead material in the NIR (0.001, compared to 0.25 in the VIS-spectrum, see Table 5 in Sellers et al., 1996b) does not allow the radiation to penetrate the canopy which, again, reduces radiation trapping. Hence, higher albedos in the near-infrared are expected during periods with substantial dead material, applicable for late autumn.

### 5.4.2 Spectral albedos for snow and ice conditions

After the establishment and discussion of snow free values for $\alpha_{VIS}$ and $\alpha_{NIR}$, the allocation of spectral albedos for snow- and ice-covered conditions have to be specified.

In order to keep the relationship between albedo and temperature as close to the original ECHAM4 algorithm as possible, the linear relationship between snow albedo and temperature was maintained. In the following, the procedure to obtain the maximum and minimum broad-band albedos in the visible and NIR spectrum (Table 5.4) is outlined.
Table 5.4: Minimum and maximum surface albedo as used in the ECHAM4 control run and ranges for $\alpha_{VIS}$ and $\alpha_{NIR}$ which are derived from a literature study from various sources as outlined in the text.

<table>
<thead>
<tr>
<th>surface type</th>
<th>$\alpha_{min,VIS}$</th>
<th>$\alpha_{min,NIR}$</th>
<th>$\alpha_{max,VIS}$</th>
<th>$\alpha_{max,NIR}$</th>
<th>$\alpha_{min}$</th>
<th>$\alpha_{max}$</th>
<th>$\alpha_{min}$</th>
<th>$\alpha_{max}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sea ice</td>
<td>0.5</td>
<td>0.56</td>
<td>0.44</td>
<td>0.75</td>
<td>0.85</td>
<td>0.65</td>
<td>0.85</td>
<td>0.65</td>
</tr>
<tr>
<td>land ice</td>
<td>0.6</td>
<td>0.67</td>
<td>0.53</td>
<td>0.8</td>
<td>0.90</td>
<td>0.70</td>
<td>0.90</td>
<td>0.70</td>
</tr>
<tr>
<td>Snow on land, $a_f = 0$</td>
<td>0.4</td>
<td>0.57</td>
<td>0.39</td>
<td>0.8</td>
<td>0.95</td>
<td>0.65</td>
<td>0.95</td>
<td>0.65</td>
</tr>
<tr>
<td>Snow on land, $a_f = 1$</td>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
<td>0.4</td>
<td>0.4</td>
<td>0.4</td>
<td>0.4</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Sea-ice albedo

The sea ice albedos are computed according to Allison (1993). He suggests, based on a comprehensive field program, the following relationship between the all-wave albedo $\alpha$ and $\alpha_{VIS}$ and $\alpha_{NIR}$, respectively:

$$\alpha_{VIS} = \alpha + \delta$$

$$\alpha_{NIR} = \alpha - \delta$$

(5.19)

with $\delta = -0.01 + 0.14 \alpha$. Note that in literature, other values are also proposed: Briegleb and Ramanathan (1982) report spectral albedos given in Table 5.5, thereby stressing the importance of melt ponds. Assuming identical radiation fluxes in the two spectral bands with wavelengths of 0.2 $\mu$m - 0.5 $\mu$m and 0.5 $\mu$m - 0.7 $\mu$m, results in substantially lower averages in $\alpha_{VIS}$ and $\alpha_{NIR}$ compared to the values in Allison (1993). Similar low values are proposed by Hansen et al. (1983). They assume sea ice albedos of 0.3 and 0.55 in the NIR and visible waveband, respectively. The study of Allison (1993), however, is based on more recent results acquired from a very sophisticated field campaign and is, therefore, considered to be more accurate. Note that for sea ice with mature melt ponds, the ice albedos given in Table 5.4 may yield too high reflectivity.

Table 5.5: Spectral albedo of sea ice

<table>
<thead>
<tr>
<th></th>
<th>0.2 - 0.5 $\mu$m</th>
<th>0.5 - 0.7 $\mu$m</th>
<th>0.7 - 4.0 $\mu$m</th>
</tr>
</thead>
<tbody>
<tr>
<td>no melt ponds</td>
<td>0.75</td>
<td>0.70</td>
<td>0.50</td>
</tr>
<tr>
<td>major melt ponds</td>
<td>0.62</td>
<td>0.45</td>
<td>0.08</td>
</tr>
</tbody>
</table>

Land ice albedo

Land ice is generally assigned substantially higher reflectances than sea ice in both the VIS and NIR radiation band. According to BATS (Dickinson et al., 1993), the NIR-albedo is equal to 0.6 while in the VIS radiation band, the corresponding value increases up to 0.8. These values do not vary with temperature. It should, however, be emphasized that the albedos applied in BATS are related to the land-cover type of 'ice cap/glaciers', with these regions being mostly snow covered. It is thus likely, that the ice albedo in BATS refers rather to snow covered ice than bare ice, which also applies to ECHAM. Hansen et al. (1983) assumes considerably lower values than BATS, namely $\alpha_{NIR} = 0.35$ and $\alpha_{VIS} = 0.6$.

Due to a lack of any comprehensive study for spectral land ice albedos, Eq. 5.19 was also applied to land ice. The all-wave albedo for land ice were taken from ECHAM. However, it would be more appropriate to compute the snow cover fraction over ice-covered grid elements in order to properly distinguish between snow and bare ice. This would allow for the introduction of a separate relationship between albedo and temperature for both the ice and snow cover fraction within each grid box. In ECHAM, on the other hand,
neither the snow water equivalent nor the snow melt rates over both land- and sea ice are calculated, which precludes a separate treatment of ice and snow with respect to albedo.

**Snow albedo**

Several land surface models incorporate the spectral snow albedo. In the Canadian land surface scheme for GCMs (CLASS, Verseghy et al., 1993), visible snow albedos are 0.95, 0.84 and 0.61 and near-infrared albedos are 0.72, 0.56 and 0.38 for fresh snow, old dry snow and old melting snow, respectively. The albedo in the NIR for old snow, suggested by Briegleb and Ramanathan (1982) and listed in Table 5.6, is slightly lower than the albedo of old melting snow in CLASS. The spectral albedos for fresh snow in CLASS and Briegleb and Ramanathan (1982) agree well.

The snow albedo in BATS is based on the snow model of Anderson (1976). For new snow, he proposes $\alpha_{\text{VIS}} = 0.95$ and $\alpha_{\text{NIR}} = 0.65$, which is the same as reported in Briegleb and Ramanathan (1982). Sellers et al. (1996a) report that the snow albedo is reduced by around 40% as the snow melts. Assuming this albedo characteristic and the fresh snow albedo suggested by Briegleb and Ramanathan (1982), leads to $\alpha_{\text{VIS}} = 0.57$ and $\alpha_{\text{NIR}} = 0.39$ for melting snow. These values are adopted in EXP4. $\alpha_{\text{NIR}}$ is in good agreement with the value proposed in CLASS for old melting snow. The visible albedo seems to be slightly too low when compared with other studies. Robinson and Kukla (1984) report a high correlation between $\alpha_{\text{VIS}}$ and $\alpha_{\text{NIR}}$ for snow. They computed a correlation coefficient as high as 0.96 for their data. Their findings and the assumption that $\alpha_{\text{VIS}} = 0.95$ (fresh snow) yield to $\alpha_{\text{NIR}} = 0.72$, which corresponds well with CLASS but is somewhat higher than the value suggested in BATS.

Summarized, the comparison of snow albedos derived from different studies agree reasonably well. Discrepancies occur not only as a result of other observation conditions and different measurements technics, but also due to slight differences in the definition of the visible and NIR spectrum.

The establishment of the spectral albedos for snow covered forests necessitated a further task. The following procedure was applied. In order to compute the albedo of snow covered forests, the forest is assumed to consist of a snow free and a snow covered part. Assuming an albedo of snow free forests equal to 0.14 yields, for completely snow covered forests, a hypothetical snow cover fraction of 0.394 in order to get a snow covered forest albedo of 0.4 which is used in ECHAM for temperatures below -10°C. This leads, with $\alpha_{\text{NIR}} = 0.05$ and $\alpha_{\text{NIR}} = 0.23$ for snow free forests (cf. Table 5.4), to $\alpha_{\text{NIR}} = 0.395$ and $\alpha_{\text{VIS}} = 0.406$. This indicates that, in a good approximation, the albedo of snow covered forest in the visible and near-infrared range hardly differ. Therefore, $\alpha_{\text{VIS}} = \alpha_{\text{NIR}} = 0.4$ was applied in EXP4 for snow covered forests at temperatures below -10°C (cf. Table 5.4).

**Open ocean albedos**

A review of literature implies, that for water surfaces, the spectral albedo in the VIS and NIR virtually do not differ (e.g., Kondratyev, 1969, Briegleb and Ramanathan, 1982).
However, many studies have shown the strong relationship between the ocean albedo and the solar zenith angle (e.g., Kondratyev, 1969; Cogley, 1979; Henderson-Sellers and Wilson, 1983; Laffleur et al., 1997 and others). Therefore, EXP4 was run with an ocean albedo dependent on the cosine of the zenith angle $\mu_0$. In order to accurately express the ocean albedo, the parameterization given in Briegleb and Ramanathan (1982) was adopted:

$$\alpha_{water} = \frac{0.05}{1.1\mu_0^{1.4} + 0.15}. \quad (5.20)$$

This relation is limited to clear sky conditions. The albedo for diffuse radiation has been set to $\alpha_{water} = 0.08$, independent of wavelength and solar angle (Kondratyev, 1969). The splitting of total incoming radiation into its direct and diffuse components is discussed in Section 5.5.

**Total surface albedo versus the ratio of the incoming shortwave radiation in the VIS- and NIR-band.**

The ratio between the incoming shortwave radiation in the visible and near-infrared range determines the total radiation weighted surface albedo:

$$\alpha_{surf} = \frac{\alpha_{VIS} \cdot SW_{VIS,inc} + \alpha_{NIR} \cdot SW_{NIR,inc}}{SW_{VIS,inc} + SW_{NIR,inc}}, \quad (5.21)$$
where \( SW \) denotes the shortwave radiation and the suffix 'inc' indicates that only the incoming fluxes are considered.

Figure 5.24 shows the relationship between the ratio \( \frac{SW_{VIS,inc}}{SW_{NIR,inc}} \) (hereinafter \( n_{VN} \)) and the total surface albedo for a number of surface types. Over snow and ice surfaces, the surface albedo increases with increasing \( n_{VN} \). The snow- and ice albedos as obtained with ECHAM4 and SiB2, are identical for \( n_{VN} = 1 \), the exception being the curves which display the snow albedo at a temperature of 0°C. Figure 5.24c shows the albedo for the Sahara desert and the rain forest. The albedo values for the rain forest are obtained by an average over the grid boxes situated in the Amazonian rain forest. It is evident that the dependence of the (radiation weighted) surface albedo on \( n_{VN} \) is more pronounced for rain forests than deserts. Thus, changes in the cloud amount yield more significant albedo deviations in rain forests than over deserts. This also applies to the Taiga or the Eurasian continent where substantial influences on surface albedo due to changing cloud conditions are expected. The curves in Figure 5.24d, representing the SiB2 model, evolve in parallel which is due to similar reflectivity properties in Eurasia and the Taiga (cf. Fig. 5.23, Region R1 and R4).

5.4.3 Results from the 3-D experiment

A compilation of the annual long-term differences EXP4 - CTRL is given in Table 5.7. As seen in most other experiments, the 10-year global mean of most variables generally change very slightly, being statistically not significant. The annual mean of land albedo decreases by approximately 0.007. Despite the lower albedo, the absolute net shortwave radiation is reduced. This is due to enhanced cloud water amount in the atmosphere which yields a lower global radiation. The higher cloud water amount leads to an enhanced downward longwave radiation which compensates for the negative bias in the shortwave range. This implies a nearly constant surface temperature. A slightly more humid global climate is established, with more precipitation (1.4%), higher relative soil moisture (1.3%) and more clouds (0.8%) (Table 5.7).

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP4</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover fraction</td>
<td>%</td>
<td>22.6</td>
<td>22.9</td>
<td>-0.03</td>
<td>-0.14%</td>
</tr>
<tr>
<td>Surface albedo</td>
<td></td>
<td>0.291</td>
<td>0.298</td>
<td>-0.007</td>
<td>-0.22%</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>Wm(^{-2})</td>
<td>130.3</td>
<td>131.4</td>
<td>-1.2</td>
<td>-0.9%</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm(^{-2})</td>
<td>174.5</td>
<td>176.3</td>
<td>-1.8</td>
<td>-1.0%</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>Wm(^{-2})</td>
<td>-44.2</td>
<td>-44.9</td>
<td>0.7</td>
<td>-1.5%</td>
</tr>
<tr>
<td>Net. LW, surface</td>
<td>Wm(^{-2})</td>
<td>-62.9</td>
<td>-63.5</td>
<td>0.6</td>
<td>-1.0%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>281.53</td>
<td>281.46</td>
<td>0.08</td>
<td>0.08%</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm(^{-2})</td>
<td>-40.5</td>
<td>-40.7</td>
<td>0.2</td>
<td>0.4%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm(^{-2})</td>
<td>-18.6</td>
<td>-18.9</td>
<td>0.3</td>
<td>1.6%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>60.9</td>
<td>60.1</td>
<td>0.8</td>
<td>1.3%</td>
</tr>
<tr>
<td>Total cloud cover</td>
<td>%</td>
<td>53.7</td>
<td>53.3</td>
<td>0.4</td>
<td>0.8%</td>
</tr>
<tr>
<td>Vert. int. cloud water</td>
<td>gm(^{-2})</td>
<td>58.8</td>
<td>58.0</td>
<td>0.8</td>
<td>1.3%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/d</td>
<td>2.09</td>
<td>2.06</td>
<td>0.3</td>
<td>1.4%</td>
</tr>
</tbody>
</table>

Ratio of the spectral incoming shortwave radiation in the VIS- and NIR-band

The 3-D simulation has shown that the quotient of the incoming SW in the two spectral bands varies significantly due to changes in the cloud amount, or, more precisely, in the total cloud water and water vapour within the atmosphere. Since water vapour and
cloud droplets absorb more efficiently in the NIR than in the VIS spectrum, the ratio $SW_{VIS}/SW_{NIR}$ increases with increasing cloud amount. The scattering and absorption properties of aerosols may also lead to variations in the ratio between the visible and NIR incoming radiation, but this subject has not been included in the present studies.

The investigation of $n_{VN}$ ($SW_{VIS,inc}/SW_{NIR,inc}$) as a function of a number of variables has elucidated some interesting properties (Fig. 5.25). Fig. 5.25a shows the strong relationship between $n_{VN}$ and the vertically integrated cloud water. Higher cloud water amounts favour the visible incoming radiation due to enhanced absorption in the NIR spectrum. A linear fit yields $\Delta (\text{total cloud water})/\Delta n_{VN} = 450 \ \text{gm}^{-2}$ and a correlation coefficient of 0.87. For clear sky cases (Fig. 5.25b), it is most reasonable to relate the vertically integrated water vapour to $n_{VN}$ since cloud droplets are missing for clear sky cases. The exponential-like relationship between $n_{VN}$ (clear sky) and the vertically integrated water vapour ($vicl$) in [kgm$^{-2}$] can be approached by the sum-of-least-squares fit:

$$vicl = 0.01 \cdot e^{0.2 \cdot n_{VN}} - 8. \quad (5.22)$$

From this, it can be inferred that the sensitivity of $n_{VN}$ to $vicl$ decreases with increasing vertically integrated water vapour due to a saturation effect. As $n_{VN}$ does not exceed 1 for clear sky cases, it can be concluded that the additional absorption in the NIR range
due to cloud droplets plays a major role since, including all cases (denoted as 'all sky'), \( n_{V_N} \) reaches values up to approximately 1.5.

The pointcloud in Fig. 5.25b shows two clusters. For values of \( n_{V_N} \) larger than approximately 0.87, a further cluster with distinctly higher atmospheric absorption in the NIR develops. A more detailed investigation shows that these values belong to grid boxes in the arctic regions with generally low solar angles. Low solar angles cause a high optical mass and imply longer paths for photons to reach the Earth’s surface, leading to an enhanced absorption in the NIR and thus a lower value of \( n_{V_N} \).

Fig. 5.25c represents the relationship between \( n_{V_N} \) and the extinction in the atmosphere, computed from the downward fluxes at the surface and at the top of atmosphere (TOA). The graphic reflects the expected relationship: The higher the extinction, the more cloud water is present which absorbs mainly in the NIR, leading to a higher ratio \( n_{V_N} \). As in Fig. 5.25b, the pointcloud is split up into two “fingers”. The lower cluster is again related to arctic climate while the upper cluster is associated with predominantly (sub-) tropical climate.

**Changes in surface albedo over Africa due to spectral modification of the incoming shortwave radiation**

Africa is ideal to show that significant changes in the (radiation weighted) surface albedo (Eq. 5.21) occur due to changes in the spectral components of the radiation fluxes. Fig. 5.26a displays the albedo deviation in July between EXP4 (with a realistic distribution of \( n_{V_N} \)) and the (radiation weighted) surface albedo as calculated with \( n_{V_N} = 1 \). The albedo anomalies are thus dependent on both \( n_{V_N} \) and the difference between \( \alpha_{VIS} \) and \( \alpha_{NIR} \). The map of the albedo anomalies stresses the importance to separately consider the shortwave fluxes in the VIS and NIR spectra. Fig. 5.26 illustrates that a decrease in surface albedo is associated with high cloudiness due to \( \alpha_{NIR} > \alpha_{VIS} \) and preferred absorption in the near-infrared. \( n_{V_N} = 1 \) indicates that the total surface albedo is calculated according to \( \alpha_{surf} = (\alpha_{VIS} + \alpha_{NIR})/2 \). The introduction of unequal radiation fluxes in the VIS and NIR range leads to a significant increase of the albedo’s gradient of approximately 0.05 between the cloudy areas (\( \sim 8^\circ \)N) and regions with low cloudiness (\( \sim 15^\circ \)S).

Fig. 5.26b shows the close relation between the cloudiness and \( n_{V_N} \). There are quite large regions where \( n_{V_N} \) is lower than 0.8 or larger than 1.2.

**Annual cycles over the Sahara and adjacent steppes**

This section points out that the value of the surface albedo over northern Africa may play a major role in the climate. Fig. 5.27 shows the annual cycles of a number of variables for the Sahara including the bush vegetation adjacent to the south to approximately 15°N. Most surface variables are significantly affected by the introduction of \( \alpha_{VIS} \) and \( \alpha_{NIR} \) and the response is mostly statistically significant on the 95% level using the t-statistics test. The surface albedo is noticeably reduced by about 0.04 - 0.05. The SRB surface albedo climatology indicates that the control simulation is in better agreement with the observation than EXP4. Note that this feature is not attributed to an inadequate partitioning into the VIS and NIR radiation flux but an unrealistic mapping of \( \alpha_{VIS} \) and \( \alpha_{NIR} \) in the SiB2-model. The underestimation of the desert’s albedo in SiB2 is supported by Barker and Davies (1989), with albedo estimates for the western Sahara ranging from 0.35 to 0.45.

The comparison of further observed and simulated variables also indicates that the ob-
Figure 5.26: a) Albedo anomalies over Africa due to deviation from \( n_{VN} = 1 \) (\( n_{VN} = \) incoming shortwave radiation in the VIS-band divided by the shortwave radiation in the NIR spectrum). b) Shaded: Cloudiness; contour lines: \( n_{VN} \). 10-year averages of July as calculated in EXP4.

served SRB albedo is sufficiently accurate: The relative soil moisture, which is considerably overestimated in the control simulation, increases in the current simulation to unrealistically high values of over 50% in August and September. This is related to a more humid climate, mainly in the summer season with more precipitation, cloudiness and cloud droplets, and a higher evapotranspiration. A detailed investigation reveals the following possible explanations for this drastic "climate change" over northern Africa:

1. A lower albedo yields more available net shortwave radiation and thus, higher surface temperatures: This simple conclusion is incorrect in regard to the summer season of the displayed region. The reduction of the global radiation due to enhanced cloudiness and higher heat consumption caused by increased evaporation, reduces the surface temperature. Therefore, the hypothesis that lower albedo leads to higher surface temperature and thus more convection and precipitation must be rejected.

