

EVALUATION OF HTSVS USING OBSERVATIONAL DATASET

A
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ABSTRACT

This study summarizes some of the results obtained from one European field programs. WINTEX , undertaken in the Boreal/Arctic regions in 1996-98. These studies enhance our knowledge of the various water and energy fluxes of the region Sodankylä (Finland).The hydro-thermodynamic soil-vegetation model HTSVS (Hydro-Thermodynamic Soil-Vegetation Scheme) was initialized taking into account the soil frost and the snow metamorphism process. Long wave down radiation, short wave down radiation, air temperature, precipitation data were among data used for forcing, simulations were run based on these data. Sensitivity analysis was carried out by varying various parameters to see the difference in the results. The simulated results were then compared with the field data and it was found that simulated results for the snow height followed the observed data; though in the first month it is slightly under predicted and towards the end of the period it over predicts the snow height it is suspected that this deviation from the observed could be due to the radiation data. Different soil types were found to fit different soil layers which emphasize the need to consider vertical heterogeneity of soil These results are encouraging and the importance of proper initialization is also realized. Thus these results lead to better understanding of how and what parameters to change to improve the land-surface models that is essential to gain an insight into the climate system.

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CHAPTER 1

INTRODUCTION

1.1 Origin and Scope of the Research

There has been considerable effort in the past few years to study climate and to identify the parameters that play an important role in global climate and climate change. Under changing climate conditions, there is an acute need to predict water availability for various applications like, for example, food production or water resource management. Investigations on the effect of altered climatic conditions require state-of-the-art models, which incorporate soil moisture, soil temperature and other processes. To investigate the effect of these processes on the hydrological cycle. A strong emphasis has been placed on predicting ground water recharge and discharge in response to climate. Typically, the water table will rise during long persistent precipitation events, sink, and even uncouple from soil moisture during long lasting drought episodes (e.g., Mölders and Rühaak, 2002). Recycling of precipitation will decrease during drought period and increase during extreme precipitation events, because a high water table ensures that sufficient moisture is available for evaporation of water from soils, and transpiration by plants, hereafter referred to as evapotranspiration.

The evaluation of the hydro-thermodynamic soil-vegetation scheme (HTSVS) and the present study will allow improved weather forecasting. The local recycling of precipitation would be better predicted, along with showers,

fogs to name a few. Better land surface models will improve flood forecasts and adequate warnings can be sent out in advance reducing damage to life and property. Understanding of these processes increases the possibility to predict icy rain or how long a snow cover will last. This knowledge can help to minimize the economic damages.

1.2 The Hydrologic Cycle

The hydrologic cycle is fueled by solar radiation. The movement, circulation, and conservation of the earth's total water content are termed the "hydrologic cycle". The hydrologic cycle starts with evaporation of water from oceans and/or from the continents into the atmosphere. When moist air is lifted, it cools and condensation results in cloud formation. Moisture is transported in form of water vapor, cloud- and precipitation hydrometeors around the globe until it returns to the surface as precipitation in liquid (rain) or solid form (snow, graupel, hail). Once precipitation reaches the ground, a part of it infiltrates into and percolates through the ground and leading to groundwater formation. Groundwater seeps into the ocean, rivers, lakes, or streams. When the rate of rainfall exceeds the land's ability to absorb it, the surplus water - termed generally as runoff - either discharges into a lake or some other water body. Finally, evaporation provides water back to the atmosphere. Some part of the water infiltrated into the ground is taken up by plants through their roots, and released into the atmosphere through transpiration (evaporation of water from plants).

The water balance is a quantitative view of the hydrologic cycle. The movement of water through the cycle holds the key to the distribution of moisture over the surface of our planet.

1.3 Soil Frost and Snow Metamorphism

Most of the arctic and the sub-arctic regions have permafrost or frozen soil during winter. Freezing or thawing of soils plays an important role in the climate in high latitudes. These processes influence winter temperatures because, during the freeze process, the effective heat capacity of the soil is increased by a factor of twenty (Viterbo *et al.*, 1999). This can be explained by the fact that the thermal conductivity of ice is about four times of that of water. During spring, a major portion of the radiation goes in melting of frozen ground and snow. This energy will be available again in early winter when the ground refreezes. Frozen ground or soil frost impedes infiltration and percolation, thereby reducing the mobility of soil water, which exacerbates the spring snowmelt flood (Cherkauer and Lettenmaier, 1999). Some areas might even become water logged due to this effect. Another important aspect of permafrost is the local equilibrium between ice, gaseous and liquid (Mölders and Walsh, 2004). Any small change in heat diffusion and conduction will affect snow thickness and consequently the three water phases in the soil (soil liquid water, soil ice, water vapor in the pore space) and due to phase transition processes there will be either latent heat released or energy consumed leading to a change in soil temperature. This change further

propagates in the sense that any change in soil temperature will influence the freezing and thawing process, which will release or consume the latent heat again altering soil temperature. Not much water is available until spring when the atmosphere has warmed enough and the ground has thawed. This leads to low evaporation in the winter season and a reversed situation during autumn. In winter, transpiration plays a very small role because deciduous forests have lost their leaves and even stomatal conductivity values are low even for coniferous forests compared to during warmer weather. This results in storage of moisture in the soil.

