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**BOREAL LAND SURFACE WATER AND
HEAT BALANCE**

**MODELLING SOIL-SNOW-VEGETATION-
ATMOSPHERE BEHAVIOUR**

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“As models become more complicated, their behaviour becomes more and more unpredictable, much like reality.”

Tennekes (1987) as cited by Goudriaan (1989)

PREFACE AND ACKNOWLEDGMENTS

Strange enough, my interest in snow and winter research was awakened in the sunny summer of 1997, when I as a Master student was set to re-program an old computer model. Since then, I have also experienced all the exiting aspects of making measurements in the field: sunny days when you have to ski to work, animals who like to eat your cables, not to mention tractors who pull them out of the ground. The sub-title of this thesis should not be misunderstood. For me, modelling is a dualistic approach combining observations and predictions to understand the behaviour of nature.

The thesis work has been carried out both at the Department of Soil Sciences, Swedish University of Agricultural Sciences (SLU), Uppsala and at the Department of Land and Water Resources Engineering, Royal Institute of Technology (KTH), Stockholm. There are many people who have supported my over the years.

First of all I am grateful to my supervisors Per-Erik Jansson, Department of Land and Water Resources Engineering, KTH, Lisbet Lewan, Department of Soil Sciences, SLU, Manfred Stähli, Eidg. Forschungsanstalt für Wald, Schnee und Landschaft, Switzerland, and Lars-Charister Lundin, Department of Geosciences, Uppsala University, for your guidance and encouragement, both as colleagues and as friends.

Secondly, I would like to thank all my colleagues at KTH and SLU for an enjoyable working environment, both in the field and at the office.

The financial support from the Swedish Research Council is gratefully acknowledged.

Finally I would like to thank all my friends and family, who have encouraged me along the way, and posed the really critical question: "What are you really doing?"

David Gustafsson
Stockholm 2001

ABSTRACT

The water and heat exchange in the soil-snow-vegetation-atmosphere system was studied in order to improve the quantitative knowledge of land surface processes. In this study, numerical simulation models and available datasets representing arable land, sub-alpine snowpack, and boreal forest were evaluated at both diurnal and seasonal timescales.

Surface heat fluxes, snow depth, soil temperatures and meteorological conditions were measured at an agricultural field in central Sweden during three winters and two summers from 1997 to 2000 within the WINTEX project. A one-dimensional simulation model (COUP) was used to simulate the water and heat balance of the field. Comparison of simulated and measured heat fluxes in winter showed that parameter values governing the upper boundary condition were more important for explaining measured fluxes than the formulation of the internal mass and heat balance of the snow cover. The assumption of steady state heat exchange between the surface and the reference height was inadequate during stable atmospheric conditions. Independent estimates of the soil heat and water balance together with the comparison of simulated and measured surface heat fluxes showed that the eddy-correlation estimates of latent heat fluxes from the arable field were on average 40 % too low.

The ability of a multi-layered snowpack model (SNTHERM) to simulate the layered nature of a sub-alpine snowpack was evaluated based on a dataset from Switzerland. The model simulated the seasonal development of snow depth and density with high accuracy. However, the models ability to reproduce the strong observed snowpack layering was limited by the neglect of the effect of snow microstructure on snow settling, and a poor representation of water redistribution within the snowpack.

The representation of boreal forest in the land surface scheme used within a weather forecast (ECMWF) model was tested with a three-year dataset from the NOPEX forest site in central Sweden. The new formulation with separate energy balances for vegetation and the soil/snow beneath the tree cover improved the simulation of seasonal and diurnal variations in latent and sensible heat flux. Further improvements of simulated latent heat fluxes were obtained when seasonal variation in vegetation properties was introduced. Application of the COUP model with the same dataset showed that simulation of evaporation from intercepted snow contributed to a better agreement with the measured sensible heat flux above forests, but also indicated that the measurements might have underestimated latent heat flux. The winter sensible heat flux above the forest was further improved if an upper limit of the aerodynamic resistance of 500 s m^{-1} was applied for stable conditions.

A comparison of the water and heat balance of arable land and forest confirmed the general knowledge of the differences between these two surface types. The forest contributed with considerably more sensible heat flux to the atmosphere than the arable land in spring and summer due to the lower albedo and relatively less latent heat flux. Latent heat flux from the forest was higher in winter due to the evaporation of intercepted snow and rain. The net radiation absorbed by the forest was 60 % higher than that absorbed by the arable land, due to the lower surface albedo in winter.

Key words: soil; snow; land surface heat exchange; forest; arable land; eddy-correlation.

LIST OF PAPERS

This thesis is based on the following papers that are referred to by their Roman numerals:

- I. Gustafsson D, Stähli M, Jansson P-E (2001) The surface energy balance of a snow cover: comparing measurements to two different simulation models. *Theoretical and Applied Climatology*, 70: 81-96.
- II. Gustafsson D, Lewan E, van den Hurk BJJM, Viterbo P, Grelle A, Lindroth A, Cienciala E, Mölder M, Halldin S, Lundin L-C, (2002) Boreal-forest surface parameterisation in the ECMWF model – 1D test with NOPEX long-term data. *Journal of Applied Meteorology*, in press.
- III. Gustafsson D, Waldner P, Stähli M (2002) Factors governing the formation and persistence of layers in a sub-alpine snowpack. Submitted to *Hydrological Processes* June 2002.
- IV. Gustafsson D (2002) Surface heat and water balance of a boreal arable field - long-term measurements and simulations. Manuscript.
- V. Gustafsson D, Jansson P-E, Lewan E (2002) Water and heat balance of the boreal landscape - comparison of forest and arable land. Manuscript.

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INTRODUCTION

Land surface and vegetation are important components of the climate system. The land surface influences the input of water and heat to the atmosphere, as well as long-term variations of atmospheric chemistry (Bonan et al. 1995). In turn, the climate influences the distribution of the vegetation on the surface of the Earth. The climate variations seen in the past have been associated with large migrations of the terrestrial ecosystems, and it has been shown that the interaction between climate and forest retreat in high northern latitudes may have a significant role for the onset of ice ages (Gallimore and Kutzbach 1996). The relationship between climate and the properties of different land surfaces has long been recognized. The climatic effects of deforestation were discussed already in the late 1700s (Feldman 1992). Today, much attention is paid to the response of the terrestrial ecosystems to a possible climate change and increased atmospheric CO₂ caused by human activities.

The presence of snow greatly alters the energy balance and hydrology of the land surface due to its high reflection of solar radiation (albedo) and temporal storage of water. Snow also acts as a thermal insulator of the soil, which reduces the soil freezing. The large forest areas in the boreal¹ zone have a significant impact in this context. It has been shown that these forests contribute to a warmer and more humid climate compared with open areas due to higher absorption of radiation and more evaporation (Bonan et al. 1995; Harding and Pomeroy 1996). The winter

albedo of a boreal forest is typically lower than that of an open snow-covered area since the trees shade the high albedo of the snow.