2. The westerly component of the horizontal wind strongly increases at the west coast of Africa between 10°N and 30°N (not shown). Thus, the southwest (monsoon) winds which have developed up to ~ 20°N, transport more humidity from the Atlantic towards the coastline and the inlands of Africa. This would enhance the available water vapour and thus, the precipitation rate. This hypothesis is supported by a particularly strong increase
Figure 5.27: Annual cycles of nine surface variables. Averages over the Sahara desert including the Steppes adjacent to the south up to ~15°N. Obs: Observation; CTRL: Control run; EXP4: EXP4. The observations are based on (cf. also Section 2): NCEP (Kalnay et al., 1996) for soil moisture; Legates (Legates and Willmott, 1990a) for 2-m-temperature; SRB (Barkstrom et al., 1989) for surface albedo and GPCP (Rudolf et al., 1996) for precipitation.

Ocean albedo

EXP4 includes the zenith angle dependence of the ocean albedo according to Equation 5.20 which applies only to the direct part of the incoming shortwave radiation. This requires the computation of the diffuse and direct component of the global radiation which is detailed
in Section 5.5.3. The effect of the zenith angle on the mean ocean albedo is clearly visible (in Fig. 5.28), i.e. The lower the mean solar angle the higher the albedo. The albedo depends also on the cloudiness: The lower the cloudiness, the higher the ratio between direct and diffuse radiation and hence, the more important the zenith angle dependence. Thus, the gradient in the ocean albedo between the tropics and the high latitudes would be substantially larger when neglecting all cloudy cases.

Figure 5.28b illustrates that in the tropics, the ocean albedo in EXP4 and CTRL agree well while in the high latitudes the ocean albedo in the modified experiment is higher by 0.015. The negative albedo anomalies correspond to generally ice-covered ocean areas. These cold areas are characterized by very low integrated water vapour and cloud water and thus a ratio $n_{VX}$ of less than 1.0 which leads to a decreasing ice albedo (Figure 5.24b).
Table 5.8: Ice free ocean albedo as simulated in EXP4 and the deviation from CTRL. 10-year means. Last column: Change (Δ) in the reflected shortwave radiation, neglecting all other feedbacks, i.e. \( Δ = Δα \cdot \text{global radiation}. \)

<table>
<thead>
<tr>
<th>month</th>
<th>EXP4</th>
<th>CTRL</th>
<th>EXP4 - CTRL</th>
<th>Bias in SW (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>8.3%</td>
<td>7.0%</td>
<td>1.3%</td>
<td>2.1 Wm(^{-2})</td>
</tr>
<tr>
<td>Feb</td>
<td>8.0%</td>
<td>7.0%</td>
<td>1.0%</td>
<td>1.8 Wm(^{-2})</td>
</tr>
<tr>
<td>Mar</td>
<td>8.0%</td>
<td>7.0%</td>
<td>1.0%</td>
<td>1.8 Wm(^{-2})</td>
</tr>
<tr>
<td>Apr</td>
<td>8.2%</td>
<td>7.0%</td>
<td>1.2%</td>
<td>2.1 Wm(^{-2})</td>
</tr>
<tr>
<td>May</td>
<td>8.5%</td>
<td>7.0%</td>
<td>1.5%</td>
<td>2.4 Wm(^{-2})</td>
</tr>
<tr>
<td>Jun</td>
<td>8.6%</td>
<td>7.0%</td>
<td>1.6%</td>
<td>2.4 Wm(^{-2})</td>
</tr>
<tr>
<td>Jul</td>
<td>8.5%</td>
<td>7.0%</td>
<td>1.5%</td>
<td>2.4 Wm(^{-2})</td>
</tr>
<tr>
<td>Aug</td>
<td>8.3%</td>
<td>7.0%</td>
<td>1.3%</td>
<td>2.1 Wm(^{-2})</td>
</tr>
<tr>
<td>Sep</td>
<td>8.6%</td>
<td>7.0%</td>
<td>1.6%</td>
<td>1.9 Wm(^{-2})</td>
</tr>
<tr>
<td>Oct</td>
<td>8.0%</td>
<td>7.0%</td>
<td>1.0%</td>
<td>1.9 Wm(^{-2})</td>
</tr>
<tr>
<td>Nov</td>
<td>8.2%</td>
<td>7.0%</td>
<td>1.2%</td>
<td>2.1 Wm(^{-2})</td>
</tr>
<tr>
<td>Dec</td>
<td>8.6%</td>
<td>7.0%</td>
<td>1.6%</td>
<td>2.2 Wm(^{-2})</td>
</tr>
<tr>
<td>Jan-Dec</td>
<td>8.2%</td>
<td>7.0%</td>
<td>1.2%</td>
<td>2.1 Wm(^{-2})</td>
</tr>
</tbody>
</table>

The average albedo difference of all ice-free ocean grid points are compiled in Table 5.8. The mean annual albedo is approximately 0.012 higher in EXP4 than in the control simulation. A bimodal evolution can be detected: Two maximum differences of surface albedo are generated at the summer and winter solstices while minima occur during spring and autumn. This means that the mean solar zenith angle for all ice-free ocean grid boxes, weighted with the incoming shortwave radiation, is largest during the two solstices. The model simulation shows that the mean monthly differences oscillate between 0.01 and 0.016. This leads, neglecting all other feedbacks, to a substantial increase in the annual average of reflected radiation by 2.1 Wm\(^{-2}\). The bimodal characteristic of the mean monthly albedo difference was also confirmed by the satellite-based observation of surface albedo (SRB climatology), with an amplitude of 0.009 being somewhat larger in the observation than in EXP4. This justifies the implementation of an ocean albedo dependent on the solar zenith angle. Furthermore, the increasing water albedo with increasing zenith angle has been confirmed by numerous theoretical and observational studies (Kondratyev, 1969; Hummel and Reck, 1979; Cogley, 1979; Robock, 1980; Briegleb and Ramanathan, 1982).

Finally, the monthly mean albedos (all-sky) of open water as suggested in Cogley (1979) has been compared to the values as calculated in EXP4. The ocean albedo estimates of Cogley (1979) are mainly based on albedo measurements by Grishchenko (1959). A comparison is carried out for 10° latitude belts from 0°N - 10°N up to 60°N - 70°N (Table 5.9). The albedo increase with increasing latitude is distinctly lower in ECHAM4 than in Cogley (1979). While, from EXP4, slightly higher albedos are computed in the tropical regions, Cogley (1979) obtains higher albedos in the higher latitudes. However,
the albedos extracted from EXP4 and tabulated in Cogley (1979) agree relatively well. Additionally, Cogley (1979) stresses that “the empiricism of these normals deserves to be noted”.

Summarized, the analysis of the climatologies derived from EXP4 revealed a number of interesting findings. It has been shown that the separate treatment of the shortwave radiation fluxes in the visible and near-infrared yields substantial differences due to changing cloud conditions. The evaluations further show that it is important to implement an ocean albedo which is dependent on the solar zenith angle.

5.5 EXP5: Incorporation of the BATS parameterization for snow albedo

5.5.1 Overview of parameterizations incorporating a snow age factor

The majority of snow models do not incorporate an explicit relationship between temperature and snow albedo. It is more common to introduce a function for the aging of snow which may account for snow grain size changes, impurities in the snow and the presence of liquid water in the snow. The aging function generally uses the time elapsed after the last snowfall as a key variable for the description of the albedo decay. Thereby, the rate of the albedo decrease is often dependent on whether the snow is melting. In the following section, some approaches related to snow aging are discussed.

In the Météo-France climate model (Douville et al., 1995b), two temperature regimes above and below 0°C are distinguished in which the aging process due to metamorphism and pollution exhibits a weak exponential and a linear decrease with time, respectively (Eq. 5.23). To keep in line with various field measurements, the minimum and maximum snow albedo are set to $\alpha_{s,\text{min}} = 0.5$ and $\alpha_{s,\text{max}} = 0.85$, respectively. A snowfall refreshes the albedo back to 0.85 when it exceeds the threshold value of 10 mm:

$$
\alpha_s(t + \Delta t) = \alpha_s(t) - \tau_a \frac{\Delta t}{\tau_1} ; T_s < 273.15 K
$$

$$
\alpha_s(t + \Delta t) = (\alpha_s(t) - \alpha_{s,\text{min}}) \exp \left[ -\tau_f \frac{\Delta t}{\tau_1} \right] + \alpha_{s,\text{min}} ; T_s \geq 273.15 K,
$$

(5.23)

where $\tau_a = 0.008$, $\tau_f = 0.24$, $\tau_1 = 86400$ s and $\Delta t$ the time step in seconds.

A similar approach is applied in the Canadian land surface scheme for GCMs (CLASS, see Verseghy (1991)) but with an exponential decay for both freezing and non-freezing conditions (Eq. 5.24). For temperatures $T < 0°C$ the albedo is allowed to vary between 0.7 and 0.84 while $\alpha_{s,\text{min}} = 0.5$ is applied for melting snow. A snowfall refreshes the albedo back to 0.84. Equations 5.23 and 5.24 are identical for the non-freezing case.

$$
\alpha_s(t + \Delta t) = (\alpha_s(t) - 0.70) \exp \left[ -\tau_f \frac{\Delta t}{\tau_1} \right] + 0.70 ; T_s < 273.15 K
$$

$$
\alpha_s(t + \Delta t) = (\alpha_s(t) - \alpha_{s,\text{min}}) \exp \left[ -\tau_f \frac{\Delta t}{\tau_1} \right] + \alpha_{s,\text{min}} ; T_s \geq 273.15 K
$$

(5.24)

The Canadian Climate Centre (CCC) second-generation model (McFarlane et al., 1992) incorporates an age factor which accounts solely for aging processes due to increased grain size and contaminations (Eq. 5.25), but the temperature dependence is neglected:

$$
f_{\text{age}}(t + \Delta t) = (1 - R) \left[ f_{\text{age}}(t) + \frac{\Delta t}{\tau} \right],
$$

(5.25)

where $R$ is a weighting factor (0 if no snow during the previous time step and up to 1 if recent snowfall is heavy), $\Delta t = 20$ min (time step) and $\tau = 40$ days. The spectral albedos can then be determined:

$$
\alpha_{s,\text{VIS}} = 0.9 - 0.15 f_{\text{age}}
$$

$$
\alpha_{s,NIR} = 0.7 - 0.15 f_{\text{age}}.
$$

(5.26)
Loth et al. (1993) suggest a linear decrease in the clear sky albedo with time but the decrease rate is different for the premelt period, the melting period as well as the postmelt period (Table 5.10).

Table 5.10: Decay in clear sky albedo after Loth et al. (1993)

<table>
<thead>
<tr>
<th>decrease rate</th>
<th>validity interval</th>
</tr>
</thead>
<tbody>
<tr>
<td>-0.0061/day</td>
<td>in premelt periods</td>
</tr>
<tr>
<td>-0.071/day</td>
<td>in melting periods, for depth &lt; 0.25 m</td>
</tr>
<tr>
<td>-0.015/day</td>
<td>in melting periods, for depth &gt; 0.25 m</td>
</tr>
<tr>
<td>-0.196/day</td>
<td>in postmelt periods</td>
</tr>
</tbody>
</table>

5.5.2 Overview of modifications

The current experiment incorporates the snow albedo following the Biosphere Atmosphere Transfer Model (BATS, Dickinson et al., 1986). BATS incorporates a prognostic variable for snow aging which essentially determines the snow albedo in the VIS and NIR spectra. Diffuse and direct radiation is treated separately. The computation of the diffuse ratio is based on PROBE data (Stokes and Schwartz, 1994).

The evaluation and validation identified two deficiencies in the BATS formulation: The computation of snow density using the snow water equivalent as well as snow aging over the Antarctica can be significantly improved by changing the appropriate parameter.

The snow albedo model in BATS

The sophisticated snow albedo parameterization used in BATS (Dickinson et al., 1986) includes the temperature effect in a prognostic equation for the age of the snow ($\tau_{age}$) which accounts, in an empirical way, for both the grain growth due to vapour diffusion and for an additional effect of melt water.

The key parameter in the BATS formulation for snow albedo is the snow aging factor $f_{age}$. $f_{age}$ is defined as

$$f_{age} = \frac{\tau_s}{1 + \tau_s} \quad (5.27)$$

where $\tau_s$ is a nondimensional age of snow, defined as

$$\tau_s^{N+1} = (\tau_s^N + \Delta \tau_s)[1 - \max(0, \Delta S_n)/\Delta P_s], \quad (5.28)$$

where $N$ denotes the current time step, $\Delta S_n$ is the change of snow water equivalent (in mm) in one time step $\Delta t$, and $\Delta P_s = 10 \, kgm^{-2}$ is the amount of fresh snow which is required to refresh snow albedo. This means that a snowfall of 10 mm water equivalent, or more, is assumed to restore the surface age which increases the snow albedo to its maximum value.

$\Delta \tau_s$ is parameterized as

$$\Delta \tau_s = (r_1 + r_2 + r_3) \frac{\Delta t}{\tau_0} \quad (5.29)$$

where $\tau_0 = 10^6 \, s$. $r_1$ represents the effects of grain growth due to vapour diffusion and is expressed as

$$r_1 = exp \left[ 5000 \left( \frac{1}{273.16} - \frac{1}{T_s} \right) \right], \quad (5.30)$$
where $T_s$ is the surface temperature; $r_2$ represents the additional effects of grain growth near or at the freezing of meltwater,

$$r_2 = (r_1)^{10} \leq 1;$$  \tag{5.31}

and $r_3$ represents the effect of dirt and soot,

$$r_3 = \begin{cases} 0.01 & \text{over Antarctic} \\ 0.3 & \text{elsewhere.} \end{cases}$$  \tag{5.32}

The parameterization of snow albedo is based on Wiscombe and Warren (1980):

$$\begin{align*}
\alpha_{VIS} &= \alpha_{VIS,D} + 0.4f(\psi)[1 - \alpha_{VIS,D}], \\
\alpha_{NIR} &= \alpha_{NIR,D} + 0.4f(\psi)[1 - \alpha_{NIR,D}],
\end{align*}$$  \tag{5.33}

where $\psi$ is the solar zenith angle, $\alpha_{VIS}$ the albedo for $\lambda < 0.7 \, \mu m$ and $\alpha_{NIR}$ the albedo for $\lambda \geq 0.7 \, \mu m$. The subscript $D$ denotes diffuse albedos as given by

$$\begin{align*}
\alpha_{VIS,D} &= [1 - C_S f_{age}]\alpha_{VIS,0}, \\
\alpha_{NIR,D} &= [1 - C_N f_{age}]\alpha_{NIR,0},
\end{align*}$$  \tag{5.34}

where $C_S = 0.2$ and $C_N = 0.5$. The albedos for visible and near-infrared solar radiation incident on new snow with a solar zenith angle less than 60° are $\alpha_{VIS,0} = 0.95$ and $\alpha_{NIR,0} = 0.65$. The function $f_{age}$ is defined in Equation 5.27. $f(\psi)$ is a factor between 0.0 and 1.0, giving the increase of snow visible albedo when the solar zenith angle exceeds 60°:

$$f(\psi) = \frac{1}{b} \left[ \frac{1 + b}{1 + 2b\cos(\psi)} - 1 \right],$$  \tag{5.35}

where $b = 2$. If $\cos(\psi) > 0.5$ then $f(\psi) = 0$.

The BATS scheme requires both the diffuse and direct shortwave radiation in the VIS and NIR-spectrum. Unfortunately, ECHAM4 does not provide for the fluxes to be separated into their diffuse and direct components. Hence, an empirical relationship must be developed (see Section 5.5.3).

The main advantage of BATS is, that all physically important processes are included in the parameterization. Furthermore, direct and diffuse as well as visible and near-infrared albedos are determined separately (cf. the following section).

**Comparison of snow aging parameterizations**

To uncover some differences in the snow albedo algorithms, the evolution of the snow albedo after a heavy snowfall for temperatures below and above the freezing point is illustrated for some models (Fig. 5.29). As seen in Fig. 5.29, the temporal decrease of $\alpha_s$ after heavy snowfall shows significant differences between the selected models. For temperatures below 0°C, CLASS provides the only example with an exponential decrease, leading to a rapid decrease in snow albedo after the last snowfall while the other expressions lead to a linear albedo decay. Eq. 5.23 leads to a slightly faster decrease than the other two linear models.

For melting snow, CLASS and the expression used in the Météo-France climate model are identical. The initial albedo differs by 0.01. Since snow aging in the CCC model is independent of temperature, the decrease in the snow albedo is small. Note that the decay rates, as suggested by Loth et al. (1993), correspond to the case with $S_0 > 25 \, \text{cm}$. The assumption of a thinner snow deck would lead to a significantly faster snow albedo decay.
Measurements by Baker et al. (1990), Robinson and Kukla (1984) and others, strongly suggest that the expression used in the CCC second generation model is untrustworthy. Comparing the ratio in the snow albedo decay for the melting to the non-melting case after heavy snowfall, yields the ratios displayed in Fig. 5.29c. The albedo decrease in the Météo-France climate model is, for \( T_s < 0^\circ \text{C} \), approximately 10 times as efficient as for above-freezing temperatures (immediately after snowfall has ended), while the ratio of the decay rate between melting to non-melting conditions is much lower in the other above-discussed approaches. When assuming \( S_n < 25 \text{ cm} \), however, the ratio between the decay rates following Loth et al. (1993) is in good agreement with the corresponding ratio as computed with the Météo-France climate model.

Fig. 5.29d shows the strong relationship between the albedo decay and temperature as simulated with BATS. The temporal albedo reduction at the melting point is very similar to the decay in CLASS and in the Météo-France climate model (curves 1 and 2 in Fig. 5.29c) for melting snow. The rate of the albedo decay for \( T = -20^\circ \text{C} \) is close to the
non-melting decay rate (Fig. 5.29a) applied in the Météo-France climate model and by Loth et al. (1993).

Assessment of the snow albedo decay with observational data

Baker et al. (1990) demonstrate that the albedo decay rates for all months are generally described equally well by either linear functions or exponential functions. Their studies were based on daily albedos of snow measured between November and April, 1969 - 1987. They found a linear decay rate of about 0.8% per day (December through February), whereas during November, March and April linear rates were 2.4, 2.9, and 3.3% per day, respectively.

The albedo decrease for cold periods in BATS (0.05/week for $T = -20^\circ$C, see Fig. 5.29) is in good agreement with an albedo decay equal to 0.008/day reported in Baker et al. (1990) for December through February, 1969 - 1987. During the melt period in March and April, they found a decay rate of approximately 0.03, in good agreement with the value in BATS for $T = 0^\circ$C (0.15/week).

In order to further validate the process of the snow albedo decay due to snow-aging, the comprehensive dataset from a 6-year-period at six Russian stations (cf. Section 2.2) was used. Since temperatures during the Russian winter are very low (cf. Table 5.11), and melting periods are generally short at the Russian stations, only the non-melting-case could be investigated.

To achieve this, the albedo decay as a function of number of days after measurable snowfall (assumed to equal 0.3 cm/day) was plotted for each of the six stations (Fig. 5.30). The number of samples decreases with increasing length of the period following the last snowfall. Hence, there is less confidence for longer decay periods than for short ones. It is

![Figure 5.30: Snow albedo decay for six Russian stations, 1978 - 1983 (solid line). The decrease is displayed as a function of number of days after measurable snowfall (> 0.3 cm/day). Only cases with a total snow height above 1 cm and $\alpha > 0.4$ are considered. Dashed: linear regression curves (corresponding correlation coefficient, see Table 5.11).]
striking that for all six stations, an excellent linear relationship exists between the albedo decay and days (after a snowfall) (see Table 5.11). This is consistent with the findings of Baker et al. (1990) who found a linear decrease of approximately 0.08 per day for the non-melting period. This is slightly more than what is observed at the Russian sites (see Table 5.11). It is noteworthy that the decay rate significantly differs from station to station. The temperature effect cannot be of major importance since all stations are located in the cold sub-arctic climate in Russia. Hence, it can be assumed that pollution effects also play a significant role.

Table 5.11: Snow albedo decay for six Russian stations, 1978 - 1983. The first 14 days after snowfall are used to determine the tabulated values. In addition to the rates of albedo decay, the correlation coefficients and the mean winter temperatures (December-March) are given.