Modeling of frozen ground is challenging and has received some attention (Harlan, 1973; Guymon and Luthin, 1974; Konrad and Duquennoi, 1993). The models account for water infiltration and movement in frozen soil and are computationally very intensive since they require iterative techniques (Tao and Grey, 1994) or high-resolution finite difference schemes.

Frozen ground (soil frost) and snow cover are the most common surface conditions in the high latitudes from October to mid-May. Snow cover plays an important role in changing regional as well as global climate by affecting the surface energy balance (Yeh *et al.* 1983; Namias 1985; Walsh *et al.* 1985; Barnett *et al.* 1989) and the hydrological cycle. Important aspects of snow, which affect the hydrological cycle the most, are albedo, thermal conductivity, the reduction of roughness length and the ability of snow to store water (Slater *et al.* 2000).

Snow metamorphism refers to changes in the crystalline structure of snow due to temperature and pressure. The change could occur due to sintering, the structural strength of the snow pack, the permeability of the snow pack, the reflectivity of the snow surface, the thermal conductivity of snow, and finally yet importantly snow density. Snow metamorphism and the depth of snow regulate the soil freezing (Williams and Smith, 1989), which has implications on soil hydraulic properties and the survival of some plants (Kongloli and Bland, 2000). The presence of snow acts to stably stratify the atmospheric boundary layer (ABL) and one of the important implications of the ABL is the reduction of movement of trace gases (Segal *et al.* 1991). The presence of snow radically affects the radiation budget. Albedo changes appreciably between 0.35 and 0.9 for old and new snow (e.g. Oke, 1978; Pielke, 1984). Snow also changes the surface aerodynamic properties and acts as an insulator for the soil beneath. Snow has an overriding influence on the runoff. Snow melt is one of the major hydrological events. Trace constituents accumulated in the snow pack over the winter will be released in a very short period into the ecosystems (Dingman, 1994). The duration and timing of the snow cover influences the micro and macro climate conditions because the surface energy balance is changed by the snow cover (Zhang *et al.* 2001). A discontinuous snow cover, which means isolated soil surface surrounded by snow, can result in significant heat fluxes and vertical mixing in the ABL especially when there is strong radiation. This difference in energy budget between the snow-free and snow-covered areas can lead to a

significant advective flow similar to a sea breeze (Baker *et al.*, 1999). Low evaporation and drainage impeded by frozen ground often means that soil remains saturated with moisture during winter.

1.4 Motivation

If the soil freezing process was not included in numerical modeling too high water vapor fluxes will be predicted as there would be liquid water available. Since soil frost hinders infiltration of water into the soil and thus in case of a rain event onto frozen soil or melting of snow-pack will contribute to runoff. For accurate determination of the energy fluxes, it is crucial to know the exact depth of freezing. Knowledge of the freezing depth requires a fine resolution, which unfavorably increases simulation time.

Another important process to be simulated is snowmelt and previous snow accumulation (Foster *et al.*, 1996; Cayan, 1996). Thus, a physically adequate formulation of snow metamorphism processes is required to guarantee the transferability of a model to other regions. The length of the growing season can be affected strongly by the change in length of the active growing season, which when simulated in climate models, in high latitudes is strongly dependent on the correct timing of the snowmelt (Llyod, 2000).

High latitude terrestrial conditions and the related processes have received limited attention because of practical and financial considerations. Most mesoscale models do not consider snow metamorphism (ablation and

accumulation) and only change the surface roughness length, albedo, emissivity (Eppel *et al.*, 1995) or at the most, a single layer snow model (Koren *et al.*, 1999; Warrach *et al.*, 2001; Chen and Dudhia, 2001). Specification of an albedo and emissivity typical of snow-cover improves the prediction, but does not account for some of the micro-meteorological and micro-climatologically conditions namely the near-surface temperature, humidity, latent heat fluxes, soil heat fluxes, associated with the snow cover (Fröhlich and Mölders, 2002). Simple one-layer snow models often assume a homogenous snow pack and work with the energy budget. Sometimes snow covers only a fraction of the grid cell like BATS (Dickinson *et al.*, 1993), for instance, assumes a one-layer system with fractional coverage of the grid square. Heat and water fluxes are simulated according to the snow age and grid cell fraction covered by the snow (Dickinson, 1993). In the case of one layer snow models, once the snow reaches the freezing point, further input of energy will result in melting of snow. Once the snow is saturated with water there will be outflow.