Adequate representations of the physical processes governing the energy exchange between the atmosphere and the terrestrial ecosystems are thus a key to understanding the climate system. Over the years large efforts have been devoted to the development of numerical models for meteorological and hydrological predictions with respect to different temporal and spatial scales. Each application may have specific demands for the description of the land surface-atmosphere processes to ensure sufficient accuracy of predicted fluxes. Models for the soil-vegetation-atmosphere-transfer of water and heat are often referred to as SVAT models. For a numerical weather prediction model, the objective of the SVAT model is to provide correct surface fluxes of heat, moisture, and momentum at an hourly timescale, since these constitute the lower boundary of the atmospheric processes (Viterbo 1996). In a large-scale hydrological model it may be most important to estimate the evaporation and the snowmelt rates (Bergström 1996) on a daily timescale. Predictions of snow microstructure are essential for avalanche risk assessment, and are sensitive to spatial and temporal variations of surface temperature and water and heat fluxes within the snowpack (Colbeck 1980; Colbeck 1989; Brun 1989). The seasonal variation of the soil physical microclimate may be most important for studies of vegetation growth and the turnover of carbon and nitrogen. Recent studies indicate that biological activity related to thawing of frozen soils greatly

¹ Boreas, the God of the north wind

enhances the emissions of greenhouse gases from agricultural and forest soils. Nevertheless, the number of details included is not a guarantee for the success of a model; compensating errors in different model parts or inadequate representations of single processes may be difficult to distinguish (Pomeroy et al. 1998).

It has been shown that climate predictions are highly sensitive to the accuracy of the estimated land surface hydrology and energy balance. A poor representation of the surface processes can force the predicted climate into an unrealistic state, for example predictions of precipitation regimes are very sensitivity to the estimated surface evaporation (McGuffie and Henderson-Sellers 1997). To ensure that correct descriptions are made of the surface characteristics and the corresponding sensible and latent heat fluxes, it is necessary to evaluate the behaviour of land surface models without the atmospheric feedback, with respect to the long-term effect on the heat and moisture budget (Viterbo and Beljaars 1995). In most climate models, the surface of the earth is divided into grid cells of tenths to hundreds of kilometres size. It is important to evaluate not only the composite behaviour of the mixed surface, but also the ability to represent the individual surface types (Verseghy 2000). To represent the average behaviour of a heterogeneous grid surface many models uses different surface types, and calculates the average surface fluxes by fractional area weighting (Koster and Suarez 1992).

Scientific problems

Knowledge about the land surface-atmosphere interactions at high northern latitudes is fairly general. It is well known

that evaporation from intercepted snow may reduce the total amount of snow accumulating in forests compared to open areas (Harding and Pomeroy 1996; Koivusalo and Kokkonen 2002). Transpiration is regulated by leaf stomata closure, as governed by radiation, atmospheric vapour pressure, soil moisture conditions and soil temperature (Jarvis 1976), whereas evaporation from wet surfaces is controlled merely by the available energy and transport. The variation of aerodynamic properties in time and space is fairly well understood for boreal forest and open areas (Garratt 1993; Mölder and Lindroth 1999).

However, there are still considerable uncertainties in the representation of these processes in SVAT models when applied to specific land surface types. One problem is related to the appropriate approximation level needed to describe the desired phenomena and another uncertainty is related to the selection of parameter values. The ideal situation would be that parameter values represent properties of the system, which may be estimated from independent observations. However, the selection of parameter values and model concepts often depends on the specific purpose of the model application, and may not always be related to a measurable physical property.

There are also several specific processes where an improved fundamental understanding of governing mechanisms would be helpful. For instance, the parameterizations of turbulent heat exchange over melting snow (Pomeroy et al. 1998), methods for modelling interception and evaporation of snow in forest canopies (Lundberg and Halldin 2001), and inter-annual variations in tree

transpiration correlated to harsh winter conditions (Mellander 2001).

The need for long-term datasets for evaluation of SVAT models has been addressed. The number of available datasets is increasing, since several land surface experiments have been initiated over the last decade. However, Only recently have long continuous data been made available from the boreal zone in northern high latitudes (for instance BOREAS, Canada, Sellers et al. 1997; NOPEX, Scandinavia, Halldin et al. 1998) The present study was done within the WINTEX framework, the winter extension of the NOPEX project devoted to the land surface processes of the boreal forest zone in northern Europe.

Objectives

The overall objective of this thesis was to improve understanding of the governing processes at a local scale, contributing with quantitative knowledge that might be useful also on larger scales. More specifically the objectives of the individual papers were to:

- Evaluate the importance of the formulation of internal snow properties in mathematical models for the estimation of surface fluxes (I),
- Evaluate the representation of boreal forest in a SVAT scheme used within a global circulation model at the European Center for Medium Range

Weather Forecasts (ECMWF) (Viterbo and Beljaars 1995; van den Hurk et al. 2000) with measurements from the NOPEX forest site on Norunda Common (II),

- Investigate the effect of strongly varying surface conditions on the seasonal development of snow pack layering (III),
- Quantitatively analyse the water and heat balance processes of an arable field on a seasonal and diurnal time scale (IV), and
- Compare the water and heat balance processes of forest and arable land (V).

Limitations

The spatial scale represented by the available observations sets the limit for the interpretation of the model applications. The scale of the observations ranged from a few centimetres (soil moisture and temperature) to hundreds of metres or several kilometres for the measurements of the surface heat fluxes.

Detailed observations of the surface energy balance components have been used to develop and evaluate SVAT models, forced by measured near surface atmospheric conditions. Thus, the feedback from the surface to the atmospheric conditions have not been addressed.

THEORETICAL BACKGROUND

The fundamental theoretical background for land-surface atmosphere interactions is given in this chapter. Specific processes concerning snow cover and the interaction

with vegetation are considered in more detail in the following chapters.

The soil-vegetation-atmosphere system

About 70 % of the solar energy input to the climate system is absorbed by land surfaces and oceans, and released to the

atmosphere as thermal radiation, water vapour (latent heat), and sensible heat (McGuiffie & Hendersson-Sellers 1997). From this point of view, the temperature and moisture availability at the land surface are important driving forces for the conditions in the atmosphere (Viterbo 1996). However, when modelling the land surface-atmosphere exchange processes the opposite perspective is often more useful. The soil is then considered as the heat and water storage, with the vegetation cover as an interface to the forcing atmospheric conditions (Figure 1).

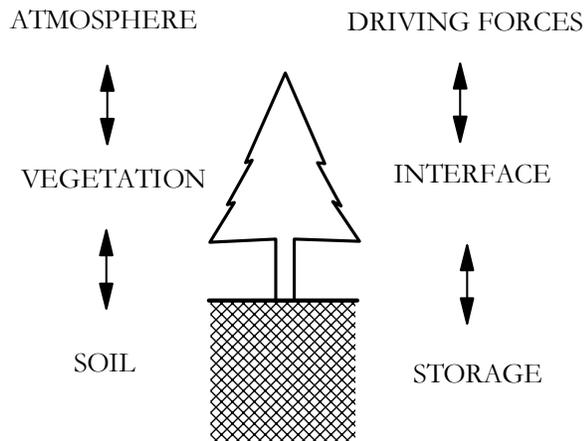


Figure 1 Basic components and their function in the soil-vegetation-atmosphere system, as represented in a typical SVAT model.

In this basic outline of the soil-vegetation-atmosphere system (Figure 1), the 'vegetation cover' represents any type of soil cover; bare soil, snow, grassland, trees, or a mixture of all of these (typical boreal region). It should also be noted that temporal storages of heat and water within the vegetation cover in many cases are as important as the storage in the soil, and consequently included in the models.