<table>
<thead>
<tr>
<th>station</th>
<th>albedo decay/day</th>
<th>linear correlation coefficient</th>
<th>mean temperature, Dec.-Apr.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yershov</td>
<td>-0.0033</td>
<td>-0.90</td>
<td>-16.0°C</td>
</tr>
<tr>
<td>Tulun</td>
<td>-0.0063</td>
<td>-0.96</td>
<td>-17.0°C</td>
</tr>
<tr>
<td>Uralsk</td>
<td>-0.0074</td>
<td>-0.95</td>
<td>-12.3°C</td>
</tr>
<tr>
<td>Kostroma</td>
<td>-0.00417</td>
<td>-0.96</td>
<td>-10.9°C</td>
</tr>
<tr>
<td>Khabarovsky</td>
<td>-0.0080</td>
<td>-0.93</td>
<td>-16.1°C</td>
</tr>
<tr>
<td>Ogurtsovo</td>
<td>-0.0067</td>
<td>-0.94</td>
<td>-11.4°C</td>
</tr>
</tbody>
</table>

Summarized, the results derived from the BATS snow model accurately represent the temporal snow albedo decay deduced from other models and field observations. Both findings from the literature and results obtained by the evaluation using albedo data measured at six Russian sites point out that a linear snow albedo decay for non-melting snow is beneficial to capture the rate of albedo changes after a snowfall. Since the BATS model includes all relevant processes to capture changes in snow albedo and, a number of observational studies support the use of an aging factor, the BATS algorithm was preferred for implementation in ECHAM4.

5.5.3 Ratio of diffuse/direct radiation

The total shortwave radiation can be separated into the direct and diffuse component. The solar radiation that passes directly through to the Earth’s surface is called direct solar radiation. The radiation that has been scattered out of the direct beam is called diffuse solar radiation. The diffuse ratio \( r_d \) is defined as the diffuse shortwave radiation divided by the total shortwave radiation. Under clear skies, \( r_d \) is typically only about 10 - 15% near local solar noon, increasing to 100% before sunset. For a cloudy sky, observations indicate that \( r_d \) is close to 1. Though factors such as surface albedo, aerosol distribution and atmospheric water vapor have an impact on \( r_d \), its magnitude for clear skies is primarily determined by the solar angle.

Fig. 5.31 shows the relationship between the solar angle and the diffuse ratio derived from three studies. The dashed line is a fit to the diffuse ratios given in Goudriaan (1977), his Table 1. These ratios are based on Wit (1965) and are valid for a very clear sky:

\[
 r_d = \frac{5}{\beta} + \frac{\exp\left(\frac{\beta}{50}\right)}{100} 
\]

where \( \beta \) is the solar angle in degrees.

The solid line is the sum-of-least-squares fit for the high-frequent measurements (every 5 minutes) of the Pilot Radiation Observation Experiment (PROBE). PROBE was conducted at Kavieng, Papua New Guinea, from November, 1992 through February, 1993.
The principle objective of PROBE was to carry out detailed measurements of the surface radiation budget and the influence of clouds on this budget (for a discussion of the measurement program, see Stokes and Schwartz, 1994). The best-fit equation for the diffuse ratio based on PROBE is

\[ r_d = 0.1220.85e^{-4.8\mu_0} \]  

(5.37)

where \( \mu_0 \) is the cosine of the zenith angle. The two relationships discussed above agree well for solar angles above 10°. PROBE gives slightly higher diffuse ratios than Goudriaan (1977). This is likely related to the fact, that the observations for the ratios presented in Goudriaan (1977), are based on very clear sky conditions. Further, as global radiation is small at large zenith angles, the differences in the two above-mentioned studies play a minor role.

The clear sky diffuse ratios measured during flight missions between 10 February and 24 March 1983 (cf. Robinson and Kukla, 1984) are distinctly higher (by approximately 0.1). The reasons remain unknown. However, it may be that discrepancies occur due to the different measurement methods (aircraft measurements and ground based measurements).

Since the dataset in PROBE provided a large number of samples and the fitted curve approaches the diffuse ratios tabulated in Goudriaan (1977), Eq. 5.37 was implemented in the simulation runs EXP4 and EXP5 to split up the shortwave incoming flux into its direct and diffuse part. This relationship was applied only to the cloud free fraction in the grid boxes whereas in the cloudy part, the incoming radiation was assumed to be diffuse.

### 5.5.4 Results from the 3-D experiment

Based on the simulation EXP5, some features attributed to the snow albedo implemented in BATS, are investigated. A principle question arises whether the BATS algorithms can be implemented in ECHAM without changes, or, if certain parameters have to be readjusted. Since BATS incorporates spectral albedos in the visible and near-infrared band, EXP4 has been used as the control experiment throughout this chapter to preclude changes due to the modification in the surface albedo fields. Thus, the modified ECHAM4
source code used in EXP4 was selected as a basis to introduce modifications required to integrate the BATS snow albedo formulation.

**Long-term annual means**

The incorporation of the BATS snow albedo into ECHAM4 leads to small changes in the long-term annual means. However, regional and continental changes become statistically highly significant. To analyze the large-scale effect of the modification, the deviations from the control experiment (remember that the control simulation refers to EXP4 throughout this chapter), averaged over the Northern Hemisphere with $S_n > 0.1$ cm in February, are compiled in Table 5.12. The positive bias in the surface albedo yields a distinct decrease in the annual surface temperature which favours snowfall over rain. Despite slightly lower total precipitation rates, which may be caused by cooler temperatures and less turbulence (which is confirmed by decreasing turbulent heat fluxes), the snowfall rate increases by ~2%. Since changes in the annual global radiation and precipitable water are minimal, the enhanced surface albedo results in a negative SW radiation bias of more than 1 Wm$^{-2}$. The change in hydrological surface variables indicates that the hydrological cycle may be slightly attenuated.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP5</th>
<th>EXP4</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover</td>
<td>%</td>
<td>34.7</td>
<td>33.1</td>
<td>1.3</td>
<td>3.9%</td>
</tr>
<tr>
<td>Snow water eq.</td>
<td>cm</td>
<td>3.35</td>
<td>3.20</td>
<td>0.15</td>
<td>4.7%</td>
</tr>
<tr>
<td>Surface albedo</td>
<td></td>
<td>0.31</td>
<td>0.30</td>
<td>0.01</td>
<td>3.2%</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>Wm$^{-2}$</td>
<td>94.7</td>
<td>95.9</td>
<td>-1.2</td>
<td>-1.2%</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm$^{-2}$</td>
<td>126.7</td>
<td>126.0</td>
<td>0.7</td>
<td>0.6%</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>Wm$^{-2}$</td>
<td>-32.0</td>
<td>-30.1</td>
<td>-1.9</td>
<td>6.4%</td>
</tr>
<tr>
<td>Net LW, surface</td>
<td>Wm$^{-2}$</td>
<td>-50.7</td>
<td>-51.1</td>
<td>0.3</td>
<td>-0.7%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>272.75</td>
<td>273.09</td>
<td>-0.34</td>
<td>-0.1%</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm$^{-2}$</td>
<td>-31.5</td>
<td>-32.0</td>
<td>0.5</td>
<td>-1.5%</td>
</tr>
<tr>
<td>Evapotranspiration</td>
<td>mm/day</td>
<td>-1.06</td>
<td>-1.07</td>
<td>0.01</td>
<td>-1.5%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm$^{-2}$</td>
<td>-1.65</td>
<td>-1.78</td>
<td>0.13</td>
<td>-7.3%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>74.1</td>
<td>73.9</td>
<td>0.2</td>
<td>0.3%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>1.69</td>
<td>1.70</td>
<td>-0.01</td>
<td>-0.5%</td>
</tr>
<tr>
<td>Snowfall</td>
<td>mm/day</td>
<td>0.67</td>
<td>0.65</td>
<td>0.02</td>
<td>3.1%</td>
</tr>
<tr>
<td>Total cloud cover</td>
<td>%</td>
<td>63.4</td>
<td>63.1</td>
<td>0.3</td>
<td>0.4%</td>
</tr>
</tbody>
</table>

**Albedo difference (EXP5 - EXP4) in winter**

Fig. 5.32a demonstrates that the surface albedo increases in the relatively warm areas (Europe, northern parts of USA, as well as the west coast of North America) while in large parts of Russia and Canada, the surface albedo in the modified run is lower than in EXP4. Note that, in warmer regions, the precipitation rates are generally higher than in the cold continental and subarctic regions. The question arises whether the key factor for a negative albedo bias is the distribution of precipitation, temperature or another parameter as, e.g. the diffuse ratio or the ratio between the incoming shortwave radiation in the VIS and NIR. The latter hypothesis can be rejected since both $\alpha_{VIS}$ and $\alpha_{NIR}$ (Fig. 5.32b, c) show similar anomalies in both magnitude and regional distribution. More frequent and heavier snowfalls lead to a reduced temporal decrease in the snow albedo (Eq. 5.28) and lower temperatures decelerate the snow aging (Eq. 5.28 and Fig. 5.29d). Fig. 5.29d illustrates
that snow aging is largely dependent on temperature. Linear regressions between the albedo difference and the corresponding changes in other surface variables suggest that surface temperature is more crucial for the snow aging process. This is primarily applicable to months in spring with rapidly increasing temperatures. Positive albedo anomalies can also be related to positive deviations in the snow water equivalent. Relating differences in surface albedo to differences in snow water equivalent (not shown), however, demonstrates a fairly low correlation.

\[ \text{Figure 5.32: Differences in winter (DJF) surface albedo: EXP5 - EXP4. a) Total surface albedo, b) spectral albedo for the VIS-band (\( \alpha_{VIS} \)), c) spectral albedo for the NIR-band (\( \alpha_{NIR} \)). Grid boxes with a mean monthly global radiation less than 1 Wm}^{-2} \text{ in February have been omitted (polar night).} \]
Long term annual cycles averaged over the Northern Hemisphere

We first focus on the differences between EXP5 and EXP4 (represented by thick lines in Fig. 5.33), EXP4 being used as control simulation in this chapter. Most differences show higher fluctuations in North America than in Eurasia. Largest differences in the radiation budget occur, in both Eurasia and North America, in the month of May. In North America, the negative anomaly of net radiation is considerably higher than 10 Wm\(^{-2}\) in May, primarily due to an increase in surface albedo, resulting in a loss of net shortwave radiation. The higher snow albedo and lower surface temperature in May imply a delayed snow melt, which is clearly demonstrated in Fig. 5.33. The reduced available radiation leads to a decrease (in magnitude, note the sign convention) in the turbulent heat fluxes which amounts to approximately double the value in North America than in Eurasia. It is surprising that reduced evapotranspiration leads to more cloud cover (Fig. 5.33). This is likely to be caused by advective processes, thereby inducing some moisture convergence.

![Graphs showing monthly means for various climate variables](image)

**Figure 5.33:** Differences of annual cycles based on 10-year monthly means EU: Eurasia, restricted to areas with a monthly snow water equivalent in February larger than 0.1 cm; NA: As EU, but for North America. The observation (Obs) are based on: NCEP (Kalnay et al., 1996) for soil moisture; Legates (Legates and Willmott, 1990a) for 2-m-temperature; SRB (Barkstrom et al., 1989) for surface albedo, GPCP (Rudolf et al., 1996) for precipitation and NOAA (1973 - 1996) for snow cover fraction. Sign convention: downward fluxes are counted positively.
In January, the anomalies between the two continents differ significantly. In Eurasia, the impact of the changed LAI distribution on the surface albedo is negative while the corresponding change over North America is substantially smaller. The negative albedo anomaly over Eurasia yields enhanced net shortwave radiation and a distinct warming of the surface by approximately 1 °C in February and March. The change in the surface temperature and the net radiation over Eurasia are highly correlated, whereas the correlation over North America is fairly low. This is likely due to the different sizes of Eurasia and North America, leading to a higher percentage of Eurasia with predominantly continental climate.

Fig. 5.33 displays for a number of variables the difference between EXP5 and the observations (thin lines). The maximum differences between EXP5 and EXP4, and EXP5 and the observation, respectively, are similar for surface albedo. Regarding snow cover fraction, surface temperature and soil moisture, however, the differences between EXP5 and the observations are significantly larger than the differences between EXP5 and EXP4. Soil moisture is strongly underestimated during winter and spring. In North America, relative soil moisture is approximately 2% higher in EXP5 than in EXP4 throughout the year while in Eurasia the corresponding change is minimal.

The large deviations from the observed surface temperature are mainly related to a very poor interpolation method used in Legates and Willmott (1990a) over mountainous regions with a generally sparse station network, and are thus not relevant. Fig. 5.33 shows that the delayed snow melt in ECHAM4, and consequently its excessive snow amount in spring, is more pronounced in the experiment with the BATS snow albedo than in the control simulation. This leads to an enhanced cooling in late spring in both Eurasia and North America.

**Snow density vs. snow water equivalent**

Snow density $\rho_s$ plays a major role in the thermal diffusivity $k_{sn}$ of snow, which greatly influences the speed of a temperature signal into the snow pack. $\rho_s$ is also crucial for the water vapour flux within a snow pack and thus the growth of snow crystals. Since snow height is usually observed, the simulation of snow density also makes the validation easier. Furthermore, the branch of winter tourism is interested in the snow height rather than the snow water equivalent.

Snow models usually compute the snow water equivalent and, therefore, do not provide any information on the snow depth. The BATS snow model, however, allows the computation of the snow density:

$$\rho_s = \rho_{new} \cdot (1 + 3 f_{age}), \quad (5.38)$$

with

- $f_{age}$ snow aging factor (Eq. 5.27)
- $\rho_s$ snow density
- $\rho_{new} = 100 \text{ kgm}^{-3}$.

Fig. 5.34 illustrates that for small snow water equivalents of up to a few centimetres, the relationship between $\rho_s$ and $S_n$ derived from EXP5 is in line with the observations, while the density for thicker snow packs is substantially underestimated. The observations suggest a fast increase of up to approximately $\rho_s = 250 \text{ kgm}^{-3}$ and a subsequent slower increase of snow density with $S_n$. The underestimation in snow density is consistent with results presented in Yang et al. (1997). They suggest the following explanation: As the computation of the snow aging is intended to mimic the surface snow processes,
the density computed this way refers to the surface snow density. Thus, after heavy snowfall large errors may occur. Yang et al. (1997) have shown that a modification in ∆Pₜ (Eq. 5.28), the parameter which describes the rate of increase in snow albedo due to snowfall, leads to a significant improvement. They suggested an increase of ∆Pₜ from 10 kgm⁻² to 60 kgm⁻² (only for the computation of snow density and not of snow albedo). This has been tested in an additional three-dimensional model experiment which led to a good agreement between the model and the observation (Fig. 5.34).

Albedo over the Antarctica

The snow albedo algorithm in BATS computes the parameters r₁ (Eq. 5.30), r₂ (Eq. 5.31) and r₃ (Eq. 5.32) which describe the aging of snow. r₂ can be neglected in very cold regions such as Antarctica, but the values of r₁ and r₃ play a major role regarding the snow albedo. Fig. 5.35 illustrates that ECHAM4, including the BATS algorithm for snow, substantially underestimates the surface albedo in Antarctica (Roesch et al., 1999 and Table 5.13). Table 5.13 illustrates that the radiation fluxes and thus surface temperature change considerably when reducing the mean surface albedo by 0.072. The net shortwave radiation increases (absolutely) by more than 20 Wm⁻² which yields a strong warming of approximately 5°C. In Table 5.14, a comparison between the observed and simulated radiation fluxes is provided for the South Pole. The modeled fluxes are generated with the control run (CTRL). In ECHAM4/T42, the South Pole is represented by the 87.9°S
Figure 5.35: Simulated (EXP5) minus observed surface albedo of Antarctica during southern summer (DJF). Observation: SRB (1984 - 1990). Contour lines: Simulated monthly precipitation rates (mm) for the same period.

Table 5.13: Differences EXP5 - EXP4 in Antarctica, DJF. Sign convention: downward fluxes are counted positively. The observed surface albedo is based on SRB data (Surface Radiation Budget).

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP5</th>
<th>EXP4 (&quot;Control&quot;)</th>
<th>EXP5 - EXP4</th>
<th>Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface albedo</td>
<td>%</td>
<td>71.3</td>
<td>78.5</td>
<td>-7.2</td>
<td>82.2</td>
</tr>
<tr>
<td>Shortwave radiation (upward)</td>
<td>Wm^{-2}</td>
<td>-217.3</td>
<td>-246.7</td>
<td>29.4</td>
<td></td>
</tr>
<tr>
<td>Global radiation</td>
<td></td>
<td>304.4</td>
<td>313.2</td>
<td>-8.8</td>
<td></td>
</tr>
<tr>
<td>Net shortwave radiation</td>
<td>Wm^{-2}</td>
<td>87.1</td>
<td>66.5</td>
<td>20.6</td>
<td></td>
</tr>
<tr>
<td>Longwave radiation (upward)</td>
<td>Wm^{-2}</td>
<td>-236.6</td>
<td>-216.7</td>
<td>-19.9</td>
<td></td>
</tr>
<tr>
<td>Downwelling longwave radiation</td>
<td>Wm^{-2}</td>
<td>170.6</td>
<td>157.8</td>
<td>12.8</td>
<td></td>
</tr>
<tr>
<td>Net longwave radiation</td>
<td>Wm^{-2}</td>
<td>-66.1</td>
<td>-59.0</td>
<td>-7.1</td>
<td></td>
</tr>
<tr>
<td>2-m-temperature</td>
<td>°C</td>
<td>-19.5</td>
<td>-24.6</td>
<td>5.1</td>
<td></td>
</tr>
</tbody>
</table>

latitude belt at a mean height of 2691 m a.s.l., which is approximately 140 m less than the altitude of the South Pole station. Both the upwelling longwave flux and thus surface temperature agree well with the observation. Large deviation, in contrast, are found in the solar radiation fluxes. Global radiation is underestimated by almost 38 Wm^{-2}. This significant deviation is hardly to explain as the atmosphere is very dry at South pole and the integrated water vapour negligibly small. The net shortwave flux agree, due to an error cancellation, excellently well. Assuming an exchange of the observed and simulated surface albedo would yield a difference in the net shortwave radiation of approximately 22 Wm^{-2}. Calculations with the isolated BATS algorithm show that for temperatures typically measured in the higher elevated areas of Antarctica in DJF (approximately -20°C) the snow aging due to dirt and soot (represented by \( r_3 \)) is negligible, compared to snow aging caused by grain growth due to vapour diffusion (represented by \( r_1 \)) (e.g., \( r_3/r_1 = 0.04 \) for \( T = -20°C \)). Thus, the too low surface albedos in cold regions with low precipitation rates, indicate that the effect of temperature on snow aging is significantly overestimated. The frequency and amount of snowfall is not a key parameter since, in contrast to \( r_1 \), the accurate value of \( r_3 \) does not play an important role in Antarctica. In addition, the computed total annual precipitation seems to be realistic (Ohmura et al., 1996b). A simulation with the full three-dimensional ECHAM4 as used in EXP5, but with
### Table 5.14: Simulated and observed radiation budget at the South pole, DJF. Sign convention: downward fluxes are counted positively. Observation are climatological means from the Amundsen-Scott South Pole Station, tabulated in Kusunoki (1985).

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>CTRL observation</th>
<th>CTRL - Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface albedo %</td>
<td>%</td>
<td>79.6</td>
<td>82.2</td>
</tr>
<tr>
<td>Shortwave radiation (upward)</td>
<td>Wm$^{-2}$</td>
<td>-265.6</td>
<td>-304.2</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm$^{-2}$</td>
<td>332.0</td>
<td>369.8</td>
</tr>
<tr>
<td>Net shortwave radiation (upward)</td>
<td>Wm$^{-2}$</td>
<td>66.4</td>
<td>65.8</td>
</tr>
<tr>
<td>Longwave radiation (upward)</td>
<td>Wm$^{-2}$</td>
<td>-188.7</td>
<td>-192.1</td>
</tr>
<tr>
<td>2-m-temperature °C</td>
<td>°C</td>
<td>-32.6</td>
<td>-31.9</td>
</tr>
</tbody>
</table>

$r_1 = r_1/2$ on the Antarctica ice sheet revealed a substantial improvement in the surface albedo over Antarctica.