These aforementioned often-used simplifications can lead to unrealistic results (1) for thick snow packs, (2) if frequent freeze and thaw cycles occur, and for (3) non-homogeneous snow packs (Loth, 1995). So far, it is not clear how much detail of the physical processes within a snow pack can be determined and what degree of complexity is required of snow models for use in atmospheric models (Yang *et al.*, 1997; Slater *et al.*, 2000).

Various studies have been performed to overcome the shortcomings of one layer snow model. Various authors (e.g., Loth *et al.*, 1993; Lynch-Stieglitz, 1994; Schlosser *et al.*, 1998; Slater *et al.*, 2001; Fröhlich and Mölders, 2002) found that models, which consider multi-layers instead of a single layer, provided better results especially for the high latitudes. Different techniques were used for a comparative study of (1) Use of mixture theory (This is a part of volume fraction theory which models the energy and mass interchange between the constituents which make up the mixture in this case snow pack, The second law of thermodynamics is used to impose restrictions upon the various constitutions) by the snow model of interest to simulate multiphase water and energy transfer processes in the snow layers (2) A simple three layer snow model (At the each layer the temperature, snow water equivalent and other parameters are simulated) (3) Force restore method (In this method the forcing by sum of fluxes to the atmosphere is modified by a restoring term which contains the deep soil temperature term Bhumralkar's [1975]) used to calculate snowmelt from the energy budget and snow temperature. The different models used all the above-mentioned checks. These models ran simulations; each of them simulated the time series of snow water equivalent, surface temperature and fluxes very well. The mixture theory gave the best match, but had a drawback of requiring intensive computational efforts. The three layer model gave results similar to mixture approach but was less computational intensive compared to the mixture

approach. The force restore method performed averagely requiring computational efforts similar to the three layer snow model. (Jin *et al.*, 1999).

In recent times, there has been work done to develop and improve soil models by inclusion of soil-frost parameterizations for use in mesoscale meteorological models (Koren *et al.*, 1999; Boone *et al.*, 2000; Warrach *et al.*, 2001). The role of soil moisture is important since anomalies in soil moisture affect the precipitation, temperature and the wind fields at the global (Shukla and Mintz, 1982) and regional scales many studies have been conducted on the role of liquid soil moisture in climate whereas the role of seasonally frozen soil moisture receiving little attention. The hydrological significance of freezing processes for the 35% of the Earth's surface, which is subject to freezing and thawing (Williams and Smiths, 1989), is well, documented (Rouse, 1984; Woo, 1986). Given the importance of soil frost, it is rather surprising that there is lack of validation of land surface schemes in simulating soil frost. Models, which apply the force-restore method, are limited in resolving the soil horizontally. The above mentioned force-restore models cannot simulate the vertical distribution of soil processes like the diurnal variation of the boundary between an unfrozen upper and a frozen deeper soil layer as they work with two or three reservoirs.

Some efforts have been made to incorporate the affect of soil frost processes in largemodels. Koren *et al.* (1999) tested and evaluated a soil-frost model offline that was developed for the NCEP (National Center for Environmental Prediction) Eta model. Warrach *et al.* (2001) also designed and

evaluated a one-layer snow parameterization along with soil frost for use in hydrological and atmospheric models.

The freeze-thaw process influences the thermal and hydrological regime (Slater *et al.*, 1998; Viterbo *et al.*, 1999; Boone *et al.*, 2000; Warrach *et al.*, 2001). Most of the mesoscale meteorological models, even if they include soil-frost, ignore cross-effects, i.e. interactions between the soil thermal and moisture regimes, namely the Ludwig-Soret effect (a temperature gradient causes in water flux and changes the soil volumetric water content) and Dufour effect (a moisture gradient can contribute to the heat flux and alters soil temperature; de Groot, 1951; Prigogine, 1961; Kramm *et al.*, 1996). For short term predictions these cross-effects can be neglected, but for long term prediction, and when chemical are considered, dry soil conditions suddenly enter the wet mode, around freezing point, snow melts, these processes may influence other variables or other processes. For example, the Ludwig-Soret process influences water recharge by 5% of the total recharge and Dufor effect changes soil temperature up to 2K over the long-term (Mölders *et al.*, 2003a). A study conducted by Mölders and Walsh (2004) shows the importance of these cross-effects. The soil volumetric content was found to vary by up to 0.188 mm/d between simulations with and without inclusion of these cross effects. In the high latitudes, largest variations were found during the snowmelt, during freezing, and freezing/thawing processes of the active layer. Mölders and Walsh [2004] find that at soil temperatures below freezing point liquid water content decreased due to an increase in soil ice. The

decrease in liquid water caused the Ludwig-Soret effect to become more prominent. Also around the freezing line, the gradient in the soil moisture may become strong, resulting in changes in the soil temperature (Mölders and Walsh, 2004). During the snowmelt, soil moisture changes quickly due to thawing of frozen ground and the uppermost layer is most affected. While the deeper layers of soil are more sensitive during the freeze up (Mölders and Walsh 2004). The importances of the Ludwig-Soret and Dufour effects have been ignored in the past in formulations of soil models (Koren *et al.*, 1999) because of the increase in computational time if they were incorporated.