The most important processes for the land surface-atmosphere interaction are thus, as summarized by Male and Granger (1981):

- radiation balance of the land surface
- turbulent exchange of momentum, heat and humidity between the land surface and the atmosphere
- storage of energy in the land surface system

These can be formalized into the fundamental energy balance equation for the land surface:

$$R_n = H + LE + G \quad (1)$$

where the net radiation (R_n) equals the sum of the turbulent surface fluxes of sensible (H) and latent (LE) heat (evaporation), and the heat flux to the ground and/or vegetation cover (G). H and LE are defined as positive when directed upwards to the atmosphere, whereas the positive direction for R_n and G is downwards.

The net radiation is the net sum of downward and upward short-wave (solar) and long-wave (thermal) components, as governed by the atmospheric forcing, the surface reflectivity and the surface temperature. The fluxes of sensible and latent heat are functions of both the efficiency of the turbulent transport (the aerodynamic resistance) and the availability of heat and moisture at the surface. The storage term may incorporate several heat sinks in the land surface system, such as soil heat flux, latent heat of snowmelt, and heat for biogeochemical processes. The surface temperature is the one unifying variable for the energy balance components, and it can be derived by solving Eq. 1. However, water in its different states is probably the most important factor, since it is the dominating component of the overall heat capacity of the system, and its availability for evaporation governs the partitioning on latent and sensible heat flux.

Radiation balance

The net radiation of the surface is the sum of net short-wave (solar) and long-wave (thermal) radiation. The ratio of the incident solar radiation that is reflected at the surface is defined as the surface albedo, which of course is highly altered in the presence of snow. The albedo of newly fallen snow is about 90 % (Plüss 1997), which is high compared to the albedo of bare soil (10-30 %) and vegetation (10-25 %) (Monteith and Unsworth 1990). Long-wave radiation is emitted by the surface as a function of its surface temperature. Gases in the atmosphere absorb part of the long-wave radiation emitted by the surface, which raises the temperature of the atmosphere. Downward long-wave radiation emitted by the atmosphere is therefore an important component in the surface energy balance, keeping the Earth's surface temperature higher than it would be without this so-called greenhouse effect (e.g. Houghton 1997).

Turbulent exchange

The velocity of air blowing over a solid surface decreases towards the surface, due to the friction between the air molecules and the surface. The friction between the air molecules travelling with different velocities will transfer horizontal momentum vertically in the wind field. The flow is laminar in a very thin layer close to the surface and breaks down to a chaotic pattern of swirling motions in the air above. Fluxes of scalars, such as heat and moisture, along vertical gradients are greatly enhanced by the turbulent nature of the airflow.

The behaviour of the turbulent boundary layer depends on the relationship between mechanical and buoyancy forces. The

boundary layer is called stable if potential air temperature increases with the distance from the surface, since the denser air close to the ground tends to reduce the turbulent motions. Conversely, the boundary layer is called unstable when potential air temperature decreases with height. At neutral stability the wind profile is found to be logarithmic:

$$u(z) = \frac{u_*}{k} \ln\left(\frac{z}{z_{0m}}\right) \quad (2)$$

where $u(z)$ is the wind speed at the height z over the surface, u_* is the characteristic friction velocity, z_{0m} is the roughness length for momentum, and k is the von Karman's constant. A similar logarithmic profile for temperature (T) defines the characteristic scales T^* and the corresponding roughness lengths for heat z_{0h}^2 . The vertical fluxes of momentum (τ) and sensible heat (H) are defined by:

$$\tau = -u_*^2, \quad H = -\rho c_p u_* T^* \quad (3)$$

where ρ and c_p are density and specific heat of air.

There is no general theory for the turbulent transport of momentum and scalars in the surface layer at non-neutral stability. However, the Monin-Obukhov similarity theory predicts that the wind and temperature profiles in the surface layer can be described by universal functions of the atmospheric stability, based on general transport equations and an assumption of a constant flux layer above the surface (Högström and Smedman 1989). The bulk fluxes of momentum, sensible and latent heat from the surface to a reference height z are given by:

² The moisture flux can be derived by the same procedure, but it is left out here for simplicity.

$$\begin{aligned}\tau &= \rho C_M u_z^2 \\ H &= -\rho c_p C_H u_z (\theta_z - \theta_s) \quad (4) \\ LE &= -\frac{\rho c_p}{\gamma} C_E u_z (e_z - e_s)\end{aligned}$$

where C_M , C_H , and C_E are the exchange coefficients for momentum, heat and water vapour respectively. The turbulent exchange coefficients are commonly expressed as aerodynamic resistance $r_a^{-1} = C_H u_z$. The exchange coefficient for heat is described by:

$$\begin{aligned}C_H &= \frac{k^2}{\ln(z_a/z_{0M}) - \Psi_M(z_a/L) + \Psi_M(z_{0M}/L)} \quad (5) \\ &\times \frac{1}{\ln(z_a/z_{0H}) - \Psi_H(z_a/L) + \Psi_H(z_{0H}/L)}\end{aligned}$$

where L is the Obukhov stability length. Ψ_H and Ψ_M are the integration of the universal stability functions, which have to be determined experimentally.

Högström (1996) reviewed the experimental evidence for the theory, and concluded that it applies for unstable to near-neutral stable conditions. However,

the application of experimentally obtained stability functions in SVAT models results in unrealistic surface cooling under stable conditions (Beljaars and Holtslag 1991). These deficiencies can be reduced, for instance by modifying the Ψ functions in the stable regime (e.g. Holtslag and De Bruin 1988). However, other mechanisms than those predicted by the MO-theory are involved in strongly stable conditions. Zilitinkevich et al. (2002) recently presented a new formulation of C_H in strongly stable stratification, which take into account the influence of the free atmosphere on the exchange in the surface layer.

In this study only a simple correction was used, which minimized the aerodynamic resistance in stable condition. A windless exchange coefficient was introduced in Eq. (5) according to Jordan (1991):

$$c_H^* = c_H + c_{H0}. \quad (6)$$

EDDY-CORRELATION MEASUREMENTS

Field observations for the quantification of water and heat balance in the soil-vegetation-atmosphere-system were a major part of the thesis work. Three years of continuous measurements were conducted from November 1997 to August 2000 at an agricultural field in Marsta (59°55'N, 17°35'E), 9 km north of Uppsala, Sweden. The sensible and latent heat fluxes between the atmosphere and the ground were the most important variables and were also associated with the largest uncertainties. The method used for

these observations (eddy-correlation) is therefore discussed in more detail below. Details about other observations, such as soil temperature, soil moisture content, snow depth, radiation, and meteorological forcing can be found in the appended papers (I and IV).