**Summary**

To summarize, the snow albedo algorithm from BATS is an efficient tool to investigate the evolution of the snow albedo. The use of a snow-aging function has proved to be reasonable and the linear snow albedo decay with time, as found in observational data (for low temperatures), is well simulated with the BATS snow model. The computed surface albedo seems to be realistic for land points excluding Antarctica. In Antarctica, the strong underestimation of the surface albedo is induced by the factor $r_3$, which is too high in cold and dry regions. A faster decrease of $r_3$ with decreasing temperatures substantially improves the surface albedo distribution in Antarctica. The strongly underestimated snow density can be eliminated by substituting $\Delta P_s = 10 \text{ kgm}^{-2}$ with $\Delta P_s = 60 \text{ kgm}^{-2}$.

### 5.6 EXP6: Annual cycle of the LAI

EXP6 investigates the changes attributed to the implementation of the annual cycle of the leaf area index (LAI) in more detail. ECHAM4 provides for each grid element a constant LAI throughout the whole year.

The leaf area index is defined as the surface of all leaves projected on a horizontal plane. Many processes in the vegetation layer are strongly dependent on the amount of leaves available, e.g., the transpiration, the radiation transfer through the canopy or the intercepted water (rain or snow) on the canopy layer after a precipitation event. Most of the GCMs introduce the leaf area index as one of the key parameters for the inclusion of vegetation. Unfortunately, the LAI is not straightforward to measure, particularly on a global scale. Different assumptions can lead to significant differences in the global distribution of the LAI.

#### 5.6.1 Overview of modifications

The current ECHAM4 version prescribes a constant LAI-value for each grid box. In this experiment, the monthly LAIs from the International Satellite Land Surface Climatology Project (ISLSCP, Sellers et al., 1996) are used to improve the quality of the leaf area index. The global dataset is transformed from a regular 1° x 1° grid onto the T42-grid by using an area-weighing interpolation. A spline function is applied to derive daily values from monthly means. Computed LAIs are divided by the vegetation ratio to relate the LAIs to the vegetated area rather than to the entire grid box.
5.6.2 Compilation of global LAI datasets

In order to determine the monthly LAI on a global scale, two alternative methods can be applied. The first method is based on a remote sensing global dataset whereas the second procedure is based on the use of a global vegetation set and the allocation of typical minimum and maximum LAI-values to each vegetation type. Both methods are described and discussed in the following sections.

Method 1

ISLSCP (Sellers et al., 1996) provides comprehensive datasets which are necessary to describe boundary conditions and to initialize and force a wide range of land-atmosphere models. These Initiative I data sets span the 24-month-period of 1987 - 1988 at one-monthly time resolution and are mapped to a common 1° x 1° grid. The derivation of the global land cover data is based on the normalized difference vegetation index (NDVI, cf. Section 4.3.3) which is measured with an advanced very high resolution radiometer (AVHRR) (Los et al., 1994). Remote sensing takes advantage of the different reflection of green vegetation: less than 20% in the green-red, but approximately 60% in the NIR. For a detailed derivation of vegetation parameters from NDVI see Sellers et al. (1996a).

The derivation of the required LAI field from the ISLSCP data is as follows:

1. Interpolation of the monthly LAI from a common 1° x 1° grid onto the T42-grid. This was completed by calculating the percentage of the 1° x 1° grid elements within each T42-cell and multiplying this value by the corresponding LAI. Ocean grid cells were omitted. To obtain daily values, a spline-interpolation was applied. However, it would be sufficient to assume monthly constant LAIs.

2. The ECHAM-parameterizations which include the LAI only relate to the vegetated area, while the ISLSCP leaf area index represents the total area mean. Therefore, the LAI values are divided by the vegetation ratio which is assumed to be constant for each mesh during the year.

Method 2

The second procedure is to create the 12 month LAIs on a global scale using a global vegetation set. The Olson dataset (Claussen et al., 1994), was favoured since the forest and vegetation fractions in ECHAM4 are also deduced from Olson. The high-resolution 0.5° x 0.5° global dataset of Olson et al. (1983) is divided into 45 major ecosystem complexes. The allocation of these vegetation types or ecosystem complexes to the minimum (for the season of dormancy) and maximum (for the growing season) LAIs is given in Claussen et al. (1994) (Table 8).

The seasonal cycle of the LAI is incorporated using Eq. 5.39 of the Europa-modell (EM), described in detail in Edelmann et al. (1995). The annual cycle depends on the latitude \( \phi \), the day of the year \( JD \) and the height above sea level \( z_s \).

\[
LAI(\phi, JD, z_s) = LAI_{min} + f_v(\phi, JD)(LAI_{max} - LAI_{min})f_h(z_s) 
\]

(5.39)

with

\[
f_v(\phi, JD) = \text{Max}\{\min[1, C \sin(\pi max(0, JD - BVP)/DVP)]\}
\]

\[
f_h(z_s) = \exp\{-5 \cdot 10^{-9} \phi_s^2\}
\]

and
BVP(φ) : Start of the vegetation period
DVP(φ) : Duration of the vegetation period [days]
Ψ_s : Geopotential of the Earth’s surface [m^2 s^{-2}]
JD : day of the year
C = 1.12 (that is, during about 30% of the vegetation period, f_v = 1 is applied).

BVP(φ) and DVP(φ) are derived from monthly mean temperatures and are expressed as linear functions of latitude φ. Fig. 5.36 shows the difference between the two methods for

![Figure 5.36: LAI difference: ISLSCP minus LAI derived with Method 2 for summer (JJA) and winter (DJF).](image)

...summer (JJA) and winter (DJF), respectively. During the northern winter (Fig. 5.36a), the satellite-derived dataset tends to be lower than the LAI derived from the Olson data. Largest differences are found particularly in regions with Taiga vegetation, in the rain forests and the Rocky Mountains. The satellite “sees” substantially lower values for forests.
The ISLSCP dataset hardly shows any difference between the LAI of the Amazonian rain forest and the adjacent savanna in Brazil. The LAI compiled with Method 2, on the other hand, shows a very sharp gradient in the LAI at the border of the (closed) rain forest area. During the growing season (Fig. 5.36b), the LAI based on the ISLSCP data is significantly lower than the LAIs derived with Method 2. This is mainly in western Europe, South East Asia and extended regions in USA and Canada. Most tropical regions with minimal seasonal temperature variations, show a negligible amplitude in the annual cycle of the LAI. Further, the difference pattern between the two above-described methods shows only a minimal seasonal variation.

Both methods pose many problems. Unfortunately, it is difficult to decide which method leads to a more realistic distribution of the leaf area index.
In the current simulation, however, the satellite-derived observations are chosen as new boundary condition for the following reasons:

1. The averaging method is more simple and realistic.
2. No allocation of minimal and maximal LAIs to vegetation types is necessary.
3. Each (Olson) grid element is described by one single vegetation type.
4. The application of the EM-algorithm to compute the annual cycle of the LAI is avoidable.
5. Claussen et al. (1994) report that the allocation of leaf area indices to vegetation types is rather poorly correlated and that it seems more reasonable to infer LAI from data concerning net primary production of vegetation.

In order to correctly analyze the effect of the modified LAI pattern on the surface climate, it is necessary to know the differences between the LAI distribution in the ECHAM4 control simulation and the ISLSCP data. Fig. 5.37 displays the differences of the satellite-derived LAI and the global boundary conditions incorporated in the ECHAM4 control simulation. During the Northern Hemisphere winter (DJF) ISLSCP is lower for the whole globe, excluding some regions in South America, e.g. the savannas in Brazil. Substantial differences occur in the northern regions of Eurasia and major parts of West Europe and North America/Canada. This may have a significant impact on the surface albedo of snow covered forests when the parameterization of snow masking includes the leaf area index. Absolute differences in the LAI tend to be higher in more forested areas, whereas savannas and desert areas have smaller (absolute) differences. Claussen et al. (1994) have allocated the rain forests a LAI of 9.3, whereas the satellite-derived LAI is approximately 6. This leads to the dark spots over the Amazonian rain forest and islands of South East Asia throughout the year. The mean summer LAI (JJA) in ECHAM4 is lower than in ISLSCP, primarily in the northern parts of Eurasia and Canada with large forested areas. An exception seem to be the Rocky Mountains with its large evergreen forests where the satellite “sees” lower values throughout the year. The LAI over the Indian subcontinent, with predominantly agricultural lands, is substantially lower in ECHAM4 throughout the year. This may be either due to difficulties in allocating each grid cell to one single vegetation type or to rapid changes in the vegetation pattern.

Fig. 5.38 shows the annual cycle of the LAI for the whole Earth (land points without ice cover) and three continents. The seasonal cycle of the LAI is significant over North America and Eurasia, while in Africa, the annual amplitude is minimal. Only two very weak maxima are seen in the evolution of the LAI in Africa. Note that the satellite-derived LAI distribution is substantially lower than the LAI derived from the Olson vegetation data. The difference amounts to nearly 0.8 (or more than 30% of the global (land-) average of the LAI in ECHAM4) for the whole world. It seems that the ISLSCP-maxima in summer over Eurasia and North America reach values close to the annual average in ECHAM4. It can thus be stated that, to a first approximation, ECHAM4 uses the satellite-derived LAI of the growing period throughout the year. Hence, summer transpiration should not be significantly changed when introducing the annual cycle of the ISLSCP-LAI as a new boundary condition in ECHAM4. The phase shift in the dashed and dash-dotted line in Fig. 5.38 is probably due to an inaccurate parameterization of the annual cycle of the LAI in the EM (cf. Eq. 5.1). This again favours ISLSCP to Olson/EM since the ISLSCP data exclude this uncertainty. It is obvious that the Olson derived LAI pattern is closer to the ECHAM4 evolution since the global distribution of the LAI in ECHAM4 is also based on the Olson data.
5.6.3 Results from the 3-D experiment

The 3-D experiment driven with the annual cycle of the LAI, as described in the previous section, displays an important prerequisite for model simulations including any kind of canopy model (see, e.g., experiment EXP7).

The off-line sensitivity studies (cf. Section 4.3.3) have revealed that transpiration strongly increases with increasing LAI, thereby generating a substantial decrease in soil moisture, runoff and bare soil evaporation. The positive impact of the LAI on the latent heat flux leads to a cooler surface boundary layer with increasing LAI.

In 3-D simulations, in contrast, the effect of variations in the LAI on the turbulent heat fluxes and the surface temperature is typically small due to numerous feedbacks between the land surface and the atmosphere which were excluded in off-line simulations. Nevertheless, some notable results are presented in the following sections. The focus is on global averages and some regional features are also presented.

Global averages

The 10-year means over land for a number of surface variables are given in Table 5.15. The impact of the new LAI boundary condition on the surface is minimal on a global scale, i.e. global differences of surface variables between the control simulation and EXP6 are typically small. The total evapotranspiration in the current experiment is reduced by over 1% when compared with the control simulation, corresponding to a change in the latent heat flux of 0.5 Wm$^{-2}$. On the other hand, the sensible heat flux increases...
slightly by approximately 0.5 Wm\(^{-2}\). Therefore, the two components of the turbulent heat fluxes compensate each other. This feature is in line with findings in the off-line experiments. The reduced evapotranspiration is likely related to the LAI which is in EXP6 substantially lower than in CTRL (cf. Fig. 5.38). This yields a decrease in transpiration and in the amount of water held on the canopy’s foliage, thereby reducing the latent heat flux. This enhances the soil moisture content by 1.0%, which leads to more water discharge of approximately 2.7%. Further, higher surface temperatures are expected due to lower energy consumption used in the evaporation process. The changes in the annual global means of the turbulent heat fluxes induce a slight increase of the Bowen ratio (ratio of sensible to latent heat flux), which largely influences the global climate (e.g., Shukla and Mintz, 1982; Yeh et al., 1984).

![Graphs showing Evapotranspiration, Rel. soil wetness, Total precipitation, and Surf. temp. over the Indian subcontinent as calculated in the control simulation and EXP6.](image)

**Figure 5.39:** Simulated annual cycles of evapotranspiration, relative soil wetness, total precipitation and surface temperature over the Indian subcontinent as calculated in the control simulation and EXP6.

Processes which are strongly dependent on the vertical distribution of liquid water vapour in the atmosphere, such as radiation processes, cloud amount and precipitation, show negligible sensitivity to the leaf area index.

It is of particular interest if local changes in the LAI can be related to (local) changes in climate variables. This can be analysed by computing the correlation between the differences in the LAI and the induced changes in surface climate variables. The correlation was very low, which confirms that the hydrological cycle is not a local phenomenon and that heat is efficiently redistributed by advection. Schär et al. (1999) report in their introduction that “the average water molecule travels a large distance before it again returns to the Earth’s surface in the form of falling precipitation”. This precludes any local or direct effect between evapotranspiration and precipitation.
Table 5.15: Comparison of 10-year-means between the control climate and EXP6. Last column: percentage differences. Sign convention: downward fluxes are counted positively. Figures refer to averages over all land points excluding ice covered grid boxes.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP6</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>mm/day</td>
<td>1.543</td>
<td>1.559</td>
<td>-0.016</td>
<td>-1.0%</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm⁻²</td>
<td>-45.0</td>
<td>-45.5</td>
<td>0.48</td>
<td>1.0%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm⁻²</td>
<td>-24.3</td>
<td>-23.7</td>
<td>-0.57</td>
<td>2.4%</td>
</tr>
<tr>
<td>Total runoff</td>
<td>mm/day</td>
<td>0.709</td>
<td>0.691</td>
<td>0.018</td>
<td>2.6%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>2.269</td>
<td>2.268</td>
<td>6 × 10⁻⁴</td>
<td>0.03%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>63.5</td>
<td>62.6</td>
<td>0.9</td>
<td>1.5%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>287.00</td>
<td>286.94</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
<td>Net surface SW rad.</td>
<td>Wm⁻²</td>
<td>143.89</td>
<td>143.93</td>
<td>-0.03</td>
<td>0.02%</td>
</tr>
<tr>
<td>Net surface LW rad.</td>
<td>Wm⁻²</td>
<td>-65.82</td>
<td>-65.56</td>
<td>-0.3</td>
<td>-0.04%</td>
</tr>
</tbody>
</table>

The only correlation coefficient greater than 0.5 was found for the relationship between the LAI and the skin reservoir content. This is reasonable since the maximal amount of the skin reservoir content is directly proportional to the LAI (Eq. 4.12). The sensitivity of the skin reservoir content with respect to the LAI in the 3-D simulation is very close to the corresponding value obtained from the offline simulations (see Fig. 4.14), i.e. the annual skin reservoir content decreases by 0.018 mm when the LAI is reduced by 1.

Impact on India

On a regional scale, larger differences between the current experiment EXP6 and the control simulation are found. To detect regions with statistically significant differences, significance tests were performed using the t-statistic test. The tests for significance were carried out on a monthly basis (95% level). India was the region where, for several parameters, statistically significant differences between the control simulation and EXP6 were detected. Thus, annual cycles of some parameters, averaged over the Indian subcontinent, are presented in Fig. 5.39. Consider for illustration the global pattern of the LAI during summer and winter (Fig. 5.37). The ISLSCP data suggest a substantially lower LAI over India than the original LAI field used in ECHAM4. This explains why evapotranspiration is significantly reduced, which leads to a distinct increase in soil moisture as well as to warmer surface temperatures. This warming is most pronounced during July and August, which also corresponds to the months with a substantial decrease in precipitation (and cloud amount, not shown). Results from model simulation suggest a certain relationship between the difference in evapotranspiration and the difference in precipitation. The multiple R-squared (the correlation squared coefficient) amounts to 0.40 which means that 40% of the deviation in precipitation can be explained by altered evapotranspiration. The conditions over India suggest that, at least on a regional scale, distinct changes in LAI may lead to significant alteration in turbulent heat fluxes and in the hydrological cycle. The radiation budget is also influenced by changes in surface temperature and cloud amount. In July and August global radiation increases by approximately 20 Wm⁻², which increases total net radiation by close to 10 Wm⁻².

Impact on Africa during June through September

The LAI in Africa is significantly reduced in the modified experiment when compared to the control run, i.e. the mean monthly LAI decreases from slightly less than 2 to approximately 1 (cf. Fig. 5.38). It is thus of interest to investigate how the surface climate responds on this decrease in foliage. In Table 5.16, the relevant surface variables
as well as the deviation in EXP6 from the control run are tabulated. The averages refer to the period June through to September. The surface climate over Africa in northern summer/early spring tends to be more humid. This is supported by a distinct increase in the soil moisture (+4.6%), more runoff (+4.3%), a higher evapotranspiration (+3.3%) as well as higher precipitable water, cloud cover (+2.0%) and precipitation (+4.1%). The increased cloudiness reduces the global radiation. This decrease in incoming solar radiation and the cooling effect due to enhanced evapotranspiration, to a slight surface cooling. The direct impact of the LAI, i.e. the less foliage the lower the transpiration rate and thus evapotranspiration, seems to be not of major importance. It is thus likely that an increase in moisture convergence by advective transport from the surrounding sea areas also plays a certain role.

Table 5.16: Comparison between the 4-monthly means (June - September) of the control climate and EXP6 for Africa. The last column depicts the percentage differences. Sign convention: downward fluxes are counted positively.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Unit</th>
<th>EXP6</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>mm/day</td>
<td>1.57</td>
<td>1.52</td>
<td>0.05</td>
<td>3.3%</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>Wm⁻²</td>
<td>-45.5</td>
<td>-44.0</td>
<td>-1.5</td>
<td>3.3%</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>Wm⁻²</td>
<td>-45.5</td>
<td>-44.0</td>
<td>-1.5</td>
<td>-1.8%</td>
</tr>
<tr>
<td>Total runoff</td>
<td>mm/day</td>
<td>0.73</td>
<td>0.70</td>
<td>0.03</td>
<td>4.3%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>2.23</td>
<td>2.14</td>
<td>0.08</td>
<td>4.1%</td>
</tr>
<tr>
<td>Rel. soil moisture</td>
<td>%</td>
<td>54.7</td>
<td>52.3</td>
<td>2.4</td>
<td>4.6%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>299.55</td>
<td>299.62</td>
<td>-0.07</td>
<td></td>
</tr>
<tr>
<td>Cloud cover</td>
<td>%</td>
<td>37.9</td>
<td>37.2</td>
<td>0.8</td>
<td>2.0%</td>
</tr>
<tr>
<td>Vertic. integrated water vapour</td>
<td>%</td>
<td>29.1</td>
<td>28.4</td>
<td>0.7</td>
<td>2.3%</td>
</tr>
<tr>
<td>Vertic. integrated cloud water</td>
<td>%</td>
<td>57.2</td>
<td>54.9</td>
<td>2.4</td>
<td>4.3%</td>
</tr>
<tr>
<td>Global radiation</td>
<td>Wm⁻²</td>
<td>233.7</td>
<td>236.2</td>
<td>-2.4</td>
<td>-1.0%</td>
</tr>
<tr>
<td>Net surface SW rad.</td>
<td>Wm⁻²</td>
<td>177.8</td>
<td>179.5</td>
<td>-1.7</td>
<td>-1.0%</td>
</tr>
<tr>
<td>Net surface LW rad.</td>
<td>Wm⁻²</td>
<td>-78.4</td>
<td>-79.9</td>
<td>1.5</td>
<td>-1.9%</td>
</tr>
</tbody>
</table>

Influence on the summer drying in Europe

It is of some interest to investigate how the introduction of a realistic annual cycle of the LAI affects the soil moisture over Europe. Several studies (Wild et al. (1996), Section 4.7) revealed that in ECHAM, the soil moisture is significantly underestimated during summer and autumn.

In spring, the LAI in the modified run is significantly lower than in CTRL while the difference from June to August amounts to only 0.3 (not shown). This leads, neglecting advective moisture transport, to a decrease in transpiration primarily in spring. The reduced transpiration rate slows down the decrease in soil moisture between spring and summer which results in a higher water level in the model's bucket in summer. In fact, the mean soil moisture content between June and September increases by 6.2%. The substantially moister soils yields a strong increase in surface runoff (+9.2%) but only a slight increase in evapotranspiration (+1.7%). The additional water is efficiently discharged by runoff while the lower LAI may contribute to the moderate increase in evapotranspiration. The higher evapotranspiration and potential changes in the advective moisture convergence lead to enhanced precipitable water (cloud water: +2.1%, water vapour: +2.0%). This again causes a slightly intensified precipitation rate (+1.3%). Global radiation weakens by 2.8 Wm⁻² (1.3%) due to enhanced cloudiness. This results, along with the increase in the evaporative cooling, in a cooler surface by approximately 0.1°C. The above results suggest that the unrealistic summer dryness over Europe is significantly reduced when the 3-D simulation is based on a realistic annual cycle of the leaf are index.
Summary

Summarized global 10-year means of surface variables are only slightly influenced by changes in the LAI distribution. However, on a local scale, both the hydrological cycle and the radiation budget are substantially modified in some regions. Considerable impacts on the surface climate are found in regions such as India, Africa or western Europe where the summer dryness is significantly reduced when using the LAI distribution based on remote sensing. The influence of the LAI probably increases when incorporating more processes in ECHAM which are dependent on the LAI (e.g., the surface albedo of snow covered forests (cf. EXP7) or the radiation transfer through the canopy).