There is an urgent need to consider the feedback between the land and atmospheric part of the water cycle in climate modeling. Due to complex interaction between various thermal and hydrological fluxes and their dependence on soil water phase, soil temperatures require proper initializations in an attempt to reduce errors. The aim of this study is to simulate soil frost and snow metamorphism with HTSVS and to evaluate the performance of the model. For this research, the stand-alone version of HTSVS is driven by meteorological data from the WINTEX (WINTer EXperiments) experiments. This evaluation is performed as previous evaluations were carried out either for unfrozen (e.g., Kramm 1995) or for frozen soil with vertically varying soil characteristics (e.g., Mölders *et al.* 2003a). In mesoscale and climate models, however, a vertically varying profile of soil characteristics cannot be considered as there is not three-dimensional data set in a resolution required by these models. As there are only

data sets describing the soil type at the surface, these models assume the same soil type with depth. It is obvious that such simplification will reduce the performance of any land-surface model as compared to evaluations carried out with a vertically known soil characteristic profile.

1.5 Background on Physical Processes and Parameterization of Land Models

Various offline studies and research governing on the physical processes and parameterizations have been conducted to improve the land surface models and their applications (Yang *et al.*, 1996; Loth *et al.*, 1993; Lynch-Stieglitz, 1994) further usage with general climate models (GCMs). Several key papers will be discussed, which highlight the importance of snow and frozen soil parameterization in land surface models (LSM). One of the main papers included here is from the Project for Intercomparisons of Land-Surface Parameterization Schemes (PILPS). This project was a study conducted to gain better the knowledge with respect to parameterization of interactions between the atmosphere and the continental surface in climate and weather forecast models (Henderson-Sellers *et al.* 1993, 1995). The project was subdivided into four phases; phase 1 and phase 2 involved consisted of stand alone simulations of the 21 LSMs which were driven by the synthetic and observed atmospheric forcing; while phase 3 and phase 4 investigated the coupling of Land-Surface Schemes (LSSs) with the General Circulation Models (GCMs). The experiments in Phase 2 (d) of PILPS was an experiment set to consider the effect of snow and

seasonally frozen soil on LSSs using 18 years of meteorological data from Valdai (Schlosser *et al.* 2000)

Effects of Frozen Soil on Soil Temperature, Spring Infiltration, and Runoff: Results from the PILPS 2(d) Experiment at Valdai, Russia: Lou et al. (2002)

Twenty-one LSSs of varying complexity and based on different concepts were parts of PILPS 2(d). The focus of this paper was to investigate the effect of frozen soil and its impact on soil temperature. The authors investigated the role of frozen soil while simulating the spring run off and the soil moisture along with the frozen soil scheme and further investigated the effect of snow model in LSSs over a cold region.

Soil temperature is an important factor in land and atmospheric modeling. All the models were divided into three categories according to the approach used in simulating soil frost. All the models were forced with the meteorological data. The vertical profiles of soil temperature were used to model and simulate the soil moisture.

The result from various runs using the 21 different models indicated that inclusion of soil-frost explicitly improve the simulation of the soil temperature on the broader scale. The effect on the thermal conductivity of the soil due to the inclusion of soil frost was considered to be negligible. No conclusive statement could be made about the effect of frozen soil on soil moisture in this experiment for the region of Valdai, since the top soil layer in Valdai is very close to saturation during winter and very little change in soil moisture takes place during

the snow melt period. The models based on more complex parameterization using the snow physics were able to better simulate the snow processes. The authors suggest that as land surface schemes begin to get more detailed and include the biological processes the need to simulate the soil freezing and soil temperature increases. The results obtained in this study suggest that with more explicit frozen soil model the simulations are better.

The Representation of Snow in Land-Surface Schemes; Results from PILPS 2(d): Slater et al. (2001)

In this paper, authors discussed the effect of snow on LSSs. Since 1992, the PILPS (Henderson-Sellers *et al.* 1995) has been comparing the performance of 21 LSSs. Phase 2 (d) utilized the 18 years observed meteorological data from grass-land catchments at the Valdai water-balance station 57°58'N, 33°14'E, in the former Soviet Union to run a number of simulations. The authors investigated the effect of various snow processes on various LSSs through intercomparison. All the models were evaluated against observations and to each other, the discrepancies were investigated. It was found that models were able to capture the annual cycle of snow accumulation and ablation. They found that it was not the amount of precipitation that allowed the model to simulate the observed snow cover, but the ablation events which varied with season, was a great source of bias among the various models. Over the 18-year simulations there is a clear

evidence of differences among the various models which resulted from different snow model parameterization and design.