Theoretical considerations

Sensible and latent heat can be measured with the so-called eddy-correlation technique, described in detail in e.g. Grelle (1997). The vertical flux of an entity transported by a turbulent airflow can be expressed as the product of the vertical wind component (w) and the concentration of the entity (x). In a turbulent airflow w

and x can be expressed as the sum of the mean (\bar{x}) and the fluctuation around this mean (x') according to "Reynold's convention" (Grelle 1997). If there are no sources or sinks of x within the surface layer and $\bar{w} = 0$, the average vertical transfer of x from the surface can be derived following:

$$\begin{aligned} \overline{(x + x')(w + w')} &= \overline{\bar{x}\bar{w}} + \overline{\bar{x}w'} + \overline{x'\bar{w}} + \overline{x'w'} \\ &= \bar{x} \cdot 0 + \bar{x} \cdot 0 + \bar{x} \cdot 0 + \overline{x'w'} \\ &= \overline{x'w'} \end{aligned} \quad (7)$$

where $\overline{x'w'}$ is the correlation between x and w . The eddy-correlation technique therefore depends on measurements of the vertical wind speed, temperature, and humidity at frequencies high enough to capture the turbulent frequencies responsible for the transport. The dominant turbulent frequency varies with height, stability, wind speed and surface roughness (Kaimal et al. 1972). Generally, higher frequencies (~ 20 Hz) are required closer to the ground above smooth surfaces than high above forests (~ 4 Hz) (Grelle 1997).

Field observations

The eddy-correlation system used at Marsta consisted of an R3 Gill Sonic Research anemometer and a closed-path gas analyser (LI-COR 6262)(I). The anemometer, mounted on a mast 3.5 metre above the ground, measured wind speed and air temperature simultaneously, whereas air humidity was measured in the gas analyser placed in a box on the ground. Air was sampled at the base of the anemometer, and sucked through a small tube into the gas analyser. The anemometer operated at 100 Hz but averages were stored on a local computer

at a frequency of 10 Hz. The time delay between the air inlet and the gas analyser readings was estimated as the time lag at the maximum of the correlation function between vertical wind speed and air humidity. Fluxes were calculated as 30-minute averages following the recommendations within the EUROFLUX community (Grelle 1999, personal communication). The fluxes were further corrected for frequency loss due to tube attenuation, sensor-misalignment and sensor-response time according to Massman (2000).

Evaluation of method

It was found that the latent heat flux was largely underestimated due to an unrealistic frequency loss in the vapour signal compared to temperature and vertical wind speed (I & IV; Figure 2 a). The lack of energy balance closure in the measured budget was also strong evidence of an underestimation of latent heat flux (I & IV; Figures 2 b-d). Comparison with SVAT-model simulations, in agreement with net radiation, sensible heat flux, soil temperatures and soil water storage, suggested that the latent heat fluxes were underestimated by 40 % by the eddy-correlation system (IV).

The use of 10 Hz instead of the recommended 20 Hz could not explain the deficiencies, since this would also have affected the temperature and wind speed signals. Most likely, the pathway between the air intake at the sonic anemometer and the gas analyser played an important role for the loss of frequencies. Peters et al. (2001) suggested that the filters used to remove particles might attract water vapour when they are filled, which would dilute the high frequency signals. The same

type of system has been shown to be reliable for measurements above forests (Grelle and Lindroth 1996) where the main

frequencies of the turbulence are smaller and the cleaner air makes the system less sensitive to filter changes.

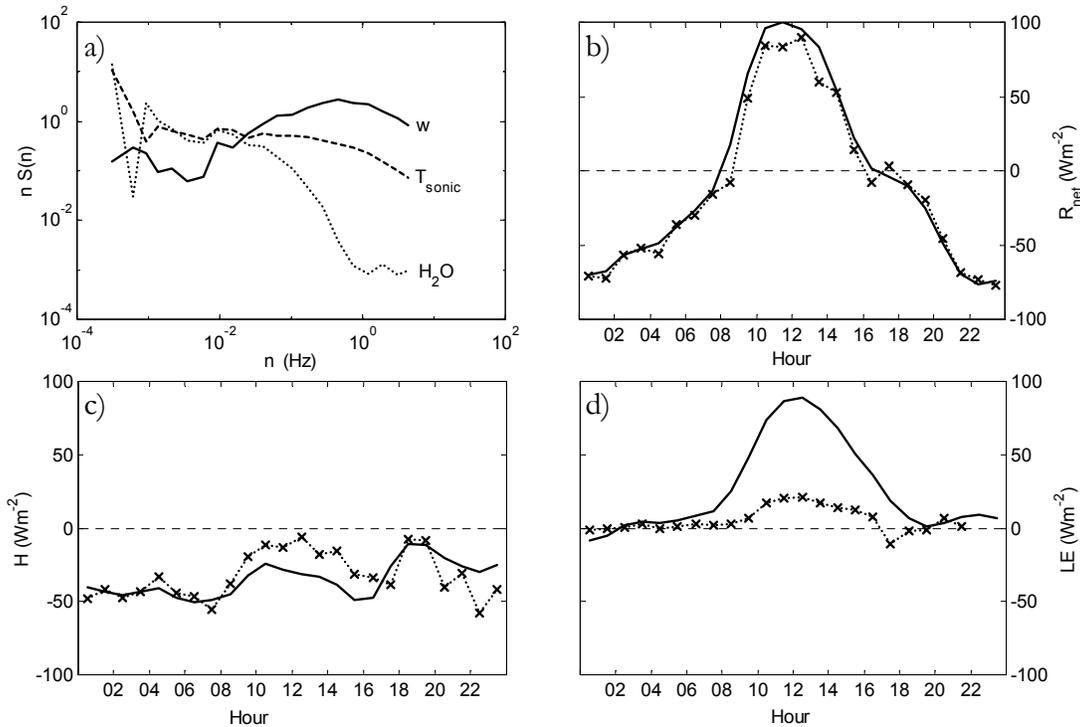


Figure 2 (a): Power spectra from the eddy-flux station estimated from 23 February 1998, showing an unrealistic frequency loss in the water-vapour signal compared to vertical wind-speed and sonic temperature signals. **(b-d):** Net radiation from the grassland (dashed line with markers) and sensible and latent heat fluxes from the eddy-flux station compared with SVAT-model estimates (solid). Modified from Halldin et al. (1999).

WATER AND HEAT BALANCE OF SNOW

The presence of snow greatly alters the land surface energy balance; net input of solar radiation is reduced due to the high snow albedo, turbulent heat exchange is lowered due to low aerodynamic roughness of the smooth snow surface, and heat flux from the ground is reduced as a result of the low thermal conductivity of snow.

Surface energy balance of snow cover

A number of snow energy balance models of different complexity have been used for different applications. Commonly, the surface energy balance and the soil heat flux are estimated to provide heat for the internal mass and water balance of the snowpack. The internal properties of the snow cover may be averaged in a single layer (e.g. Ohta 1994; Stähli and Jansson 1998; IV) or discretized into a number of

horizontal layers (e.g. Jordan 1991; Brun et al. 1989; Lehning et al. 1999).

The importance of the formulation of the internal structure of the snow pack for the surface heat fluxes was investigated in (I). Two snow models of different complexity were used as the upper boundary for the soil in the SOIL model (Jansson 1998): (1) the one-layer snow model described in Stähli and Jansson (1998) and (2) the multi-layer heat and mass-balance model SNTHERM developed by Jordan (1991). Field measurements of the water and heat balance components from an agricultural field were compared with simulations with the two different snow models. The comparison of the two snow models and the measurements showed that:

- The surface heat fluxes simulated by the two models were in many cases close to each other but differed substantially from the measurements during snowmelt and in the presence of a patchy snow cover.
- A good agreement between measured and simulated sensible heat flux were obtained for both models.
- Differences in the simulated soil temperature were attributed mainly to differences in the thermal conductivity of snow, even though the temporal dynamics of the observed soil temperature was somewhat better represented by the multi-layer snow model when snow depths exceeded 10 cm.
- The formulation of the internal structure of the snow pack was of less importance compared to the formulation of the surface heat fluxes, during conditions of shallow and patchy snow pack.