5.7 EXP7: Albedo of snow covered forests and a simple snow interception model

The experiment EXP7 has been performed to investigate and clarify potential impacts of snow covered forests on surface albedo. This requires the implementation of a canopy model suitable for GCM use and a parameterization accounting for snow on the canopy.

5.7.1 Overview of modifications

The Canadian Land Surface Schemes for GCMs (CLASS) has been identified to be appropriate for albedo computations of snow covered forests. This model requires the introduction of a new prognostic variable, the skin reservoir for the snow intercepted on the canopy. The calculation of the sky view factor, which accounts for the degree of crown closure, needs the fraction of both needleleaf and broadleaf trees in each grid square. This procedure is based on the Olson dataset (Olson et al., 1983) and look-up tables in Claussen et al. (1994).

To account for a realistic snow mass intercepted by the canopy, a simple interception model was developed. This model includes, in addition to evaporation and snowfall, the downloading of intercepted snow due to wind and temperature (melt/drip and slipping).

5.7.2 Radiation transfer canopy models

Many canopy models have been developed in the past. Here, only a few which focus on the radiation transfer through the canopy (and hence, incorporate the determination of the canopy albedo) are briefly discussed. In addition, problems inherent to its implementation in GCMs are accounted for.

The albedo of a canopy depends on the height of the plants, the angle and distribution of the leaves, as well as the density of foliage which is closely related to the leaf area index (LAI). The reflected radiation of plant canopies is generally very complex as it is composed of both reflected and multiple scattered radiation by plant organs and by the ground surface. Further complications are inferred when discussing a snow covered canopy.

Yamazaki et al. (1992) developed a simple heat-balance model with two canopy layers. The canopy is divided into a "crown space" and a "trunk space" (without leaves). The crown space is further divided into a lower and upper layer. The radiation fluxes are computed assuming an exponential decrease in the radiation with increasing depth of the canopy layer. The transmittance for direct and indirect radiation are calculated using an attenuation coefficient and randomly distributed leaves. However, since the model requires a number of morphological parameters such as canopy-top height and canopy-
base height which are difficult to determine on a global scale, this model was rejected for use in ECHAM4. Even though, the most appropriate global datasets for describing the vegetation are available in Sellers et al. (1996b), the morphological properties may be afflicted with rather large errors.

Several studies have been published on canopy radiation transfer models using a two-stream approximation (Dickinson, 1983; Sellers, 1985; Joseph et al., 1996). Their method focuses on the radiation fluxes along the vertical (Eq. 5.40). Individual leaves are treated as flat isotropic scattering elements. Therefore, upward and downward diffuse fluxes are completely isotropic:

\[
\begin{align*}
-j_i(dI \uparrow /dL + [1 - (1 - \beta)\omega]I \uparrow - \omega I \downarrow) &= \omega \mu K_0 \exp(-KL) \\

-j_i(dI \downarrow /dL + [1 - (1 - \beta)\omega]I \downarrow - \omega I \uparrow) &= \omega \mu K(1 - \beta_0) \exp(-KL),
\end{align*}
\]

where

- \( I \uparrow \) upward radiative flux
- \( I \downarrow \) downward radiative flux
- \( K \) optical depth of direct beam per unit leaf area
- \( \mu \) average inverse diffuse optical depth per unit leaf area
- \( \omega \) scattering coefficient
- \( L \) cumulative leaf area index
- \( \beta, \beta_0 \) upscatter parameters for the diffuse and direct beam.

Some further assumptions, such as leaves are randomly arranged in space and the application of correct boundary conditions lead to a closed solution of Equation 5.40 (Sellers, 1985). The canopy albedo \( \alpha_c \) can be written in the form

\[
\alpha_c = I \uparrow (0) = h_1/\sigma + h_2 + h_3,
\]

where the parameters \( h_1, h_2, h_3 \) and \( \sigma \) are functions of
- the scattering coefficient for leaves and soil
- the leaf area index
- the leaf angle distribution
- the angle of incident radiation.

Fig. 5.40a,b show canopy albedo in the visible radiation and NIR spectrum, based on leaf properties given in Table 5.17 and Eq. 5.41. Snow is presumed on the ground below the canopy with a reflectance of 0.65 and 0.95 in the NIR and VIS, respectively.

| Table 5.17: Optical properties of the “standard” leaf |
|---------------------------------|------|------|
| scattering coefficient (\( \omega \)) | reflectance | transmittance |
| VIS | 0.175 | 0.6 \( \omega \) | 0.4 \( \omega \) |
| NIR | 0.825 | 0.7 \( \omega \) | 0.3 \( \omega \) |

Fig. 5.40 clearly shows that the sensitivity of the canopy albedo to the LAI is highest for low LAIs. \( \alpha_{c, VIS} \) decreases considerably with increasing zenith angle for LAI = 1. This is due to a decreasing percentage of the direct SW radiation which reaches the bright snow covered surface. The high (ground-) snow albedo increases the probability that photons may partially escape the canopy without further scattering at the canopy’s foliage. The effect of the LAI on \( \alpha_{c, VIS} \) is minimal for LAIs larger than 3. Furthermore, the surface albedo of the surface underneath the canopy, hardly influences the total canopy albedo. The sensitivities of \( \alpha_{c, VIS} \) and \( \alpha_{c, NIR} \) to the LAI and zenith angle differ strongly. For
LAIs above 3, the response of $\alpha_{c,NIR}$ to the LAI is considerably larger than the effect of the LAI on $\alpha_{c,VIS}$. This is due to the large value of the scattering coefficient of green leaves in the NIR spectrum (Table 5.17). Furthermore, $\alpha_{c,NIR}$ is more sensitive to the soil albedo than $\alpha_{c,NIR}$ due to the large scattering coefficient of leaves in the NIR.

Fig. 5.40c displays the total canopy albedo for a clear sky. It illustrates the weak angular dependence for zenith angles less than 60°. The minimal zenith angle dependence of $\alpha_c$ is confirmed by many other studies (e.g., Jarvis et al., 1976; Verseghy et al., 1993; Yin, 1998).

The calculation of the canopy albedo using the two-stream approximation is computationally relatively cheap. However, it is not a simple task to make all the necessary input data on the T42 mesh available. Moreover, errors in the specification of boundary fields will increase the uncertainty. Therefore, the two-stream approximation was rejected for implementation in ECHAM4. This approach may be more appropriate when a detailed investigation of the transpiration rates within the canopy, or the radiation available for snow melt underneath a canopy, is required.

A different approach for computing canopy albedos has been proposed by Otterman (1984). He treats, in a simplified way, forests as a surface with vertical tree trunks. The direct solar beam is reflected at the protruding vertical plant elements. The canopy albedo, computed with the model of Otterman (1984) (not shown) reveals large deviations from other model results. Furthermore, the projection of the cylindrical plant elements on a
vertical plane (parameter $s$), is not globally available. The approach for a forest, based on cylindrical plant elements, is presumably too primitive for a correct computation of the radiation transfer within a real canopy because the foliage and branches are neglected. Several studies show that canopies reflect radiation quite anisotropically (Goudriaan, 1977; Verstraete et al., 1990; Dickinson et al., 1990). In other words, the measured reflectance of such a surface depends not only on the structure of the surface, but also on the relative position of the observer. These models are important for the interpretation of satellite-derived measurements but are too sophisticated for use in GCMs.

Since none of the previously discussed models can be easily adopted in ECHAM, the algorithm used in the Canadian Land Surface Scheme (CLASS) was implemented in ECHAM4. This scheme is described in detail in the next section.

### 5.7.3 Surface albedo of snow covered canopy in CLASS

The parameterization of the surface albedo for snow covered canopies in EXP7 is adopted from CLASS (Canadian Land Surface Scheme for GCMs) (Verseghy, 1991). This model allows for snow on the canopy. It is based on a simple algorithm but is dedicated to capture the principal relationships between the canopy albedo and snow on both the underlying ground and the canopy. The model gives reasonable results and the generated canopy albedos compare reasonably well with canopy albedos as obtained with more sophisticated models.

The key parameter for the computation of the albedo of snow covered forests in CLASS is the sky view factor (SVF) which describes the degree of canopy closure. The SVF is related to an exponential function of the LAI (Eq. 5.42).

\[
SVF = e^{-0.5LAI} \quad \text{(needleleaf trees)} \\
SVF = e^{-1.5LAI} \quad \text{(broadleaf trees)}
\]

The total surface albedo is computed as

\[
\alpha = SVF \cdot \alpha_g + (1 - SVF) \alpha_c,
\]

where $\alpha_g$ is the albedo of the ground underneath the canopy and $\alpha_c$ is the albedo of the canopy. The snow albedo on the ground underneath the forest is assumed to be the same as in the open area, which is in line with the findings of Pomeroy and Dion (1996). $\alpha_c$ is given by

\[
\alpha_c = f_{sc} \cdot \alpha_{sc} + (1.0 - f_{sc}) \cdot \alpha_{c,snowfree}
\]

The fraction of the canopy covered by snow, $f_{sc}$, is defined as

\[
f_{sc} = S_{nc}/S_{nc,max}
\]

where $S_{nc}$ is the water equivalent of snow intercepted by the canopy and

\[
S_{nc,max} = 0.2LAI.
\]

Verseghy (1991) reports that Eq. 5.46 works well for both rain and snow and for a wide variety of vegetation types and precipitation events. This means that for LAI = 5, an amount of snow equal to 1 kgm$^{-2}$ (or 1 cm of fresh snow) is sufficient to fill the canopy storage capacity. The albedo of snow covered canopy, $\alpha_{sc}$, is set equal to 0.20, from values given in the literature (e.g., Leonard and Escher, 1968; Verseghy, 1991; Harding and Pomeroy, 1996; Pomeroy and Dion, 1996).
In Fig. 5.40d, the canopy albedo is calculated for both needleleaf trees (solid line) and broadleaf trees (thin dashed line). The forest is assumed to be snow free with $\alpha_c = 0.16$. The soil is completely snow covered ($\alpha_s = 0.8$). Note the large difference between both forest types. The more sophisticated scheme used in the two-stream approximation (thick dashed line) leads to a similar functional form of the relation between the canopy albedo and LAI. A mixed forest with the same fraction of needleleaf and broadleaf trees yields a fairly good agreement between CLASS and the two-stream approximation, described in Sellers (1985). The albedos for needleleaf trees might be slightly too high for sparse foliage (Fig. 5.40d). However, the boreal forests typically do not drop their leaves. This avoids low LAIs for needleleaf forests during the cold season.

In order to apply Eq. 5.42, it is necessary to distinguish between needleleaf and broadleaf trees. The detailed procedure is as follows:

1) The Olson vegetation classification (Olson et al., 1983) is used to determine the fraction of needleleaf and broadleaf trees. Based on a look-up table in Claussen et al. (1994), therein Table 2), the fraction of needle- and broadleaf trees for each land type is determined.

2) Area weighed interpolation is applied to transform the data from a regular $1 \times 1^\circ$ grid onto the T42-grid.

3) Monthly LAIs for both forest types are determined by using tabulated maximum and minimum LAIs in Claussen et al. (1994).

4) The final monthly LAIs are calculated such as to fullfill the following conditions:

- The LAI in each grid box must be identical with the ISLSCP-derived LAI.
- The seasonal cycle of the LAI over forests is identical with the seasonal cycle of the LAI averaged over the entire grid box, i.e. respective LAI ratios between different months are maintained.

5.7.4 Snow interception on canopies

Intercepted snow on forests plays a major role in the snow hydrology of forests. The impact of intercepted snow on (long-term) forest albedos is, however, of negligible importance (e.g., Betts and Ball, 1997; Pomeroy and Dion, 1996). On a shorter time scale, however, snow interception may significantly change the reflectance of a forest. For the hydrology of cold climate forests it is likely necessary to calculate the amount of intercepted snow (Hedstrom and Pomeroy, 1998). Several studies deal with observations where significant amounts of snow exist for long periods on canopies (Nakai et al., 1994; Hedstrom and Pomeroy, 1998). Nevertheless, intercepted snow on canopies is neglected in most GCMs.

The main goal of the implementation of a simple interception model is to investigate to what extent the ECHAM GCM is applicable to analyze the different interception processes. These include the interception efficiency (interception/snowfall), sublimation and unloading of intercepted snow due to wind and melt.

While the snow interception algorithm in CLASS allows the intercepted snow to decrease only as a result of sublimation and melt, the new algorithms for the mass balance of the snow intercepted by the canopy per unit area evolves according to the following equation:

$$\frac{\partial S_{nc}}{\partial t} = P_{sc} - S_{nc} \cdot f(T) \cdot f(v) - E_{sc},$$

with

- $E_{sc}$: Evaporation rate from the skin reservoir for intercepted snow [kg m$^{-2}$ s$^{-1}$]
P_{sc} \quad \text{Snowfall rate per unit area intercepted by the canopy [kgm}^{-2}s^{-1}] \\
S_{nc} \quad \text{Snow water equivalent of the snow intercepted by the canopy [kgm}^{-2}] \\
f(T) \quad \text{Function which describes unloading of intercepted snow caused by temperature} \\
f(v) \quad \text{Function to describe wind-induced downfall of intercepted snow.}

This approach requires the introduction of a new prognostic variable in the ECHAM4, namely $S_{nc}$. It is disadvantageous that $S_{nc}$ is difficult to validate since observational data are sparse. In addition, they are all limited to isolated trees (e.g. Pomeroy and Dion, 1996) and are therefore little representative for extended forests with largely varying characteristics. Eq. 5.47 includes all various processes affecting the amount of snow intercepted by the canopy, i.e.:

(1) sublimation of intercepted snow
(2) maximal amount of intercepted snow
(3) unloading due to temperature (melt/drip and slipping) (Eq. 5.48)
(4) unloading due to wind (Eq. 5.49).

In the following, a parameterization set is developed to account for all the four above-described processes.

(1) Hedstrom and Pomeroy (1998) and Lundberg et al. (1998) quote various studies which identified sublimation as an important process affecting the accumulation of snow in forests. The sublimation of intercepted snow is assumed to have the same rate as the sublimation rates calculated for snow on the ground. This condition may overestimate the sublimation of intercepted snow. Nakai et al. (1994) suggest from observational studies to reduce the potential evaporation for partly snow-covered canopies. They express the reduction factor by the snow storage on the canopy divided by its maximum value.

(2) is simply accounted for by the assumption that the total amount of intercepted snow does not exceed the value $S_{nc,\text{max}}$, as defined in Eq. 5.46. However, in reality, the maximal snow load on the canopy is dependent on the cohesion of snow to the branch, the strength of intercepted snow masses and the branch elasticity (Schmidt and Pomeroy, 1990). It has been reported (Pomeroy and Gray, 1995; Betts and Ball, 1997; Hedstrom and Pomeroy, 1998) that after heavy snowfalls, the intercepted snow may be substantially larger than in Eq. 5.46.

(3) Unloading processes due to temperature become more significant for temperatures close to melting point. This is associated with a decrease in snow strength and an increase in bending of the branches. Schmidt and Pomeroy (1990) found that branches are substantially stiffer at colder temperatures. They report stiff branches at $T = -12^\circ C$, while at $T = -2^\circ C$, bending of branches occurred that would induce unloading of snow. Gubler and Rychetnik (1991) suggest that unloading associated with increasing air temperature is rather affected by a decrease in snow strength than by the increase of bending. However, visual observations in the boreal forests suggest that the increased bending is more important than the decrease in snow strengths. Pomeroy and Gray (1995) reflect that these differences are most probably due to changes in the upward longwave radiation during the night. This, in turn, leads to differences in the amount of heat required to warm the intercepted snow.

In order to keep Eq. 5.47 as simple as possible, $f(T)$ is specified as a linear function in $T$ [$^\circ C$]

\[
f(T) = \frac{3^\circ C + T}{60^\circ C}. \tag{5.48}
\]

The function was determined in such a way as to fulfill $f(-3^\circ C) = 0.0$. The unloading rate is assumed to vanish for temperatures below $T = -3^\circ C$. The value of the denominator
(= 60°C) allows for unloading half of the intercepted snow during 12 hours at $T = 0$°C (see Fig. 5.41). For a temperature of 2°C, approximately 80% of the intercepted snow is unloaded during 12 hours. The considerable increase in unloading for air temperature above 0°C agrees with the findings of Nakai et al. (1994) who reported a rapid decrease of snow on trees after a snowfall caused by slipping and melt. Yamazaki et al. (1996) reported, in line with Fig. 5.41, that after snowfall, the crown-snow ratio decreases exponentially with time, depending on the air temperature and wind speed. They also found that the response time $\tau$ (reduction to ~37% of initial value) of the crown-snow is about 1/2 day when the air temperature is below 0°C, and 1 - 5 hrs when it is above 0°C. However, scattering is large and measurements presented in Yamazaki et al. (1996) contain also $\tau$-values of some days for freezing temperatures. This in reasonable agreement with the newly-developed model.

![Figure 5.41: Unloading of intercepted snow due to temperature and wind. At the time $t = 0$, the canopy is assumed to support the maximum snow load, that is (normalized) interception $I = 1$. a) is based on Eq. 5.49, b) on Eq. 5.48. Time step is $\Delta t = 24$ min.](image)

(4) Release of intercepted snow, triggered by branch movement due to wind influence, can generally be observed after snowfall. Unloading may also be caused by the atmospheric shear stress exerted by wind on the branches and snow. Betts and Ball (1997) investigated Boreal Ecosystem-Atmosphere Study (BOREAS) measurements from 1994 and 1995. They revealed that (winter) forest albedos above 0.3 correspond to days with low wind speed of less than approximately 3 ms$^{-1}$. Miller (1962) reports that snow interception considerably decreases when the wind speed during snowfall is larger than 2 ms$^{-1}$. To make the approaches consistent, the algorithm for the wind induced downfall of intercepted snow was assumed to be similar to that of temperature. The function $f(T)$ was replaced by $f(v)$, with $v$ representing the wind speed at 10 m above the ground, which corresponds, to a first approximation, to the mean canopy height (Eq. 5.49). As above for $f(T)$, $f(v)$ is specified as a linear function in $v$ [ms$^{-1}$]:

$$f(v) = \frac{v}{150}\text{ ms}^{-1}.$$  \hspace{1cm} (5.49)

The denominator has been selected such as to unload 50% of the intercepted snow within 6 hours for $v = 5$ ms$^{-1}$. This interception model does not presume that the intercepted snow load approaches zero between each snowfall event as most simple interception models do (Hedstrom and Pomeroy, 1998).
5.7.5 Results from the 3-D experiment

The simulation with the snow submodel implemented in CLASS was conducted to investigate the albedo and snow conditions of snow covered forests. The new formulation, developed for the unloading of intercepted snow, is discussed in detail. Since CLASS incorporates the leaf area index as a key parameter, EXP6 (which includes a realistic seasonal cycle of the LAI) was designed as the control experiment. Thus, all the differences in this chapter refer to EXP6 (cf. Section 5.6).

Annual means

Annual means of the Northern Hemisphere (restricted to regions with measurable snow deck in February) are presented in Table 5.18. Considerable differences are found for snow water equivalent, snow cover fraction, surface albedo and surface temperature. The decreasing albedo leads to higher temperatures and, thus, to less snowfall and earlier snow melt in spring. The higher temperatures yield reduced stability and enhanced turbulent heat fluxes. This enhances the vertically integrated water vapour (and cloud water) by about 2%, which leads to a slight increase in precipitation. Nevertheless, the snowfall rate decreases due to higher temperatures, mainly during the transient seasons. It is noteworthy that the above discussion is based on purely local considerations. Local processes are, however, also affected by large-scale advective processes (e.g. moisture convergence).