After a systematic analysis of differences among the various models, three key processes for representing snow were identified (i) the vapor exchange at the snow surface, (ii) the amount of energy available, and (iii) the structural and thermal properties of the snow. A thin snow cover is a common source of discrepancy among the models because here small changes in energy input to the snow will melt it completely. The models showed sensitivity to long-wave radiation. This could be due to the stability induced feedback and the varying capacity of each model to exchange turbulent energy with the atmosphere. Slater *et al.* (2001) discussed the weakness in macro scale snow modeling and various parameterization schemes, for instance, the parameterization of frozen soil moisture. Investigation of differences in parameters and parameterizations of snow thermal conductivity, emissivity, and density were carried out and their influence on simulation results noted. Authors concluded that it is hard to suggest what particular combination of parameterization method is best for a given application, but the biggest sources of errors were identified, and the areas where most attention should be paid isolated to an extent.

Simulations of a Boreal Grassland Hydrology at Valdai, Russia: PILPS Phase 2(d): Schlosser et al. (1999)

This paper is one among the first series of the PILPS project. This study was part of PILPS 2(d) phase, using the hydrological data from Usadievskiy spanning 18 years. A suite of 21 LSSs participated; they are listed below at the end of this section. The main aim was to test the sensitivity to downward long-wave radiative forcing, and controls of hydrologic variability. The experimental design was planned in such a manner that the intermodel differences were minimized. Simulations were run on a set of consistent data, which were based on observed properties of the Usadievskiy catchment as much as possible. When it was not possible to obtain observational dataset for a particular parameter, then a value was set based on the overall vegetation and soil properties observed at Usadievskiy (Schlosser *et al.*, 1997), though the problem of different parameterization across LSSs persists (Polcher *et al.*, 1996). The threshold for rain-snow temperature was set at 0°C. The hydrological soil column was set to a depth of 2 m for all the models. The models were forced with the observed meteorological data of 18 years. All the models were run with the repeat of the first year, until they achieved equilibrium condition. This method was also used in other PILPS studies (Slater *et al.*, 2001; Henderson-Sellers *et al.*, 1993, 1995; Schlosser *et al.*, 1997). After reaching equilibrium all the models were integrated with 17 years of forcing (1967-1983). The outputs of the simulation for the models were requested to be daily averaged value of 24 output variables.

Another control run was performed to test the sensitivity to long wave radiation in a similar manner. The results analyzed after the completion of the simulation showed that the model scatter is present. Overall all the models showed reasonable agreement for the simulation of the root-zone soil moisture and it falls within the observed spatial variability. Few disagreements are seen in the anomalies of root-zone soil moisture. The precipitation was simulated reasonably accurately, the partitioning of the total water cycle was also well simulated; almost all the models showed a large portion of incident precipitation to evaporation rather than runoff, similar to observation. Efforts were made to improve the results of those, which showed extreme results by changing the code and parameterization. The models showed discrepancies in the phase and magnitude of the spring runoff peak, which was a result of differences in the snow ablation rates and partitioning between runoff and infiltration for run water. All models show a considerable sensitivity to the long-wave forcing for the simulation of the snow pack. Choice of snow parameterization was found to be important for the winter snow pack. Higher sensitivity to ablation rather than accumulation was found. Importance of frozen soil was underlined by the simulations with no spin up and was found that it can affect the long-term variability of the models. With spin up the runs show large biases, compared to the control integrations of the corresponding year. Recursive year runs were found to be of great relevance especially for the assessment of the LSSs interannual variations and also help in the reduction of biases. The results

indicate the importance of LSS seasonal to interannual climate application and the effects of improper LSS initialization at high latitudes.

Simulation of Freeze-Thaw Cycles in a General Circulation-Model Land-Surface Scheme: Slater et al. (1998)

Observed meteorological data collected from four sites in the former Soviet Union was used to run a LSS (Land surface scheme) model Best Approximation of Surface Exchanges (BASE; Desborough, 1997) for a period of six years. The main objective was to see how well the model is able to capture the freeze-thaw cycles. The data used for this study was collected between 1978 and 1983 for four sites namely Yershov, 51.4° N, 48.3°E; Uralsk, 51.3°N, 51.4 °E; Khabarovsk, 48.5°N, 135.2°E; Ogurtsovo, 54.9°N, 82.9°E). All of the sites were within 10 degrees of latitude but with widely spaced longitude. The vegetation type was taken to be grass. For the purpose of offline experiments, BASE was forced with meteorological data set which did not include observation for long-wave downward radiation (L_{\downarrow}). Thus L_{\downarrow} was derived.

In their experimental design for each simulation run, BASE was initialized using estimates based on the observations. A spin-up was used until equilibrium was reached. Other soil parameters included porosity, wilting level and soil type from Robock *et al.* (1995) and saturated hydraulic conductivity. BASE used three soil layers and depending on the available data the soil column was divided into 10, 40 and 50 cm slabs; the second layer being the bottom of the root zone. As

for the vegetation parameters the roughness length of 0.1 and maximum vegetation cover of 0.8 and a maximum leaf area index of 3.0 were chosen.