A sensitivity analysis of the governing parameters for the turbulent exchange above snow and frozen soils (IV) showed that:

- Measured downward sensible heat flux at 3.5 m height was 10 W m^{-2} too low to satisfy the estimated surface temperature in stable atmospheric conditions (Figure 3).
- The surface temperature was underestimated by 1°C on average during these conditions, which were found during 7 % of the study period (1 November 1997-1 August 2000)
- A minimum heat exchange coefficient corresponding to $5 \text{ W m}^{-2} \text{ }^\circ\text{C}^{-1}$ was needed to correctly simulate the snow surface temperature. 5 W m^{-2} roughly corresponds to a temperature decrease of $4 \text{ }^\circ\text{C hour}^{-1}$ in the air below the eddy-correlation instrument³.

The layered nature of a snow cover

A snow cover typically has a layered structure after a number of snowfall and snowmelt events. The sharp gradients of density and snow grain properties between adjacent layers greatly affect several important physical processes, such as melt water movement, adsorption of solar radiation, and avalanche release. A number of multi-layered snowpack models have been developed. The French model CROCUS (Brun et al. 1989) and the Swiss SNOWPACK (Lehning et al. 1999), have mainly been applied in high alpine terrain,

³ If we neglect the latent heat, the heat capacity of the air layer below the eddy-correlation instrument located at 3.5 m is $[3.5 \text{ m} \cdot 1.22 \text{ kg m}^{-3} \cdot 1004 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}] = 4.3 \text{ kJ }^\circ\text{C}^{-1} \text{ m}^{-2}$, which means that a heat flux of -1 W m^{-2} corresponds to a temperature decrease of approximately $[3.6 \text{ kJ hour}^{-1} \text{ m}^{-2} / 4.3 \text{ kJ }^\circ\text{C}^{-1} \text{ m}^{-2}] = 0.8 \text{ }^\circ\text{C hour}^{-1}$.

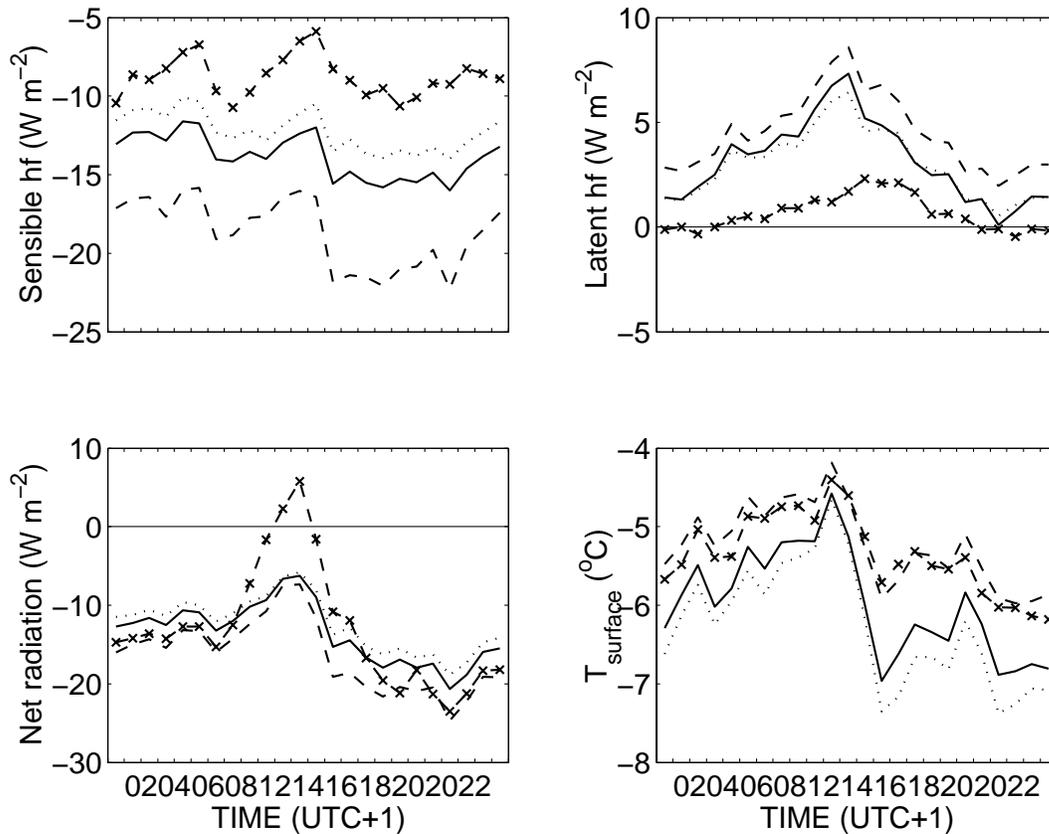


Figure 3. Diurnal average courses of net radiation, sensible heat flux, latent heat flux and surface temperature, for days with simulated snow depth deeper than 5 cm and measured surface temperature below zero, measurements (dashed line with markers) and three simulations with windless exchange coefficient for heat 0.001 m s⁻¹ (dotted line), 0.00163 m s⁻¹ (solid line), and 0.005 m s⁻¹ (dashed line), for Marsta 1997-2000 (re-drawn from IV).

whereas the SNTHERM (Jordan 1991) has been applied in low terrain both in open areas and in forests (I; Hardy et al. 1998; Koivusalo et al. 2001; Koivusalo and Kokkonen 2002). However, few reported studies have shown the ability of these models to simulate the development of snowpack layering in a sub-alpine environment, which is subject to frequent melting and re-freezing cycles.

The ability of the SNTHERM model to simulate the seasonal development of such a snowpack on a sub-alpine meadow in Switzerland was evaluated in paper (III). An extensive data-set from a field site in

the Alptal valley (Central Switzerland, 47°N, 1200 m.a.s.l) was available from the unusually snow-rich winter 1998/99 (Waldner et al. 2000). The maximum snow depth was 1.8 m with a corresponding snow water equivalent of 550 mm. The observations included snow physical properties such as snow depth, snow water equivalent, snow density profiles, and photographs of translucent snow profiles. The comparison of observations and simulations showed that:

- The model successfully simulated the seasonal development of total snow depth and average snow density, as well

as the surface energy balance and temperatures in snow and soil.

- The model reproduced fairly well the overall observed density profiles, but the layering was much more pronounced in the observations.

A detailed sensitivity analysis of different model parameters was done to identify the most important processes for the layering and why the model was unable to reproduce the observations correctly. It was found that the formation of distinct layers in the model simulations was most sensitive to parameters governing the water redistribution within the snowpack. A reduced hydraulic conductivity increased the densification of individual layers, due to increased re-freezing of infiltrating melt water. The numerical strategy for mixing thin layers was also believed to have contributed to the more smooth density profiles found in the simulations compared to the observations. The simulated layering was less sensitive to other processes included in the model, for instance snow surface energy balance, compaction of snow, and snow grain growth.