Higher deviations between the surface climate in EXP7 and EXP6 occur when reducing the area to, e.g., boreal forests. Several deforestation experiments (of boreal forests) have revealed similar tendencies as for the differences between the control run and the simulation with the modified model: Thomas and Rowntree (1992) showed that the removal of the boreal forests increases the land surface albedo and snow depth but decreases air temperature, surface net radiation, sensible heat flux, latent heat flux, and precipitation during the months of March, April and May, compared to simulations which include the boreal forests.

The impact caused by changes in the surface albedo is likely larger when including the oceanic feedback instead of prescribing the sea surface temperature (Bonan et al., 1995): colder winter climate increases the extent of sea ice, thereby reinforcing the cooling caused by higher ocean albedos.
Geographical distribution of albedo and temperature in March

Largest deviations between EXP7 and EXP6 (defined as control simulation throughout this chapter) occur in spring. Fig. 5.42 shows the long-term differences for the surface albedo and 2-m temperatures in March. The pronounced decrease in the surface albedo leads to substantially higher net shortwave radiation and, consequently, to enhanced surface temperatures. Fig. 5.42c indicates that the differences are statistically significant on the 95% level, using the t-statistic test. Largest albedo differences are found over the boreal forests in both the higher latitudes of Eurasia and North America. This feature is mainly attributed to Eq. 5.43 which assumes the albedo of snow covered forests to be 0.2. This value has been confirmed to be realistic by several authors (Verseghy, 1991; Harding and Pomeroy, 1996; Pomeroy and Dion, 1996; and others). A crude estimate of the maximum surface albedo of snow covered forests, as calculated with CLASS, leads to approximately 0.25, in agreement with observational studies. This is distinctly lower than the albedos suggested in ECHAM (0.4 for boreal forests in winter with $T < -10^\circ$C). The above estimate is based on the following assumptions: sky view factor $SVF = 5 - 6\%$ (Eq. 5.42 for needleleaf trees with LAI = 6), snow on the ground with $\alpha_s = 0.8$.

The impact of changes in the forest albedo on the absorbed SW radiation and surface temperature from December through February is small due to low sun and therefore low global radiation.

Assessment of the changes in the Taiga

This section outlines the effect of a reduced surface albedo in the Taiga on various surface climate variables. The Taiga is defined as the grid boxes where the vegetation type 'Main Taiga', according to Olson et al. (1983), is dominant. This definition agrees well with the grid boxes situated north of 45°N and which contain a forest fraction $a_f > 0.5$. The climate of the boreal forests have been investigated by many authors (Bonan et al., 1992; Chalita and Treut, 1994; Bonan et al., 1995; Douville and Royer, 1997; Harding and Pomeroy, 1996; Pomeroy and Dion, 1996). The marked impact of the masking effect on the surface albedo, and consequently on the energy balance, has often been emphasized. It has been shown, that the surface albedo of the winter boreal landscape greatly affects the Northern Hemisphere climate at all times of the year.

Fig. 5.43 presents the monthly differences for EXP7 - EXP6 over the Taiga. The marked decrease in the surface albedo is clearly visible and equals approximately 0.1 in winter, the main contribution being the lower albedo of snow covered forests (see previous section). The average Taiga albedo (including non-forested areas) amounts to 0.499 in the control simulation, whereas EXP7 provides a value of 0.40. This change affects the net shortwave radiation mainly in late winter and spring due to a rapidly increasing TOA incoming shortwave flux. Greatest differences between the two 3-D simulations typically occur in April. Changes in net shortwave radiation are similar to changes in net radiations as net longwave radiation is hardly affected by albedo modifications (Fig. 5.43). The higher temperature in spring yields an earlier snow melt, further reducing the albedo (positive feedback). Comparisons on a daily basis have revealed that the retreat of the snowline in the Taiga is approximately one week earlier in EXP7 than in EXP6, which is in better agreement with the observations (cf. Section 3.4). The earlier snow melt is also documented in Fig. 5.43: positive differences in March/April but negative in May/June.

Despite slightly increased precipitation in April, the snowfall rate decreases by close to 0.1 mm/day (~10% of total snowfall) due to distinctly increased surface temperatures. The increase in the cloud water amount in spring (Fig. 5.43) is probably related to the higher temperatures reinforcing the hydrological cycle, which is in line with enhanced
Figure 5.42: a) Mean differences between EXP7 and EXP6 for March (10-year means). a) Surface albedo, b) 2-m-temperature, c) t-statistic test for surface albedo - dark shaded areas differ on the 95% level. Areas shaded in bright grey depict the area with measurable snow in February.

Turbulent fluxes (not shown). The cloud water amount increases by almost 6 gm\(^{-2}\) or \(\sim 10\%\). The relative deviations in the vertically integrated water vapour amount (not shown) are of similar magnitude. The increased amount of atmospheric water reduces the global radiation by a few Wm\(^{-2}\), which counteracts the increase in net shortwave radiation due to the lower surface albedo. However, the reduced albedo dominates the enhanced amount of cloud droplets and water vapour with respect to net shortwave radiation.

The surface pressure distinctly decreases in spring when using the snow submodel of CLASS (Fig. 5.43). The pressure change of approximately 1 hPa implies a warming in a fairly thick atmospheric air layer: A rough estimate results in an air layer with a thickness of more than 2 km to obtain the above detected pressure anomaly. The computation is based on (i) the ideal gas equation and an isotherme atmosphere (Ling, 1982), (ii) a mean layer temperature of 8.5\(^\circ\)C with its lower level being at sea level, (iii) an increase in sea surface pressure of 1.0 hPa, and (iv) a warming by 1\(^\circ\)C. A comparison of the surface wind
pattern between EXP6 and EXP7 (not shown) revealed an intensification of the wind especially over Europe and western Russia. To what extent this is related to the pressure anomalies would need further investigation.

The memory effect of the atmosphere seems to be short for all surface climate variables: In June, the differences between the two model simulations (EXP7 and EXP6) are generally small and thus not significant. The differences might be higher with interactive oceans included, due to much longer response times as are typical for the atmosphere.

**Sky view factor and albedo in the boreal forests**

This section concerns the mean sky view factor \( SVF \), cf. Eq. 5.42 and its impact on total surface albedo, focusing on the boreal forests.

The mean LAI, averaged over the Taiga, amounts to approximately 5.5 in winter and 6 in summer. The respective simulated sky view factor are 5.6% and 4.5%, respectively. The snow covered fraction of the canopy, \( f_{sc} \) (Eq. 5.45), is also calculated and stored. The simulated value of \( f_{sc} \) is close to 0.2 for winter boreal forests, corresponding to a mean snow load of approximately \( S_{sc} = 0.2 \text{ mm} \) (Eq. 5.46 with \( \text{LAI} = 5 \)). Based on these values as well as \( \alpha_g = 0.8 \) (snow on ground), and the albedo of snow free evergreen forest being 0.13 (Claussen et al., 1994), Eq. 5.43 yields a total surface albedo of \( \alpha = 0.18 \). This value is in close agreement with ground based observations of winter boreal surface albedo (Betts and Ball, 1997; Harding and Pomeroy, 1996; Pomeroy and Dion, 1996).
Interception

This section outlines how much snow is intercepted on the canopy and how this snow is unloaded or sublimated. The features attributed to the interception processes are shown for the month of March, the relevant processes being similar for other (winter) months.

**Figure 5.44:** Snow interception in March (10-year means) as simulated in EXP7. a) Absolute amounts (mm/month); b) Intercepted snow divided by total snowfall (%).

Fig. 5.44 shows the accumulated intercepted snow in March (10-year average). The broad belt of boreal forests with high winter interception caused by high LAIs, is clearly visible. Fig. 5.44b displays the ratio between the intercepted snow and the total snowfall. Typical ratios of evergreen forests range from 20 - 30%. This value is in the right order of magnitude: Pomeroy and Schmidt (1993) report that as much as 60% of cumulative snowfall may be intercepted by the boreal forest in mid-winter. Hedstrom and Pomeroy (1998) compare results from an interception model with observations located in a mid-continental southern boreal forest. The measurements from both black spruce and jack pine suggest that approximately half of the snowfall is intercepted, again in relatively good agreement with EXP7. The relatively low values in northeastern Canada are related to the high snowfall rates in this area: higher snowfall rates generally lead to, everything else being the same, higher throughfall (precipitation that penetrates the canopy and reaches the soil surface).

In addition to the snow accumulation on trees, it is relevant to know the magnitude of the processes which reduce the intercepted snow. In the model configuration, downloading due to temperature (melt/drip/slipping), wind (vibration) and sublimation are distinguished (see previous section). The geographical distribution of the above processes, expressed in percentages of total intercepted snow, is shown in Fig. 5.45. It is evident that the melt and slipping due to temperature (Fig. 5.45a) increases with increasing temperature and thus, generally increases from north to south. Since the unloading is zero for temperatures
below $T = -3^\circ$C, extended regions with monthly mean temperatures below approximately $T = -5^\circ$C, are coloured in bright grey. Highest values are close to 70%, being in accordance with observations, which suggest fast melt and slipping of intercepted snow close to and above melting point (Nakai et al., 1994).

Fig. 5.45c shows the percentage of intercepted snow released by sublimation. High values are found in most regions excluding very cold areas in Canada, Alaska and Siberia. These regions are characterized by very low evapotranspiration rates which are, according to the control simulation, in the range of 0.1 - 0.2 mm/day, implying low sublimation rates as well. In boreal forests, in contrast, substantially higher sublimation rates (typically 50% of total intercepted snow) are generated. This is likely due to slightly warmer temperatures, rougher surfaces (forest) and lower stability of the boundary layer. This is in quite good agreement with annual sublimation losses ranging from 30% - 40% of annual snowfall for completely coniferous canopies reported in Pomeroy and Schmidt (1993). The tongue-like area over the U.S. Great Plains with rather low values is caused by mean temperatures close to freezing point which favour melt and slipping of intercepted snow (see Fig. 5.45a).
Fig. 5.45b illustrates how the wind-induced downfall generally increases with latitude, the main reason being not an increasing wind speed but a reduction of the sublimation rate. Since in the northern parts of Eurasia and Canada, the temperature influence is essentially zero in March and sublimation rates are low, only the vibration of trees contributes significantly to the downloading of intercepted snow: Relatively high values are simulated in a broad belt in western Russia between 50°N and 60°N, caused by relatively high wind speeds of frequently more than 5 ms⁻¹ in March. Since only few observational studies on the wind induced downfall of intercepted snow exist, it is difficult to argue how well the model results fit the observations. The magnitudes are likely to be realistic since the percentage of intercepted snow, the downloading triggered by temperature as well as the sublimation all agree well with observational studies.

Conclusion
To summarize, the surface albedos of snow-covered forests, as simulated in EXP7, are significantly lower than in the control simulation and agree well with observations. The implemented interception model captures the main processes with reasonable skill. The main differences between the control simulation and EXP7 are generated in the boreal forests where significant changes in the near-surface climate have been detected.

5.8 EXP8: Modification of snowflakes’ melt temperature
The percentage of precipitation reaching the ground as snow, is fundamentally for the build-up of snow decks. In most snow models, a threshold temperature is defined where snow flakes change into rain drops (snow-rain criterion). Randall et al. (1994) summarize the temperature conditions for 11 GCMs that are required for precipitation to fall as snow. Loth et al. (1993) and Yang et al. (1997) found a substantial sensitivity of (modeled) snow height to the snow-rain criterion. A short overview of threshold temperatures proposed in literature is given in the next but one subsection.

5.8.1 Overview of modifications
In this experiment, the mean layer temperature above which snow flakes are allowed to melt, is reduced from 2°C to 1°C. This model deficiency was identified by comparing simulated and observed ratios between snowfall and rain against surface temperature.

5.8.2 Snow-rain criteria
A number of snow-rain criteria are listed in Table 5.19. Note the large temperature range between 0.0 and 2.5°C where snowfall and rain are produced with equal probability. The threshold temperature obviously depends not only on the temperature but also on the location. Yang et al. (1997) and Rohrer (1992) emphasize that threshold temperatures usually differ between mountainous regions and plains. They further suggest that the magnitude of the precipitation affects the snow-rain criterion. Thus, differences are expected between summer and winter precipitation.

Fig. 5.46 illustrates the probability of snowfall as a function of the screen temperature. The figure provides an estimate of the temperature where 50% of the precipitation falls as snow: $T = 1°C$ is a good synthesis of all available observations.

In ECHAM4, another approach is applied to describe the form of the precipitation. It is assumed that snow flakes melt when the average temperature of the corresponding model
Table 5.19: Threshold temperature for the distinction of snowfall and rain (snow-rain-criterion).

<table>
<thead>
<tr>
<th>source</th>
<th>threshold temperature [°C]</th>
<th>remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rachner (1983)</td>
<td>0.0</td>
<td>cold region (Hokkaido in Japan)</td>
</tr>
<tr>
<td>Motoyama (1990)</td>
<td>0.0</td>
<td>(deduced from simple model)</td>
</tr>
<tr>
<td>Yang et al. (1997)</td>
<td>1.0</td>
<td>only snow and rain events are counted</td>
</tr>
<tr>
<td>Wilhelm (1975)</td>
<td>0.75</td>
<td>rain and mixed precipitation added in one category</td>
</tr>
<tr>
<td>Rohrer (1992)</td>
<td>1.0</td>
<td>50% snow, 73% snow and mixed precipitation</td>
</tr>
<tr>
<td>Motoyama (1990)</td>
<td>1.0 - 2.0</td>
<td>estimation by eye from graphs</td>
</tr>
<tr>
<td>BATS (Dickinson et al., 1993)</td>
<td>2.2</td>
<td>warm region (Honshu in Japan)</td>
</tr>
<tr>
<td>Auer (1974)</td>
<td>2.5</td>
<td>probability of snow and rain are equal</td>
</tr>
<tr>
<td>Anderson (1976)</td>
<td>1.0</td>
<td>wet bulb temperature</td>
</tr>
<tr>
<td>Siemer (1988)</td>
<td>2.0</td>
<td>wet bulb temperature</td>
</tr>
</tbody>
</table>

Layer exceeds $T = 2°C$. Melting is limited by maintaining the cooling of the layer (heat consumption due to snow melt) in such a way that the temperature of the layer after the melting does not fall below $T = 2°C$. Assuming an unstable stratification in the lower atmosphere, which generally applies during snowfall events in low pressure areas, the above assumption yields, for a screen temperature of $T = 2°C$, to more snowfall than rain. This is, however, inconsistent with the observational data.

Further clarification is achieved by computing the mean modeled ratio between snow and

total precipitation. For this purpose, 6-hourly model data from the T42 control simulation were extracted. Averages were determined over the whole period comprising four winter months December through March. The result is displayed in Fig. 5.46. It is evident that, for \( T > 0^\circ\text{C} \), the ratio of snowfall to rain is substantially overestimated in the control simulation. The agreement can be significantly improved when the melt of snow flakes is initialized at lower temperatures (cf. next section). It is noteworthy that ECHAM4 simulates, even for temperatures far below zero degrees, a substantial percentage of rainfall, which is not supported by any observations. This may be related to an underestimated rate of snow flakes’ melt, an overestimation of the velocity of falling rain drops or a delayed onset of the melting process. These rainfall events for temperatures below zero degrees occur throughout every season with almost equal probability. In contrast to that, calculations have revealed, that the intensity of precipitation seems to play a significant role: Heavy precipitation, defined here as events with more than 3 mm/hour, rarely falls as rain for temperatures below \( T = -3^\circ\text{C} \) \( (T = -3^\circ\text{C}: 98\% \text{ snowfall}, \ 2\% \text{ rain}) \). Thus, the model produces unrealistic ratios between snowfall and rain, predominantly for low precipitation intensities.

A review of the study on which the ECHAM formulation is based (Mason, 1971) reveals that the threshold temperature of 2°C in ECHAM4 is not well supported by that study. Mason (1971) considers a radar observational study where the maximum of the melting band lies 100 ± 50 m below the 0°C level. This finding does not exclude the initialization of the snow flakes’ melt at \( T = 1^\circ\text{C} \) instead of \( T = 2^\circ\text{C} \).

For all these reasons, it is strongly recommended to allow for melt of falling snow flakes when temperatures fall below two degrees. Thus, the simulation EXP8 has been performed using \( T = 1^\circ\text{C} \) as the threshold for melt instead of \( T = 2^\circ\text{C} \).

Further studies related to the snow-rain criterion have been conducted using off-line model experiments with forcing from Russian sites (see Section 4.8).

5.8.3 Results from the 3-D experiment

EXP8 is appropriate for investigating the ratio between snowfall and total precipitation. The experiment provides deeper insight into the relevance of the temperature at which snow flakes melt.

Half-year means for the Northern Hemisphere

In Table 5.20, 10-year means of some surface variables in the control run and EXP8 are compared. The averaging period is restricted to December through to May, since during summer and autumn the differences between the control run and the modified simulation are generally small.

Despite slightly increasing total precipitation in EXP8, compared to CTRL, snowfall is significantly reduced, the reason being the changed threshold temperature for the initiation of snow flakes’ melt in any atmospheric model layer. This decrease implies that snowfall through atmospheric layers with a mean temperature between 1°C and 2°C occurs quite often. The decreasing snowfall rate yields a lower snow melt rate. However, the decrease in snowfall is not entirely compensated by a lower melting rate. This leads to a considerable decrease in the snow mass and consequently, in the surface albedo. The negative albedo anomaly enhances the amount of absorbed energy at the ground, inducing additional heating of the Earth’s surface. This is reflected in a 0.5°C higher surface temperature. The retreat of the snowline in spring is, therefore, substantially earlier in EXP8, compared to CTRL, leading to a closer agreement with the observations in spring (cf. Section 3.4).
Table 5.20: Comparison of 10-year-means (February - May) between the control climate and EXP8. Figures refer to averages over all land points where in February, S > 0.1 cm (according to the USAF snow depth climatology, Foster and Davy (1988)), excluding ice covered grid boxes. The last column depicts the percentage differences. Sign convention: downward fluxes are counted positively.

<table>
<thead>
<tr>
<th>parameter</th>
<th>unit</th>
<th>EXP8</th>
<th>CTRL</th>
<th>Difference</th>
<th>Difference (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow cover</td>
<td>%</td>
<td>54.3</td>
<td>55.6</td>
<td>-1.3</td>
<td>-2.4%</td>
</tr>
<tr>
<td>Snow water eq.</td>
<td>cm</td>
<td>5.6</td>
<td>5.9</td>
<td>-0.3</td>
<td>-4.4%</td>
</tr>
<tr>
<td>Snowfall</td>
<td>mm/day</td>
<td>0.80</td>
<td>0.94</td>
<td>-0.05</td>
<td>-5.9%</td>
</tr>
<tr>
<td>Snow melt</td>
<td>mm/day</td>
<td>0.88</td>
<td>0.72</td>
<td>-0.04</td>
<td>-5.3%</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>%</td>
<td>30.6</td>
<td>40.4</td>
<td>-9.8</td>
<td>-2.0%</td>
</tr>
<tr>
<td>Net SW, surface</td>
<td>Wm⁻²</td>
<td>71.4</td>
<td>70.5</td>
<td>0.9</td>
<td>1.3%</td>
</tr>
<tr>
<td>SW, up, surface</td>
<td>Wm⁻²</td>
<td>-34.7</td>
<td>-35.5</td>
<td>0.8</td>
<td>-2.2%</td>
</tr>
<tr>
<td>Net LW, surface</td>
<td>Wm⁻²</td>
<td>-44.7</td>
<td>-44.5</td>
<td>0.2</td>
<td>0.5%</td>
</tr>
<tr>
<td>2-m temperature</td>
<td>K</td>
<td>264.7</td>
<td>264.2</td>
<td>0.5</td>
<td></td>
</tr>
<tr>
<td>Evapotranspiration</td>
<td>mm/day</td>
<td>0.78</td>
<td>0.77</td>
<td>0.01</td>
<td>1.9%</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>mm/day</td>
<td>1.52</td>
<td>1.51</td>
<td>0.01</td>
<td>0.7%</td>
</tr>
<tr>
<td>Convective precipitation</td>
<td>mm/day</td>
<td>0.28</td>
<td>0.25</td>
<td>0.03</td>
<td>9.5%</td>
</tr>
</tbody>
</table>

Figure 5.47: Change in the ratio between snowfall and total precipitation due to a change in the threshold temperature from 2°C to 1°C, 10-year average of April. Unit: %. Contour lines: Simulated mean monthly surface temperature in April, ±/- 5°C.