The model was able to simulate effectively the general characteristics of the seasonal freeze-thaw cycle for each of the stations for a period of six years. There were problems regarding the model's ability to simulate accurate amount of available soil moisture during the spring. The model tended to over-predict the soil moisture and it failed to match the timing of total freeze and total thaw. The authors pinned the problem down to hydraulic conductivity and when that was modified the results were more accurate for the soil moisture. This paper showed that the inclusion of frozen soil parameterization is important to better simulate the available soil moisture and if not accounted for, the errors could extend beyond the spring-thaw season. The authors suggest that the simulations can be further made more accurate by including effects like freezing fronts, ice-lensing and other periglacial processes.

Validation of the Snow Submodel of the Biosphere-Atmosphere Transfer Scheme with Russian Snow Cover and Meteorological Observational Data: Yang et al. (1996)

This paper among others discusses the effect of snow cover simulations using a snow submodel in Biosphere-Atmosphere Transfer Scheme (BATS; Dickinson *et al.*, 1993), using detailed snow measurements. The model's ability to simulate the snow cover at different stations is examined. The snow data was obtained from

meteorological observations in the former Soviet Union for six stations from 1978-1983. Each of these stations was located on a grass-covered plot. BATS used datasets of Olson *et al.* (1983), Matthews (1983) and Henderson-Sellers (1985) for the 18 classes of land cover. The various parameters related to the various vegetation kinds i.e. physical and physiological properties were used. The soil type used was loamy soil. The hydraulic properties were based on Clapp and Hornberger's (1978) empirical relationship for soil water potential and hydraulic conductivity. The snow submodel in BATS assumes one-layer snow cover with a time varying snow depth, snow density and snow albedo. Every three hours, the age of snow and the amount of grid covered by snow was calculated. These were then used to calculate the thermal conductivity, volumetric specific heat of snow and the composite soil/snow layer were derived. The force-restore method (Dickinson 1988) was used to calculate the temperature between the interface of soil and snow. The snow model is more computationally efficient and simpler than those used by Loth *et al.* (1993) and Lynch-Stieglitz (1994). The authors wanted to evaluate the model before making any changes to the equations and parameterization of the snow physics processes. The rain-snow transition temperature was set at 2.2°C. The model was then set at 50% of soil moisture capacity and both snow depth and snow age were set to zero at initialization. Long-wave radiation was parameterized based on Robock *et al.* (1995) who used a modified version of the Monteith formula to estimate the downward long-wave radiation. The scheme used was

based on Satterland (1979). The model integration was allowed to reach equilibrium with the given initial soil moisture. This was achieved by driving the model a repeat of first year's data multiple times, and then using 6 years data to drive the model.

Yang *et al.* (1996) discussed two major findings, first the role of wind correction for the precipitation data, and this was also applied to the rain-snow transition criterion. In absence of the correction factor, the model is able to simulate the snow water equivalent and surface temperature of snow to a reasonable accuracy for all the six winters under consideration, especially, the time of accumulation and end of ablation. The density change and metamorphism of snow was also well captured. With some alterations to the parameterization schemes and modification of the code, the model was able to simulate the snow cover fraction, snow depth, and surface albedo efficiently. After the wind correction was applied to the gauge-measured precipitation, the model showed increased root mean-square errors for all the stations except one. The errors in the model were explained by the possibility of snow being blown away from the area of measurement. One of the reasons for discrepancy could be that with stronger wind the snow gage is not able to capture all the precipitation and also at the same time less snow might remain on the open sites. The less precipitation and less snow on the site itself cancel each other and hence there might be no need for correction of wind in the nearby areas but there might be need for it in larger area where the snow accumulates. The author

contributed and warranted special attention to four major factors namely the parameterization of long-wave radiation and snow cover fraction, the snow metamorphism process for determination of snow density and change in albedo and choice of critical temperature for rain-snow transition.

CHAPTER 2

DATA ACQUISITION AND SITE DESCRIPTION

2.1 The Founding of NOPEX (Northern hemisphere climate Processes land-surface Experiment) / WINTEX (WINTER EXperiments) Project

Recent years large efforts have focused on a better understanding of climate processes and climate change (Halldin *et al.*, 2001). One of the main areas of interest is predicting climate change using various models (Anderson, 1976; Jordan, 1991; Loth *et al.*, 1993; Lynch-Stieglitz, 1994). For better predictions, there is acute need for reliable year-round observational data, which would indeed help in reducing considerable uncertainties in models. In the past, practical and financial constraints have prevented a detailed observational dataset from being set up. The WINTEX projects, which are a subset of the Northern hemisphere climate Processes land-surface Experiment (NOPEX) project, were among the first large-scale European research projects set up to study the various land-surface processes at a regional scale to characterize by arctic weather conditions (Halldin *et al.*, 2001).