However, the sensitivity analysis showed that further calibration could not improve

the model performance with regard to the snow pack layering. Several important processes were disregarded, for instance the effect of capillary barriers that slow down the vertical water flow, induce lateral flow, and may form ice-layers (Marsh and Woo 1984). It is also known that the compaction of snow layers is influenced by the snow microstructure, by its importance for the snow viscosity. Moreover, the surface energy balance is known to be a key-process for the formation of surface hoar (Lehning et al. 2001). Additional simulations with the SNOWPACK model (Lehning et al. 1999) showed that a more pronounced layering could be obtained if the effect of the snow microstructure on its viscosity and elasticity was accounted for.

It was thus concluded that the major limitations for a more realistic simulation of the snow pack layers were:

- neglecting effects of snow microstructure on the compaction rate,
- the current formulation of water flow in the snowpack, which does not account for capillary barrier effects, preferential flow, and lateral flow.

WATER AND HEAT BALANCE OF VEGETATED SURFACES

The description of the interface processes (Figure 1) is probably the most crucial aspect when modelling the water and heat exchange between a vegetated surface and the atmosphere. Several sub-surfaces contribute to the overall evaporation and heat exchange, for example trees with thousands of individual leaves and the soil surface below the canopy just to mention a

few. All of these may have their individual response to the environmental conditions. The following sections cover the experiences from paper II, IV and V with regard to the heat and water balance of arable land and forest in the boreal zone at high northern latitudes.

Models for vegetated surfaces

The most important sources of evaporation are the canopy transpiration, evaporation of rain or snow trapped on the canopy (interception), soil evaporation,

and evaporation from snow below the canopy.

Canopy transpiration

The Penman equation (Penman 1953) as modified by Monteith (1965), which often is referred to as the Penman-Monteith equation (PM-equation), specifies the evaporation from a surface with a specific surface resistance:

$$E_c = \frac{1}{L_v} \frac{\Delta R_{nc} + \rho c_p \delta e / r_a}{\Delta + \gamma (r_a + r_s) / r_a}$$

where δe is the vapour pressure deficit in the air, Δ is the slope of saturation pressure versus temperature, ρ is the air density, c_p is the specific heat of air, γ is the psychrometric constant (66 Pa K⁻¹), r_a is the aerodynamic resistance, and r_s is the surface resistance.

The PM-equation is often used to estimate the transpiration from a canopy, represented as one big leaf with a single canopy surface resistance (Goudriaan 1989). The surface resistance accounts for the stomata regulation of the water availability for evaporation, which is often described as a function of short-wave radiation and atmospheric vapour pressure deficit (Jarvis 1978; Lindroth 1985). The sensible heat flux from and the corresponding temperature of the canopy can be calculated using the residual of the net radiation and the latent heat flux estimates. The disadvantage with the PM-equation is that it neglects the feedback between the transpiration and the temperature and vapour pressure in the canopy air (Halldin and Lindroth 1986; Goudriaan 1989).

Surfaces below canopy

The PM-equation can also be applied for the water and heat exchange from other sub-surfaces, such as bare soil, snow

surface or intercepted water, if the corresponding surface resistances are defined. The resistance scheme for a vegetation cover and the underlying soil/snow that was used in (IV and V) is shown in Figure 4. The reduced turbulence within the canopy has been taken into account merely as an additional aerodynamic resistance for the soil/snow surfaces, but more elaborate methods have been developed (Goudriaan 1989).

However, a dynamic approach (an iterative solution of Eq. (1), where H and LE are estimated by Eqs. (4) and (5) that takes into account the feedback between net radiation, surface temperature and aerodynamic resistance is often more appropriate for soil and snow surfaces in open areas or below the canopy, where sharp temperature gradients may arise during stable conditions. This solution can also be applied for the canopy evaporation (II and V).

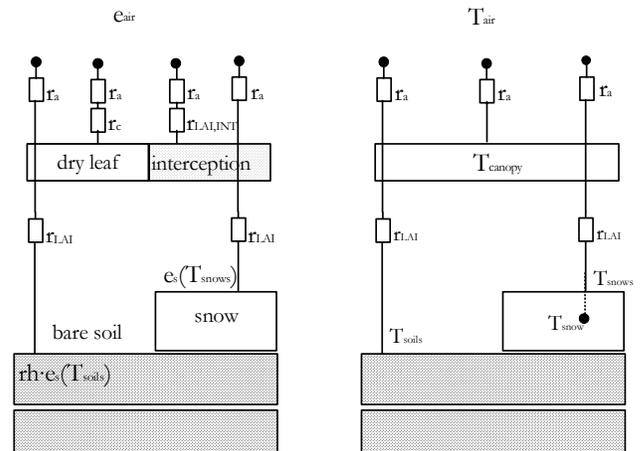


Figure 4 Resistance schemes for latent heat flux (left) and sensible heat flux (right) from the surface compartments: bare soil, snow, and canopy (transpiration and interception) as defined in the COUP model (Jansson and Karlberg, 2001).

Interception of rain and snow

The interception of precipitation in the canopy is of great importance for the surface water and heat balance. The relatively simple approach to simulating the interception process used in (IV and V) is presented in Figure 5.

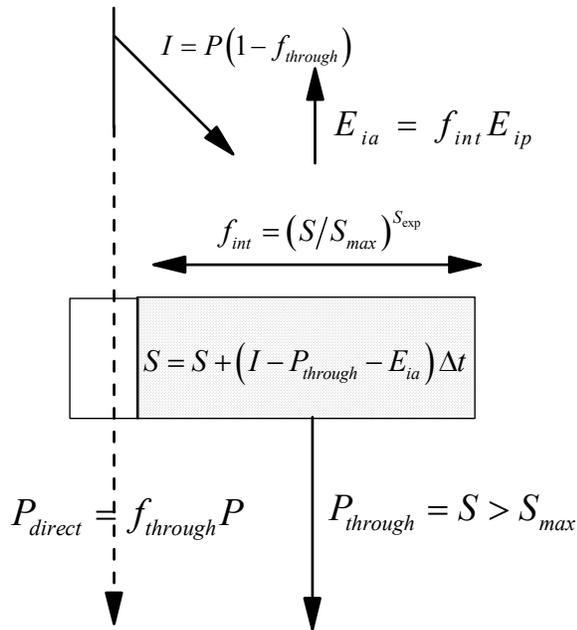


Figure 5 Interception model from Jansson and Karlberg (2001).

A given proportion of the precipitation is assumed to be intercepted in the canopy, as long as the interception storage is below the capacity. The interception capacity is estimated as a function of the fraction of ice in the intercepted water, which is modelled as a function of the air temperature, and the estimated snowmelt. Evaporation of the intercepted water is estimated with the PM-equation with a surface resistance representing the within-canopy aerodynamic resistance. The albedo of the snow-covered fraction of the canopy is calculated as a weighted average of the vegetation albedo and the albedo of the snow.