A interesting detail is the pronounced increase of the convective precipitation. Although rather small in absolute amount, a pronounced positive difference was observed during each month from December through to May. The question arises whether this is induced by decreased stability or if another process is responsible for these results. Further analyses suggest that, in the atmosphere, the enhanced melting cools the corresponding model layer and thus, reduces the stability. At the surface, in contrast, less energy for melt is consumed (due to the decreased snowfall and thus lower albedo), leading to a warmer surface in EXP8 as in the control simulation. The enhanced surface heating further destabilizes the stratification of the atmosphere.

For investigations of shorter time scales, it is relevant to note that the temporal distribution of runoff significantly differs between EXP8 and the control simulation: A higher fraction of liquid precipitation during a precipitation event enhances the runoff particularly during and after the wet period, while the melting of snow (which is reduced in the modified experiment and which also influences the rate of runoff) might occur days or weeks later.
It should be stressed here that the modification, implemented in EXP8 generally, at the same time, affects only a small percentage of the entire Northern Hemisphere. Thus, the impact of the modification in the melt temperature of snow flakes might be substantially larger in a small region than discussed above.

Ratio between snowfall and precipitation

Fig. 5.47 shows how the ratio between snowfall and (total) precipitation is changed in April when assuming that the snow flakes’ melt is initialized at 1°C instead of 2°C. The most significant changes occur in a broad belt with monthly mean temperatures between $T = -5°C$ and $T = 5°C$ (thick contour lines). The northward shift of this band west of Europe due to the gulf-stream is clearly visible. In the Southern Hemisphere, the zones with large anomalies include only a few land points but cover a broad circumpolar (storm-) belt close to 60°S. Other months (not shown) depict a similar spatial distribution of these deviations on the Southern Hemisphere, whereas on the Northern Hemisphere, covered with huge land masses, strong north- and southward shifts caused by strong seasonal variations in the surface temperature are observed.

Fig. 5.48 demonstrates how the change in the fraction of snowfall (between the current experiment and the control simulation) depends on the monthly mean surface temperature. The smoothed curves can be characterized by a pronounced peak at a temperature between approximately $T = 0°C$ and $T = 2°C$, reaching a value of ~15%. There is quite large scatter between the winter months. The graph suggests that mean monthly surface temperature is a key variable for the change in the ratio between snowfall and precipitation but the fraction is affected by other processes as well. The impact of the atmospheric stratification and the intensity of the precipitation is certainly also of great relevance.
Since the fraction of snowfall depends (in the model) on the intensity of the precipitation, as shown previously, tests have been carried out to ascertain how the results differ in Fig. 5.48, when excluding weak or strong precipitation. In fact, substantial differences are found when, e.g., all events with a mean monthly precipitation of less than 20 mm were neglected. However, no structure was found in the results, indicating a rather arbitrary relationship between (monthly) precipitation intensities and the fraction of snowfall. The fact that the mean temperature during the precipitation events need not necessarily agree with the mean monthly temperatures is probably of minor relevance.

Zonal means

This section briefly discusses the zonal means in winter (DJF) of a few variables for both Eurasia and North America (Fig. 5.49). The upper two panels show that the snow cover and surface albedo anomalies do not differ significantly between Eurasia and North America. The pronounced negative peak for snow cover and surface albedo is close to 38°N and 40°N for Eurasia and North America, respectively. While the zonal width of the peaks with respect to snow cover, surface albedo and snowfall are similar, the impact on the surface temperature is noticeable in a distinctly broader latitudinal band. This indicates that the temperature changes are not restricted to areas with considerable changes in the net shortwave radiation due to advective processes.
5.9 EXP9: Dependence of surface albedo on soil moisture

Basic radiation physics suggest that water absorbs significantly more solar radiation than dry soil does. Hence, absorption of solar radiation increases with the relative fraction of water in any mixture of soil and water (Eltahir, 1998). While ECHAM4 ignores the relation between soil moisture and surface albedo, many studies propose the introduction of such a relationship. EXP9 was developed to clarify the role of soil moisture with respect to the soil moisture content.

5.9.1 Overview of modifications

EXP9 differs from the control simulation by the introduction of a relationship between bare soil albedo and soil moisture (Eq. 5.54). The equation is derived from three assumptions: (i) annual mean of background albedo in EXP9 is equal as in CTRL, (ii) the background albedo for dry soil is double the value as for wet soil, and (iii) the near-surface (relative) soil moisture closely follows the total (relative) soil moisture.

5.9.2 Relationships between bare soil albedo and soil moisture

In the following, some available relationships between the soil moisture and the background albedo are briefly described.

Idso et al. (1975) have shown that the albedo of bare soil is linearly related to the water content in the uppermost layer ($W_1$) of the soil:

$$\alpha_b = \begin{cases} 
0.31 - 0.34 \frac{W_1}{W_{1,max}} & ; \quad W_1/W_{1,max} \leq 0.5 \\
0.14 & ; \quad W_1/W_{1,max} > 0.5,
\end{cases}$$

(5.50)

where $W_{1,max}$ is the soil moisture content at saturation (field capacity) for the top soil layer.

Barker et al. (1994) additionally suggest a linear relationship between relative soil moisture and surface albedo. Their expression has been implemented in CCCII (Canadian Climate Centre second-generation GCM):

$$\alpha_b = \alpha_{dry} - 0.07 \frac{W_1}{W_{1,max}}.$$  

(5.51)

$\alpha_{dry}$ is the albedo allocated to dry conditions.

In BATS (Dickinson et al., 1993), the following nonlinear relation is proposed:

$$\alpha_b = \alpha_{sat} + 0.01(11 - 40 \frac{W_1}{\Delta z_1}).$$  

(5.52)

$\alpha_{sat}$ is the albedo at field capacity and $\Delta z_1$ the depth of the uppermost soil layer. Values of $\alpha_b$ are limited to the range $\alpha_{sat} \leq \alpha_b \leq \alpha_{dry}$ which are tabulated for eight texture classes in Dickinson et al. (1993). They assume that the following expression accurately applies to all texture classes:

$$\alpha_{dry} = 2\alpha_{wet}.$$  

(5.53)

This relationship is confirmed by Wilson and Henderson-Sellers (1985). Eq. 5.50 generates background albedos of between 0.14 and 0.31 for moist and dry soils, respectively. This range agrees well with Eq. 5.53.

The parameterization used in EM is given by Eq. 4.30.
The computation of the background albedo in EXP9 is based on (i) a linear correlation between the bare soil albedo and the relative soil moisture content, (ii) Eq. 5.53 and, (iii) identical background albedos in CTRL and EXP9 for the mean annual soil moisture content. These conditions lead to the following formulation for the background albedo of bare soil with a soil water content $W_s$:

$$
\alpha_b = \alpha_{bm} \left( \frac{2W_{max} - W_s}{2W_{max} - W_{sm}} \right),
$$

where $\alpha_{bm}$ is the background albedo as used in CTRL, and $W_{sm}$ is the average of the soil water content calculated from the 3-D 10-year control simulation of ECHAM/T42. Eq. 5.54 may only be applied to the vegetation-free part of the grid box. Unfortunately, ECHAM maintains only a single bucket for soil water. This prevents relating the bare soil albedo to the relative soil moisture of a thin subsurface soil layer which would allow for a more appropriate parameterization.

### 5.9.3 Results from the 3-D experiment

EXP9 was conducted to investigate the interaction between soil moisture and surface albedo in more detail. The evaluation focuses on regions where the annual amplitude of soil moisture content is large.

The global or North Hemispheric annual means of surface variables are not provided here since the deviations between EXP9 and the control simulation are negligible (cf. Fig. 5.51, dashed lines). This is evident, as Eq. 5.54 requires that the mean annual soil moisture content, and thus the mean annual bare soil albedo, is similar in CTRL and EXP9.

#### Comparison of soil moisture content and background albedo

In this section, the annual amplitudes of soil moisture and background albedo in EXP9 are compared by plotting differences between winter (DJF) and summer (JJA) (Fig. 5.50). The negative correlation between soil moisture and background albedo due to Eq. 5.54 is obvious. A regression between the amplitude of soil moisture and the background albedo, including all land-points, leads to a correlation coefficient $r = -0.93$ and to $\Delta \alpha_b = -0.012$ for an increase in $W_s$ by 10%.

The north- and southward shift of the ITCZ in JJA and DJF, respectively, due to varying sun elevations is mainly responsible for the soil moisture distribution in the tropics. Soil moisture is significantly higher in NH winter than in NH summer in the following areas:

- Africa south of the Equator with the rainy season in DJF
- Large parts of Brasil and Peru where the ITCZ is shifted far to the south in January
- Europe and the Rocky Mountains.

The opposite applies to South East Asia with monsoon-induced precipitation predominantly during JJA, the steppes in Africa north of the Equator, as well as Mexico and the Caribbean.

The comparison between the observed and modeled soil moisture reveals that the summer drying over western Europe is more pronounced in ECHAM4 than in the NCEP re-analysis (1957 - 1996, cf. Section 2.1) (not shown). In contrast, soil moisture amplitudes, that are larger in the NCEP re-analysis than in the model, are found in extended regions of Russia and Canada. In Africa, South America, Australia and South East Asia, the amplitudes are reproduced with reasonable skill. Hence, assuming Eq. 5.54 to be realistic, the annual cycle of the background albedo is simulated with sufficient accuracy.
Changes of annual cycles over Africa

This section describes how the monthly means of a number of variables are modified when implementing Eq. 5.54 into ECHAM4. In Fig. 5.51 the averaging domain is restricted to the steppes in Africa with the rainy season in northern summer (shown as a thick solid line in Fig. 5.50). This domain partially overlaps with the Sahel-zone between 10°N and 20°N. From January through to March, the surface albedo is approximately 0.02 higher in EXP9 than in CTRL, whereas during the wet summer season, a distinct decrease has been modeled. The decrease in the summer albedo during NH summer generates a cold temperature anomaly. This is opposite from what is expected due to the albedo-temperature feedback, i.e. a reduced surface albedo enhances the absorption of solar radiation at the surface, thereby leading to higher temperatures. Obviously, the atmospheric water vapour tends to mask this direct feedback: The atmosphere becomes more humid, and the precipitation rate as well as the relative soil moisture increase. The enhanced atmospheric water content leads to substantially lower global radiation which overcompensates the increasing absorption of solar radiation due to the lower surface albedo. The net longwave radiation increases slightly in summer. This implies that the increase in downwelling longwave radi-
Surface albedo

2-m temp.

Rel. surf. soil wetness

Net LW, surface

Vert. int. cloud water

Total precipitation

Global radiation

Net SW, surface

Net radiation

Figure 5.51: Changes in monthly (10-year) averages: EXP9 - CTRL. Solid: Africa with rainy season in northern summer (shown as thick solid line in Fig. 5.50). Dashed: Global land monthly means.

The increased soil moisture content in summer is associated with more cloud water and thus, with increased precipitation. Schär et al. (1999) report that “the surplus of precipitation over wet as compared to dry soils derives primarily from atmospheric advection”. This result is supported by a significant increase in the westerly wind component in JJA along the west coast of Africa between the equator and 15°N (not shown). This increase in the 10-m wind speed is in the order of 1 - 2 ms⁻¹, corresponding to approximately 25% of the summer mean wind speed in the control climate. The enhanced relative humidity, following Schär et al. (1999), reduces the level of free convection and thus, favors convective precipitation. In fact, convective precipitation is mainly responsible for the increase in total precipitation during summer.

These conclusions should be compared to results presented for the Sahara and the adjacent steppes in Section 5.4, based on EXP4 which incorporates the spectral albedos in the visible and near-infrared range. During summer, both experiments lead to similar results: Lower
albedo are related to lower surface temperature, primarily due to enhanced cloudiness and hence, lower global radiation.

**Comparison with observations**

This section compares the observed and simulated annual cycles of the surface albedo. As in the previous section, the averaging domain will be restricted to Africa with the rainy season in JJA. Fig. 5.52a demonstrates that the annual amplitude of the satellite-based observation and the simulated surface albedo compare well, the albedo difference between winter and summer being close to 0.04. It can thus be concluded that EXP9 captures the annual variation in surface albedo caused by changes in soil moisture, with reasonable skill. It should, however, be emphasized that the observed annual variations in the surface albedo may also be caused by changes in the vegetation cover: The wet season supports the growth of bushes and grass with typically lower surface albedos unlike desert-like areas established during the dry season. This is also suggested by Sud and Fennessy (1985) who claimed that replacing forests with bare land may increase the surface albedo by 0.06 - 0.1.

Fig. 5.52b displays the simulated and observed relative monthly soil moisture averaged over the same domain as in Fig. 5.52a. Despite substantial differences between EXP9 and the NCEP re-analysis, the model captures the evolution of the soil moisture quite well. Adapting the observed soil moisture in the model would, assuming everything else to be the same, lead to a slightly lower annual amplitude of the simulated surface albedo.

**5.10 EXP10: Incorporation of subgrid scale orography**

In the ECHAM GCM (and in most other climate models) the grid boxes consist of flat land which does not represent the reality. In rough mountainous areas, however, it is hazardous to neglect the height variation within the grid squares. The motivation for incorporating subgrid scale orography, i.e., the deviations of height within the grid square from the grid box mean, are outlined in the next paragraph.

Accurate computation of the temperature distribution within each grid box allows considerable improvement, as follows:

1. Distinction between snowfall and rain
(2) Computation of the snow covered fraction
(3) Grid fraction with snow melt
(4) More realistic snow albedo
(5) Consideration of the "seeding" effect: Snow at highest elevations encourages growth of snow cover at the beginning of the snow season (seeding). At the end of the snow season, snow melts firstly in the valleys and encourages a retreat of snow cover.
(6) Improvement of further temperature dependent processes as, e.g., evapotranspiration.
(7) More realistic runoff (and soil moisture) due to a high resolution of the topographic data.

The principal aim of this experiment is to demonstrate that (locally) substantial errors in the temperature and snow distribution can arise from the assumption of flat grid boxes.

5.10.1 Incorporation of subgrid scale topographic variability

This section describes an algorithm which allows the inclusion of highly resolved topographic data. As it is computationally highly inefficient to use the raw topographic data, a procedure must be developed to reduce the amount of data. This difficulty was overcome by using the method described in Walland and Simmonds (1996) and is briefly summarized in the following. This procedure is based on the statistical distribution theory. If the surface heights in each grid square are normally distributed, the standard deviation ($\sigma$) is sufficient to compute the height distribution about the mean. Since real distributions differ from normal distribution, deviations from normality have to be accounted for. This can best be achieved by using statistical measures, which include the skewness ($\gamma_3$) and curtosis ($\gamma_4$) of the distribution. Abramowitz and Stegun (1965) developed a very accurate approximation for calculating the percentage of heights below a certain level by using the cumulative frequency distribution ($P$). Assuming a normal distribution of unit variance for variable $x$, it can be written:

$$P_{nor}(x) = 1 - Z(x)(a_1 t + a_2 t^2 + a_3 t^3) + \epsilon \quad \text{for} \quad 0 \leq x < \infty,$$

where

$$Z(x) = \frac{1}{\sqrt{2\pi}} e^{-x^2/2}$$

$$t = \frac{1}{1 + px}$$

and

$$p = 0.33267, \ a_1 = 0.43618, \ a_2 = -0.12016, \ a_3 = 0.93729, \ |\epsilon| < 10^{-5}.$$  \hspace{1cm} (5.58)

To account for the deviation from normality of the distribution, the skewness and curtosis are used

$$P(x) = P_{nor}(x) - \left[ \frac{\gamma_3}{6} Z^{(3)}(x) \right] + \left[ \frac{\gamma_4}{24} Z^{(3)}(x) + \frac{\gamma_3^2}{72} Z^{(6)}(x) \right],$$

where subscripts in parentheses denote order of differentiation. To compute the various statistical moments, a high-resolution topographic dataset was needed. The best dataset available so far covers the whole world in about a km x km grid, and is described in the following section.

5.10.2 Topographic dataset

The dataset used to compute the first four statistical moments of the subgrid scale orography is based on a global digital elevation model, compiled by the U.S. Geological Survey's
EROS data center in Sioux Falls, South Dakota (Bliss and Olson, 1996). The horizontal grid spacing is 30-arc-seconds (which corresponds to about 1 km) resulting in a global matrix having a dimension of 21,600 rows and 43,200 columns. The size of the global dataset is about 1.7 GB (ASCII). The elevation values range from -407 to 8752 metres. The dataset is derived from 8 sources of elevation information, the two most important being:

1. Digital Terrain Elevation Data (DTED) with a horizontal grid spacing of 3-arc-seconds (approximately 90 m) produced by the National Imagery and Mapping Agency (NIMA). It was used for most of Eurasia and large parts of Africa.
2. Digital Chart of the World (DWC), based on the 1:1000,000-scale Operational Navigation Chart (ONC) series, which is the largest scale base map source with global coverage (Danko, 1992). It was used for Australia, most of Greenland, and large areas of Africa, South America, and Canada.

Mean average, standard deviation, skewness and curtosis of the subgrid scale topography are calculated for the T42 resolution. These statistical moments are used as new surface boundaries in EXP10. Fig. 5.53 illustrates the height distribution for the T42 resolution. Fig. 5.53a illustrates the standard deviation ($\sigma_z$) of the subgrid scale orography. The computation is performed for the T42-mesh and is based on the 30-arc-seconds topographic dataset, described above. High values of $\sigma_z$ are associated with rough mountainous regions. It may be of some interest that the recalculated $\sigma_z$, based on the 30-arc-seconds global elevation datasets, yields significantly higher values than those assumed in ECHAM4 (not shown), the only reason being the lower resolution (5' x 5') of the US Navy dataset on which the former computation of $\sigma_z$ in ECHAM4 is based.

Fig. 5.53b displays the percentage of each grid square with an altitude of more than 300 metres above the mean grid height. 300 metres correspond, assuming a moist adiabatic temperature gradient, to a temperature change of approximately two degrees. Largest values are found over mountainous regions and exceed 20% in extended regions of the Himalaya mountains, the Rockies and the Andes. The relatively steep slopes at the ice-sheet edges of Greenland and Antarctica as well as the Alps can also be identified. 15% of the land grid boxes have such a height distribution that a quarter (or more) of the total grid area exceeds its respective mean height by more than 300 metres. The corresponding values for 200 m and 100 m are 27.7% and 49.6%, respectively.

The skewness reveals some properties of the height distribution within the grid cells. 71% of the grid boxes (T42 resolution) have a positive curtosis. This signifies that the median is lower than the mean or, more descriptively, that the grid box contains many low-lying regions and comparatively few very high peaks.

5.10.3 Experimental setup

It is beyond the scope of this study to adapt all physical processes to the modified height distribution. In the following, the modifications which are incorporated in EXP10, are briefly summarized.

1. Calculation of the temperature distribution within each grid box and each time step:
   - Based on the simulated mean grid temperature, the percentages of the grid box in 2°C-intervals are determined. Higher resolution (0.5°C) is provided for the "critical" range between -2°C and 2°C for a better representation of physical processes which rapidly change close to freezing point.
   - The lapse rate in the lower atmosphere is computed using the 2-m temperature and the temperature on the third atmospheric model level, corresponding to a height of approxi-
Figure 5.53: a) Global distribution of standard deviation of the subgrid scale topography derived from a high-resolution (30-arc-seconds) elevation dataset (T42 resolution). b) Fraction of each grid square that is more than 300 m higher than its mean height.

approximately 400 m above the ground.

(2) The distinction between rain and snowfall was considerably modified. The percentage of snowfall to total precipitation is determined using the rain-snow criterion, as derived from Rohrer (1992) (see Fig. 5.46). This would lead in some regions to a significant increase in the ratio between snowfall and rain (cf. Section 5.8). In order to avoid undue changes on the ratio between snow and rain, the curve for the rain-snow criterion, as suggested by Rohrer (1992), was shifted to maintain the ratio between solid precipitation and rain.

(3) Each grid element is divided into a domain above and below freezing point. For each domain a separate snow cover is maintained. Therefore, at the beginning and the end of the snow season it may happen that the lower part is snow free while the upper region is snow covered. As the freezing level deviates, snow is distributed between the two packs, conserving mass at all times. To avoid numerical instabilities and strongly diverging snow
depths between the two domains, the snow water equivalent in the lower and upper domain approach each other in an exponential form, the time constant being 14 days.