2.2 NOPEX/WINTEX

NOPEX, was part of BAHC (Biospheric Aspects of the Hydrological Cycle), which also coordinated the BOREAS (Boreal Ecosystem-Atmosphere Study), the HAPEX-SAHEL (the Hydrology-Atmosphere Pilot Experiment in the Sahel). The

main objective of the NOPEX project was to study the land-surface processes, e.g. exchange of various fluxes in an environment dominated by long winter, snow, low solar angles and land cover which comprised of boreal forest (Halldin *et al.*, 2001), on daily as well as an annual basis. NOPEX intended to provide improved parameterization schemes regarding the exchange of water, energy and carbon between the land surface and the atmosphere for the meso-scale to the global scale meteorological models. NOPEX started its field activities in 1994 with two different programmes i.e. Continuous Climate Monitoring (CCM) and the Concentrated Field Effort (CFE) (www.hyd.uu.se/nopex/). The CCM, which continues to operate, was established for studying the various processes during a long time interval. The credibility of various cycles of water, carbon, and energy can be verified against observational data over a longer time interval, along with calibration and testing of various models from long-term observation series. Other tasks of CCM, include a comparison of mass and energy balance using different methods, evaluation of climatic conditions and the collection of year round, including the hard winter months

The other field program named CFEs were carried out during the spring and summer of 1994 and 1995 (Halldin *et al.*, 2001); the activities involved consisted of intensive studies covering an area of around 100 by 50 Km².

The objective of CFEs involved local and regional scale studies, remote sensing, development of remote-sensing methodologies for use in modeling community, to test and develop measuring techniques, atmospheric and

hydrological modeling in relation to the WINTEX project. The first set of observations was recorded during the spring and summer season. On successful collection of data during that season the NOPEX group decided to carry out a concentrated field effort during wintertime. Keeping in mind the practical problems along with the difficult winter conditions, only a limited field experiments to test various assumptions and measurement techniques were established. The subproject within NOPEX under the CFEs framework was named "WINTEX-Land surface-atmosphere interactions in a winter-time boreal landscape planning of a NOPEX winter-time CFE", and was launched on December 1996 and commenced in December 1998 (Halldin *et al.*, 2001). WINTEX was constructed with the aim of providing the much sought after full-scale winter time data.

2.3 NOPEX/WINTEX Region

The WINTEX experiment was a pilot study, which was initiated with the intention of understanding the various processes regarding land-surface-atmosphere exchanges of water, energy and carbon during the winter time. The initial aim was to establish only one site for investigation at Marsta, Sweden, but the variable winter conditions led to the probability of no snowfall event taking place and this eventually led to the establishment of another site in the northern region of NOPEX around the Sodankylä Meteorological Observatory. The two sites are the southern and northern boundary of the boreal forest.