Model applications

Three different models have been used to explore the land surface heat exchange processes with respect to forest and arable land (II, IV and V). Two of the models were land surface schemes used in the ECMWF⁴ GCM model (VB95, Viterbo and Beljaars 1995; TESSEL, van den Hurk et al. 2000), whereas the third model was a SVAT model developed from a hydrological local-scale perspective (COUP; Jansson and Karlberg 2001). In addition to the dataset from Marsta agricultural site 1997-2000 (I and IV), data on surface heat fluxes, sub-surface conditions and meteorological variables from the NOPEX forest site (Lundin et al. 1999) were available for a three-year period 1994-1996 (II).

The conceptual differences between the models were mainly related to the spatial representation of surface elements, and to the formulation of some specific processes.

The VB95 model assembles the sub-surfaces (bare soil, snow, vegetation, intercepted water) into one energy balance equation, using a common surface temperature and individual estimates of evaporation. In the TESSEL model, each surface component represents an areal fraction, and the energy balance equation is solved separately for each fraction (mosaic approach, cf. Koster and Suarez 1992). The COUP model describes a vertical structure of the surface compartments, and allows for simultaneous water and heat exchange from the vegetation layer and soil/snow surface

⁴ European Center for Medium Range Weather Forecasts, Reading, UK

below (Figure 4, mixed vegetation cover approach, cf. Koster and Suarez 1992).

Four processes were defined differently in the models, namely:

1. Increased aerodynamic resistance for evaporation from snow below vegetation
2. Reduced water uptake of vegetation from frozen soil
3. The response of stomata and thus transpiration to vapour pressure deficit
4. Interception of snow

Processes 1-3 had been refined with respect to boreal forest in the TESSEL model compared to the VB95 based on an evaluation with data from the BOREAS⁵-experiment in Canada (Betts et al 1998). However, only the COUP model also accounted for interception processes in the presence of snow.

Water and heat balance of forest

The summer latent heat flux was mainly regulated by the canopy surface resistance, as controlled by leaf area index, radiation, and vapour pressure deficit. The diurnal variation of evaporation was improved in the ECMWF model due to the inclusion of a transpiration response to vapour pressure deficit (II). Comparison with evaporation estimated with constant surface resistance in the PM-equation, showed that the account for soil moisture stress substantially improved the explanation of observed variability (II).

However, all models systematically underestimated evaporation in the late summer. A sensitivity analysis (II) of some governing parameters showed that the

models could be further improved by inclusion of:

- Natural seasonal variations in leaf area
- The ability of trees to extract additional water from deeper levels in the soil during conditions when the uppermost layers are dry.
- The variation of vegetation properties within and between years related to climatic conditions during winter.

The winter latent heat flux was significantly overestimated when snow evaporation was estimated based on the aerodynamic resistance above the canopy, and a surface temperature that was allowed to exceed zero degrees (VB95). A better agreement with measured winter latent heat fluxes was obtained if snow evaporation was considered from below the canopy only (TESSEL). On the other hand, simulated evaporation from intercepted snow contributed to a better model agreement with observed winter sensible heat flux (COUP). The maximum simulated evaporation of intercepted snow was 0.32 mm hr⁻¹, which corresponded well with values reported in the literature (e.g. 0.3 mm hr⁻¹ in Lundberg and Halldin 1994; 0.56 mm hr⁻¹ in Lundberg et al. 1998).

The models generally simulated the sensible heat flux correctly, as long as the latent heat flux was represented satisfactorily. However, downward sensible heat flux in stable atmospheric conditions was underestimated unless an upper limit of aerodynamic resistance was used (II and V). This produced an unrealistic surface cooling to balance the negative net radiation, unless a higher heat exchange between the canopy and soil was assumed. However, the latter implied that heat is taken from the soil instead of from the air,

⁵ Boreal Ecosystem-Atmosphere Study (Sellers et al. 1997)

which contradicted the measurements of both sensible heat flux and soil temperatures (II). The best results with regard to forest sensible heat flux, net radiation and soil temperatures were obtained with an explicit representation of

evaporation from intercepted snow, an upper limit of aerodynamic resistance during inversions set to 500 s m^{-1} , and a thermal conductivity in the top soil layer representative for the specific site.

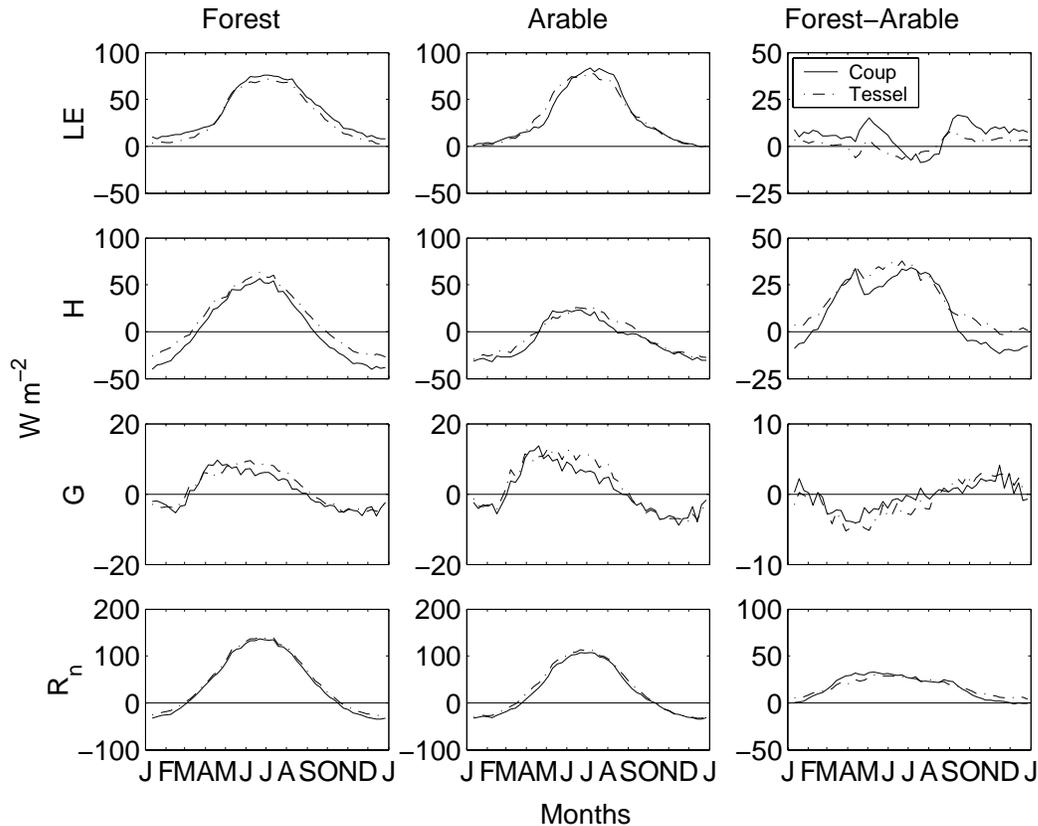


Figure 6: Annual cycles of latent heat flux, sensible heat flux, ground heat flux, and net radiation for forest (left column), arable land (middle column), and the discrepancy between forest and arable land (right column), simulations with the Coup and the Tessel models using the Marsta dataset 1970-1999 (re-drawn from V).

Water and heat balance of arable land

The seasonal variation in the surface energy balance of arable land differed greatly compared to forest (Figure 6), due to the typical seasonal variation of the arable crops. To obtain a reasonable simulation of the surface heat fluxes, a seasonal development of leaf area index and canopy height had to be used, with the start of the growing season in May, the

maximum leaf area index and maximum canopy height in July, and harvest at the beginning of September (IV and V; Persson 1997). Other important surface properties were the low roughness length and the high albedo in winter.