(4) Snow melt processes, as well as runoff and interception, are calculated for the domains above and below freezing point, using its respective mean temperature. To reduce the complexity of the modifications, the calculation of the turbulent heat fluxes as well as the specific humidity are not adapted.

(5) The mean grid square surface albedo is based on an area-weighted mean, thereby assuming that the snow water equivalent is constant in the upper and lower domain. The snow albedo is computed using the temperature distribution from (1) and the relation between temperature and snow albedo (Eq. 5.18 as implemented in EXP3). The net shortwave radiation is determined using the global radiation and the mean grid box albedo. The upward longwave radiation is also determined using the mean grid box surface temperature since the deviation from a computation using the accurate temperature distribution is marginal.

To minimize computational cost, the aforementioned modifications were only applied to grid elements with \( \sigma_z > 150 \text{ ms}^{-1} \).

5.10.4 Results from the 3-D experiment

EXP10 was developed to investigate the impact of the irregular height distribution within each grid box on the surface climate. On temporal and spacial smaller scales, EXP10 and CTRL differ significantly whereas the impact of the modifications on the global mean climate is negligibly small. The structure of EXP10 allows a detailed analysis of the climate within a grid box. Therefore, the analysis is focused on the mean state of the lower (with surface temperatures above freezing point) and the upper domain (with surface temperatures below freezing point) in the grid boxes.

Temperature distribution

In each model grid box and at each time step, the fraction of the grid box above and below freezing point was computed. In a second step, the mean temperature of the upper \((T_1)\) and lower domain \((T_2)\) was determined using the simulated temperature gradient within the atmosphere between the surface and a height of approximately 400 m above the ground. Whenever the entire grid element is above or below 0°C, \(T_1\) and \(T_2\) are equal. Fig. 5.54a illustrates the difference \(T_1 - T_2\) for the month of October. Large positive values are found over the Himalayas, excluding the Tibetan Plateau, the Caucasus and the Alps. In Scandinavia and in the Rockies, the differences are usually distinctly smaller. The regions with large deviations between the mean temperature of the upper and lower domain are mainly concentrated on rough mountainous areas (cf. Fig. 5.53) where the mean grid square temperatures do not greatly deviate from \(T = 0°C\). Flat areas with \(\sigma_z \leq 150 \text{ ms}^{-1}\) are not separated into a lower and upper domain and thus \(T_1 = T_2\) applies. The more frequently the entire grid box is above or below freezing level, the lower the difference \(T_1 - T_2\). Fig. 5.54b displays the fraction of the grid boxes with mean surface temperatures below 0°C. Approximately 10% of the Alpine area is below 0°C in October, leading mainly to solid precipitation. The control experiment, in contrast, simulates marginal snowfall over the Alps in October. For all completely flat grid squares, Fig. 5.54b corresponds to the frequency of simulated surface temperatures below 0°C. Assuming normal distribution for temperatures, the 50%-contour line coincides with the temperature, that the percentages represent. This is well confirmed in Fig. 5.54b (thick lines depict the 0°C isoline).
Assessment of the subgrid scale conditions of snow

As previously described, each grid square is subdivided into two domains, comprising the range above and below freezing level. For each domain, a specific snow regime is maintained. Based on the temperature distribution, a separate calculation for snowfall, snow water equivalent ($S_{n1}$, $S_{n2}$), snow cover fraction ($f_{s1}$, $f_{s2}$), snow albedo and snow melt is realized. Fig. 5.55 displays how the snow regime develops in November. This month is ideal to show the relevance of the subgrid scale orography in early winter. In the mountainous regions with mean monthly temperatures below $T = -5^\circ C$ (isolinc drawn as thick line), the snow cover fraction of the two domains differ by up to 25%, e.g. in the Alps or Caucasus mountains. In cold regions with mean temperatures below approximately $-5^\circ C$, the differences vanish, since the entire grid box is mostly below the freezing level. A number of areas in Russia, Siberia and Canada are rather flat and being below the threshold of $\sigma_z = 150$ m s$^{-1}$ which is required to maintain the distinction between two snow packs within each grid element. The difference $f_{s1} - f_{s2}$ over the Tibetan Plateau approaches zero due to low temperatures and relatively flat grid squares.

Differences between EXP10 and the control simulation

The autumn is ideal to investigate the impact of the modified model simulation on the climate. During the winter, the influence of the modifications is reduced and the large-scale circulation dominates the development, probably more than local modifications in mountainous areas.

Fig. 5.56 demonstrates that differences in the surface climate between EXP10 and the control simulation are typically small. In autumn, significant deviations are restricted to mountainous areas. The distinct increase in the snow cover fraction over the Alps, the Caucasus, the Rockies and parts of the Himalayas is likely to be caused by snowfall in the
grid box domain below the freezing level. This increased snow cover can be interpreted as seeding of snow cover on the "peaks". In the ECHAM4 control experiment, snowfall rarely occurs in autumn in the Alps or the Caucasus in autumn, and potential snow usually melts within a short time period. In the modified model version, however, snow in the upper reservoir remains for a longer time due to temperatures below 0°C. As seen in Fig. 5.55, the difference in the snow water equivalent in the Himalayan area shows the following pattern: the Tibetan Plateau tends to have less snow whereas in the surrounding areas, the snow water equivalent increases.

**Monthly surface albedo in the Alps**

The Alps are defined here as the domain ranging from 45.5°N to 48.0°N and 5°E to 15°E, respectively, covering an area of ~ 2.1 \times 10^5 km^2, which is approximately five times the extent of Switzerland.

Fig. 5.57 displays the annual cycles of surface albedo over the Alps from four selected sources. While the surface albedo in the control simulation remains essentially constant in September and October, the albedo in the modified model run increases significantly from September to October. As previously mentioned, this increase is primarily caused by snowfall in the domains with below freezing temperatures. In the control simulation, in contrast, precipitation usually falls as rain in September and October, and possibly fallen snow would rapidly melt. Hence, the Swiss winter starts earlier in EXP10 than in the control simulation. Compared to the control simulation, the mean monthly surface albedo is considerably greater in EXP10, mainly in autumn and spring. Obviously, the model
Figure 5.56: SON-differences: EXP10 minus control simulation. a) Snow cover fraction, b) surface albedo, c) snowfall rate.

captures the seeding effect in autumn in an appropriate manner but the representation of the unseeding effect in spring fails.

A comparison between the simulated albedos and the mean monthly surface albedos, which are kindly provided by L. Zgraggen (pers. comm.), indicates that EXP10 is in better agreement with the observation. The monthly mean albedos are reasonably well simulated in EXP10. The significant differences between the model and the observation by L. Zgraggen is likely due to differing domains, on which the averages are based: The percentage of mountains in Switzerland is higher than in the Alpine region, as defined above.

EXP10 simulates snow on the highest peaks in the Alps even in July and August. This seems realistic as perpetual snow typically occurs in regions higher than 3000 metres above sea level.

Summarized, it can be concluded that the implementation of the subgrid scale topography significantly improves the evolution of the snow pack over the Alps. However, it should be mentioned that the Alps are a region actually too small for reasonable GCM comparisons,
Figure 5.57: Monthly mean surface albedo as simulated in the control run (CTRL) and EXP10, as well as derived from the surface radiation budget (SRB) and an observational estimate, kindly provided by L. Zraggen (pers. comm.).

as the Alps comprise a domain of three grid boxes only at T42 resolution.

Summary

In orographically rough areas, substantial differences between the control simulation and EXP10 occur primarily in autumn (Northern Hemisphere) while in winter and spring, no systematic differences between EXP10 and the control simulation are found. The evolution of the surface albedo may be significantly improved in isolated mountainous areas, such as the Alpine region. Obviously, modifications (e.g. snowfall rate) in isolated domains during short periods are effectively weakened and balanced in 3-D simulations by advective processes.

The implementation of the subgrid scale topography in models with very low resolution is likely to yield markedly larger impacts on the surface climate (Walland and Simmonds, 1996). These modifications allow, even in low-resolution models, a simulation of realistic snowfall, snow depth and surface albedo. Furthermore, the incorporation of the subgrid topography allows for the investigation of local climate conditions in mountainous areas without nesting a regional model in a GCM.
6. Conclusions

This study focused on the representation of land surface processes in climate models. Special emphasis was placed on the surface albedo and snow cover. The two main objectives of this study were: (i) the assessment and improvement of the surface albedo parameterization in ECHAM with focus on snow covered conditions, and (ii) a thorough comparison of the land surface schemes in ECHAM and EM, and the response to variations in key surface parameters. In order to achieve these objectives, both off-line simulations and long-term integrations with the three-dimensional ECHAM GCM were conducted.

The main area of the present work consisted of the evaluation of available surface albedo and the development of improved parameterizations for this parameter. Furthermore, testing a number of parameterizations in the framework of three-dimensional model simulations was also of major interest. The evaluation of ten long-term ECHAM4 climate simulations revealed several deficiencies in the oversimplified albedo parameterization in the ECHAM GCM.

The impact of modifications related to surface albedo on the global climate was of different magnitude. It is thus important to rank the 3-D model simulations with respect to their response on the (near-surface) climate.

The correct conversion of simulated snow water equivalent into snow cover fraction is of prime importance for a correct simulation of snow cover fraction, and thus surface albedo, radiation balance and temperature. Hence, the simulated snow cover fraction over both flat and mountainous areas were validated. The analysis revealed that the snow cover fraction is underestimated in flat areas but overestimated over mountainous regions. Two 3-D-experiment (EXP1 and EXP2) were conducted which demonstrated a large impact on the surface climate when using a modified representation of snow cover fraction. The main problem of parameterizing the correct snow cover fraction in GCMs arises from different model resolutions and from the existence of relatively few global datasets for both snow cover fraction and snow water equivalent.

A separate treatment of radiation fluxes in the visible and near-infrared spectrum (EXP3) as well as the implementation of the snow albedo parameterization as used in BATS and CLASS (including a simple snow interception model), affects the climate primarily on a local and continental scale. Evidence is presented that the introduction of a prognostic snow-aging variable improves the simulation of the snow albedo. It is insufficient to describe snow aging solely with the current surface or screen temperature as this ignores the "memory" of the snow as well as snow pollution. BATS illustrates the importance of accurately determining the key-parameters and the difficulty to develop a parameterization valid for all climate regimes. The parameterization, as used in CLASS, is suitable for use in GCMs to describe the albedo of snow-covered forests. The largest impacts on the surface albedo are found over the boreal forests in Siberia and Canada, where the zonal averaged albedo decreases by up to 0.1 in winter. This leads to snow-covered forest albedos which are in better agreement with the literature. The subsequent significant rise in surface temperature over extended parts of Eurasia and North America yields an earlier snowmelt in spring and an accelerated retreat of the snow line. The development of a simple model for snow intercepted on the canopy proved that it is feasible to simulate extremely local processes in the framework of a GCM. It is somewhat questionable to introduce a new prognostic variable for the snow mass intercepted on the canopy as it is difficult to validate this variable on a common GCM grid. However, the simple interception model provides valuable information on the regional magnitudes of processes affecting the canopy snow reservoir.
An insignificant impact on the surface climate, both on a global and continental scale, is inherent to the model experiments which incorporate: (i) the polynomial relationship between snow albedo and surface temperature, (ii) a reduction in the melting temperature of snow flakes in the atmosphere and (iii) the subgrid orography to account for temperature and snow cover within a GCM grid square.

In the following, some suggestions are proposed how to improve ECHAM4 regarding the representation of processes related to the surface albedo. As an improvement to ECHAM4, the specific parameterization concerning the surface albedo could be combined as follows: The two relationships between snow cover fraction and snow water equivalent adapted for flat and mountainous regions could be merged into one single equation which contains both the \( \tanh \) function and the square root term containing the subgrid orography. Since the parameterization, as presented in EXP1, is primarily derived for forest-free grid boxes, it is strongly suggested to compute the snow cover fraction over forests according to CLASS (EXP7). Since the leaf area is essential for calculating transpiration rates and thus soil moisture and possibly precipitation, it is recommended to introduce the seasonal variations of the leaf area index (EXP6). Furthermore, the leaf area index plays a major role in specifying the sky view factor as used in CLASS. BATS is best qualified to represent the snow albedo by introducing a snow age factor. However, accurate determination of model parameters is required. The description of the snow albedo should not be given by a (polynomial or other) function of the surface temperature. A further modification which considerably improves the quality of the albedo parameterization in ECHAM4 consists of the separate handling of the radiation fluxes in the visible and near-infrared spectrum (EXP4). The question arises as to how other modifications can be adapted to allow for a correct distinction between the visible range and NIR. The snow cover fraction depends on, for example, the wavelength of the incoming radiation, as its absorption within the snow cover is substantially larger in the NIR than in the visible band. However, in a preliminary version of the new model, these effects could be neglected. Over bare soil it is profitable to account for albedo changes due to variations in the soil moisture (EXP9). For a correct simulation of the near-surface soil moisture which is responsible for albedo changes, it would be beneficial to replace the bucket model in ECHAM by a multi-layer soil model. Despite little impact on the simulated surface climate when using T42 resolution, it is suggested to implement a sub-model to account for temperature variation within single grid boxes (EXP10) as this allows for a more physical representation of snow cover conditions, especially for coarse resolutions (e.g. T21). Since this additional sub-model requires rather large amounts of computer time, it is unreasonable to incorporate these modifications in a high-resolution GCM.

A serious problem found in many GCMs is the delayed snow melt in spring. Comparisons of the control climate with observations revealed that a positive feedback might postpone the snowmelt as an excess of snow enhances the surface albedo and thus reduces the surface temperature. A higher surface temperature favours an earlier snowmelt directly and also indirectly by enlarging the frequency of rain. Furthermore, excessive precipitation in the northern parts of Eurasia and North America during April could enhance the positive snow bias. The 3-D experiments revealed that the reduction of the albedo over snow covered (boreal) forests is probably the most effective manner for improving the model with respect to the delayed snow melt. Validation of off-line experiments with observational data from Russian sites showed, however, that excessive snow in spring is also simulated in off-line experiments. Different reasons may lead to these results: (i) snow only melts in ECHAM4 when both the snow pack and the relatively thick uppermost soil layer exceed freezing point, (ii) modeled surface temperature being less than or equal to 0°C as long as snow patches exist and (iii) an erroneous value for downwelling longwave radiation. Issues (i) and (ii) both contribute to a delayed snow melt.
The main results that are based on off-line simulations and sensitivity tests, using the land surface schemes incorporated in ECHAM and EM yielded the subsequent results. Despite the different structure of the land surface schemes used in ECHAM and EM, the simulated surface climate is, in part, similar. The most pronounced differences, based on off-line model experiments, are found in the simulated annual cycles of surface runoff and drainage, as well as in the transpiration and bare soil evaporation due to the different soil model (bucket and non-bucket type). However, total monthly runoff and latent heat flux do not differ substantially. The diurnal amplitude of the ground heat flux is significantly overestimated in both models (EM showing slightly better results) due to the absence of a very thin uppermost soil layer which rapidly responds to changes in the net radiation flux. The key problems in the EM land surface scheme are due to the forced deep soil temperature, which yields an unrealistic simulation of the monthly ground heat fluxes. The oversimplified transpiration algorithms imply an increase that is too rapid in the transpiration rate in late spring. The EM soil model, which includes several layers for water, is better adapted to capture the correct origin of evaporation and runoff processes, whereas the bucket formulation in ECHAM requires the determination of a number of parameters to account for this problem. The simulation, including soil and plant parameters adapted for the Cabauw site leads to improved soil moisture. Turbulent heat fluxes, in contrast, show less agreement with the observations when using the adapted parameter set. This study found a significant impact of the von-Kármán-constant and the stability-correction function on the simulated monthly values in the turbulent heat fluxes. Under snow covered conditions, the two land surface schemes generate substantial deviations in the shortwave net radiation, primarily due to the different assessment of the snow cover fraction. The sensitivity studies of the land surface schemes, as used in ECHAM and EM, have clearly shown that it is insufficient to consider monthly values only, but that it is necessary to extend the analyses to diurnal cycles. As the sensitivities are mostly non-linear, it is of major importance to study the impacts over the entire parameter range. The response on surface climate usually levels off with increasing roughness length, leaf area index and field capacity. The ECHAM land surface scheme generally shows a higher sensitivity with respect to the leaf area index and roughness length due primarily to a different parameterization of the transpiration rate.

A number of issues need further investigation:

The “stand-alone” forcing used to examine the sensitivity of a land surface scheme is a useful tool for investigating the structure and general behaviour of land surface schemes, as well as for comparing them to each other. However, since the surface-atmospheric feedbacks are neglected in off-line simulations, 3-D model experiments should also be conducted to clarify the principal impact of each parameter on the climate in the framework of a fully developed GCM. The sensitivities are largely dependent on the vegetation type and the climate regime. It would, therefore, be beneficial to extend the studies to other climate regimes, such as tropical rain forests, boreal forests and more arid climates. The modified (snow) albedo submodels are primarily validated in an off-line mode due to the lack of accurate global climatologies of surface albedos. Thus, in 3-D climate simulations, only strong anomalies in the albedo pattern could be revealed when compared to observations. Therefore, the climate research community should, in addition to model development, focus on the compilation of accurate global surface climatologies.

The sea surface temperature (SST) and sea-ice are prescribed in all 3-D climate simulations with the ECHAM GCM. The sensitivity of the climate to surface parameters is likely to be larger with interactive oceans. Furthermore, the impact of albedo variations during winter and spring would also possibly affect summer air temperatures as the oceans have a long “memory”. It would thus be beneficial to expand the investigations to coupled ocean-atmosphere climate models.
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Appendix

A.1 Abstract and Summary of Paper 1

Abstract
Annual cycles of monthly albedos simulated with a General Circulation Model (GCM) are compared with surface observations. The data observed at 35 stations are retrieved from the Global Energy Balance Archive (GEBA) and drawn from the soil moisture and meteorological observations in the former Soviet Union. The model data are obtained with the ECHAM4 GCM in a ten-year simulation of the present-day climate at T106 resolution. The model calculated albedo values are modified before they are compared with the surface observations: They are interpolated to the stations and adjusted to account for altitude differences and fractional forest area. During the snow-free period, the model underestimates the albedo by up to 0.05 at the stations (with values between 0.2 and 0.25 measured over short grass) because the albedo for grassland is too low in the model. During the period with seasonal snow cover, the model underestimates the albedo by up to 0.2 at stations in Russia, Scandinavia and Canada, which experience severe winters. This underestimation is due to an oversimplified parameterization of the snow covered grid fraction and an inadequate linear relation between snow albedo and temperature. The derivative of albedo with respect to the forest fraction implemented in ECHAM is in line with the observations, although a small overestimation of the model's gradient has been detected.

Summary
Surface albedos calculated with the ECHAM4 GCM at T106 spectral resolution have been compared with observed data from the Global Energy Balance Archive (GEBA) and long term observations at six Russian sites.

A direct comparison is not possible because the model albedo is representative of a grid cell whereas the observations are point measurements. A comparison becomes feasible, however, if three modifications are applied to the model calculated values. These modifications account for (i) the distance between the station and the grid points, (ii) the difference in altitude between the station and the grid cell and (iii) the difference in vegetation at the station (usually short grass) and in the grid cell (usually a mixture of evergreen and deciduous forests with cultivated land) as a whole.

The ECHAM4 albedo is distinctly lower than that observed during the snow-free period for most sites in North America, Europe and the former Soviet Union. The modelled albedo is approximately 0.15, whereas the observations suggest values in the range of 0.20 - 0.25. The main reason for this is the lower albedo for forest as opposed to grass. However, the observed albedos over grass are, partially, significantly larger than the grass albedo which have been allocated in ECHAM4 to grassland (Claussen et al., 1994).

The model albedo is about 0.1 - 0.2 lower than the observed albedo over snow. It has
been found that the simulation of the snow water equivalent is realistic and thus, not the main reason for these differences, but rather the parameterization of the snow covered grid fraction and the temperature dependence of the pure snow albedo. It is very likely that ECHAM underestimates the snow covered grid fraction and that the temperature-albedo relation of snow needs improvements. Different data sources suggest a more polynomial relation as opposed to a linear one. The value of the maximum albedo of pure snow in the model, which equates to 0.8, seems to be a reasonable choice; however, over ice and snow covered ice, the GEBA observations suggest a higher value of about 0.85.

The derivative of the total grid-averaged model albedo with respect to the forest fraction has been confirmed by the observation. However, a slight overestimation of the model's gradient has been observed.
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