The systematic errors in the measurements of latent heat flux were a major limitation for the assessment of the surface energy balance of the arable land, as already

discussed in the previous chapters. However, the model agreement with net radiation, sensible heat flux, soil temperatures, and soil moisture storage was used as indirect evidence for the simulated latent heat fluxes (IV and V). The simulated sensible heat flux was very sensitive to the parameterization of the transpiration control and the aerodynamic resistance as a function of leaf area, canopy height and presence of snow. The diurnal variation of summer sensible heat flux was improved if the impact of radiation, vapour pressure deficit and seasonal variation in leaf area index on canopy surface resistance was accounted for (IV).

The overall agreement between simulated and measurements were very good (IV),

except for a limited period of time at the end of the vegetation season in 1999. However, the systematic underestimation of sensible heat flux during this period was also evidence of the robustness of the parameterization. The plants had been killed with pesticides at the beginning of July, and consequently transpiration was inhibited. The model assumed the plants to be active without any loss of leaf area or transpiration capacity until the prescribed harvest in September, which demonstrated the importance of water availability for the partitioning of net radiation into latent and sensible heat flux.

DISCUSSION AND CONCLUSIONS

This study focused on the formulation of the boundary conditions and the accuracy of surface flux estimates for snow cover, arable land and forest. Winter processes were given most attention, even though the seasonal dynamics of forests and open arable land were considered as well. The main contribution of this study was to describe the surface water and heat balance of the different surface types, using long-term datasets and soil-vegetation-atmosphere transfer models based on existing theories.

Snow cover processes

The range of the snow cover conditions was considerable in the study material, as was the quality of the available snow cover information. The two-metre-deep snowpack in the Alptal study (III) was in sharp contrast to the shallow and patchy snowpacks observed at the agricultural field in Marsta (I+IV), where the

maximum depth was 35 cm and the longest period of continuous snow cover one month. Obviously, a wide range of conditions is a firm base for testing the generality of a model approach, but probably the contrast between the two datasets (representing two extremes) was too large in this case. Undoubtedly, a multi-layered description of the internal heat and mass balance would have been of much more relevance for the snow surface energy balance if it had been applied to the Alptal data (III) than what was found in (I). Moreover, it is well known that the assumption of a steady-state heat flux between the snow surface and the soil tends to overestimate the soil surface heat flux for deeper snow covers (e.g. Jansson 1987). However, with the present data from Marsta it was not possible to elucidate whether the inclusion of the heat storage in the snowpack significantly improved the simulation of soil temperature. In the Alptal dataset, the temperature at the snow/soil interface

below the deep snowpack was zero throughout almost the entire winter, due to the retention of liquid water at the bottom of the snowpack. Thus, the high vertical resolution of the snowpack was of little importance for the simulation of the soil temperature. Koivusalo et al. (2001) showed that a two-layer approach was needed to describe the behaviour of the surface heat fluxes above a seasonal snowpack in northern Finland. In conclusion, a one- or two-layer approach, including one snow surface temperature and at least one snowpack temperature, is an appropriate approximation level in applications where only the bulk water and heat balances of snow are of interest.

It was found that near surface heat storage terms had to be accounted for in order to explain both the measured sensible heat flux and surface temperature in stable atmospheric conditions. However, for many model applications it is sufficient to avoid the unrealistic cooling of the surface, which otherwise may cause numerical instability. A simple correction of the aerodynamic resistance using a windless exchange coefficient was useful to quantify the effect of the phenomena. However, for applications within numerical weather prediction models, general methods are more useful, for instance the one recently presented by Zilitinkevich et al. (2002).

Vegetated surfaces

Differences in the aerodynamic properties and in the paths of evaporation from dry, wet, or snow-covered vegetation or soil were the most important explanations for the differences in seasonal water and heat budgets between forest and arable land. The total evapotranspiration from the arable land and the forests were equal. The

transpiration from the arable land was higher in summer, but the interception evaporation from the forests was higher at all times of the year, especially in winter. The forest contributed with more sensible heat flux to the atmosphere than the arable land. The latter surface type was a small net sink of sensible heat flux due to the large downward sensible heat fluxes used for snowmelt in winter, and the relatively low sensible heat fluxes in summer (V). Net radiation was higher for the forest than for the arable land (on average the forest absorbed 60 % more radiation than the arable land). The main part of this extra radiation was received from March to October. The shading by the forest trees of the snow lying beneath was thus most important at the end of the winter. However, many years were characterized by shallow snow cover and bare soil conditions that could occur at any time during the winter. In these cases the difference between the net radiation of the forest and the arable land was less pronounced during spring (V).

The general improvements by the new ECMWF land surface scheme in the representation of a boreal forest were promising. The improvements were mainly due to the introduction of separate energy balances for individual surface types and improved parameterizations of some key biotic and abiotic processes (II). Sensitivity analysis of some key parameters showed that further model improvements were possible if seasonal variation of vegetation properties was included. The corresponding evaluation of the COUP model (V) showed that the wintertime surface energy balance of forest was further improved if snow interception processes were accounted for.

The simulated sensible heat flux and the surface temperature of the forest were sensitive to the formulation of turbulent exchange coefficients in stable conditions (V). The results are consistent with the earlier discussion about the turbulent heat exchange above snow surface (IV). However, the storage terms in the air seemed to less important above the forest, since it was possible to satisfy both measured surface temperature and downward sensible heat flux by minimizing the aerodynamic resistance in these conditions.

Conclusions

- The models were useful to formalize our current knowledge of the system behaviour, and as tools to evaluate the usefulness of the measurements.
- The model agreement with measured sensible heat flux, net radiation, and soil water and heat balances demonstrated that eddy-correlation measurements might have underestimated latent heat flux by 40 % over an arable field.
- Internal heat and mass balance of a shallow and patchy snow cover was of little importance for estimation of the surface fluxes.
- Heat storage in the air below the eddy-correlation instrument had to be accounted for in order to explain measurements of surface temperature and sensible heat fluxes above snow during stable atmospheric conditions.
- The seasonal variation of vegetation properties and behaviour played a

major role for the sensible and latent heat fluxes from forest and arable land.

- The model formulations of snow evaporation and interception evaporation were demonstrated as being the major component in the description of the wintertime surface energy balance of forests.

Future directions

This work has pointed out several issues that need to be addressed in future research. Firstly, the vertical distribution of snow/water in the forest was demonstrated to strongly influence the overall surface energy balance. Unfortunately the available measurements did not fully resolve the problem, therefore independent measurements of snow interception are critical.

Secondly, this study demonstrated the models ability to describe the behaviour of forests and arable land as separate systems. It would also be of great interest to investigate the interplay between different surface types in a heterogeneous landscape using coupled high resolution models.

In addition, the interaction between abiotic and biotic processes need to be further understood in order to increase the accuracy of model predictions on both seasonal and inter-seasonal time-scales. Additional research is recommended on the feedback between atmosphere and land surface biogeochemical processes. This is not only of major interest for long term prediction of the future climate, but also to improve the present meso-scale meteorological models.

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