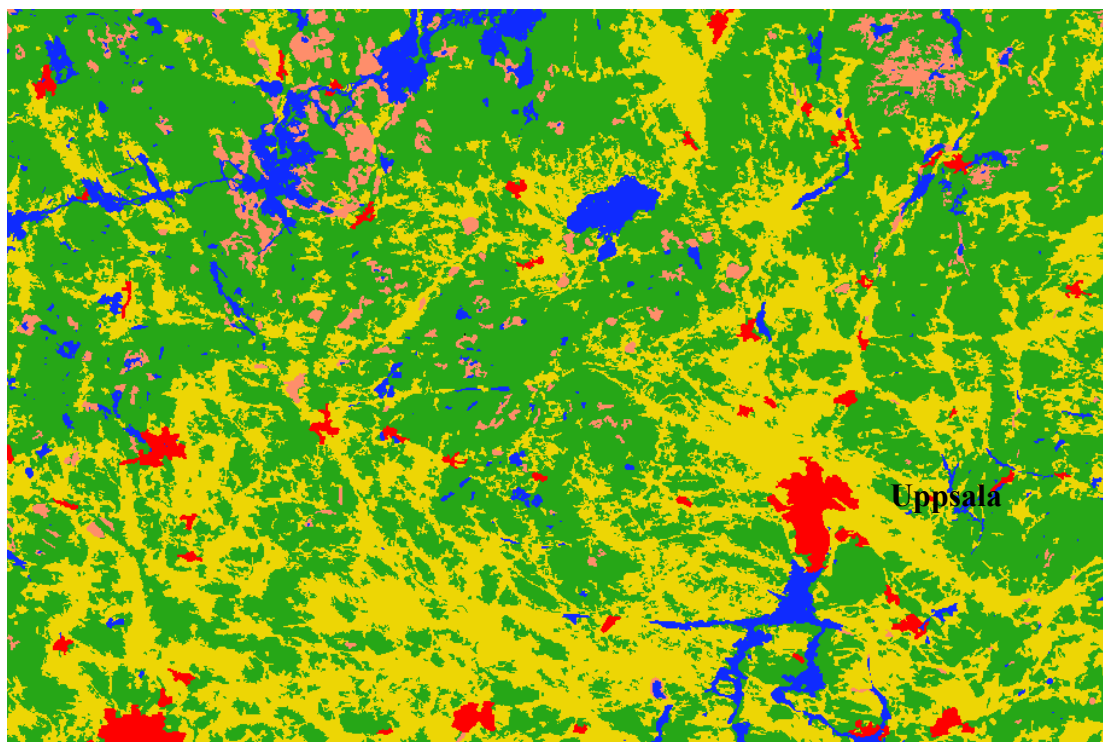


Figure 2.3.1

The Land use picture depicting the southern part of NOPEX region

[www. hyd.uu.se/nopex](http://www.hyd.uu.se/nopex)

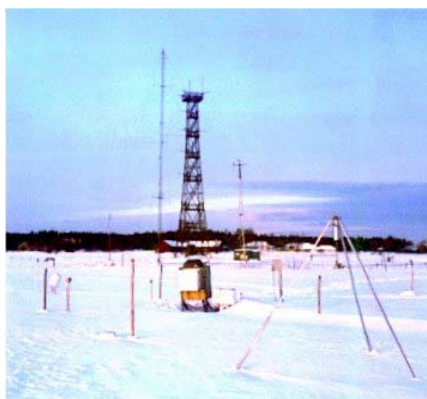
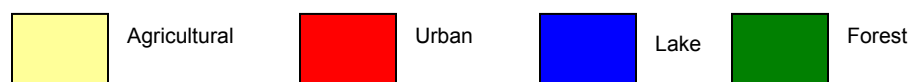


Figure 2.3.2
Marsta 30 m flux mast

2.4 Southern NOPEX/WINTEX Region: Marsta

The southern part of the NOPEX region is situated in Central Sweden near Uppsala, 9 km from Marsta. It was chosen because of its flat landscape and homogeneity in terms of a mixed boreal-forest/agricultural landscape

It is evident from figure 2.3.1, that land use is very heterogeneous with the majority comprising of forests followed by agricultural fields (www.hyd.uu.se/nopex). Most of the forests are found in the northern part of the field study and the southern part contains more agricultural land. Agriculture has a history that dates back to more than 1000 of years and the forest has been managed in a systematic manner for the last 200 years. The forest mainly comprises of spruce and pine with a small portion of deciduous forests (15%). The topography varies between 0-188 m above the sea level; the distance to the sea is about 50 km which avoids the complications due to land sea circulations. The predominant soil type is clay in the south; other soil types found are sand and silt in the northern part. The geology mainly comprises of granite, sedimentary gneiss, and leptite. Lakes and wet areas are found to be an important part of the landscape with more areas of bogs and moraines in the northern part. Three main run-off regions consists of Lake Mälaren in the south, two smaller run-off regions were River Dalälven and the Bothnian Sea. The whole NOPEX southern region is situated at the southern part of the boreal forest zone but compared to other boreal forest regions the climate tends to be more maritime.

2.4.1 Marsta Meteorological Observatory

The site description is described by Gustafsson *et al.* (2000). Marsta Meteorological observatory is situated at 59°55'N, 17°35'E about 9 km north of Uppsala, Sweden.



Figure 2.4.1.1

Map of Sweden showing the site of Marsta

The annual precipitation of the site is around 520 mm and the temperature averages about 5.5 °C. The dominant soil type is loamy clay with an organic content of 6% in the topsoil and about 2.5% in the subsoil. The site is located on an agricultural land consisting of two sectors one on the east side and one on the west side having the dimensions of 1000-4000 m and other dimension is about 500 m going from the north to the south. As a part of the soil management program, the field was ploughed prior to the first winter, but no soil management was done prior to the second winter. The data set covers two winters from November 1997 to April 1999.

2.4.2 Data Collection and Meteorological Observations for Marsta

The meteorological observations included air temperature, wind speed, humidity, downward and upward short-wave and upward long- wave and net radiation. Precipitation was collected at the site itself during the spring, summer and the autumn season as for the winter season the data for precipitation was obtained from central Uppsala. There was slight error in the observed precipitation data. It was found that observed precipitation was slightly less than the observed snow accumulation. This was attributed to a snowstorm with an S-N gradient (Gustafson *et al.*, 2001), also it is possible that the equipment used for precipitation was malfunctioning and was not able to catch the precipitation event accurately. The timing of the precipitation was adjusted by comparison of the

hourly readings and a daily observation of precipitation at a manually maintained station in Uppsala (Gustafson *et al.*, 2001).

An ultrasonic ranging sensor was used to measure snowfall events and soil temperatures were measured using copper-constantan thermocouples in two sets of profiles at depths of 5, 10, and 20 cm. It was observed that around 5% of data were missing and this was rectified by replacing the observations collected from meteorological station in central Uppsala. For long wave radiation missing observations data were simulated using the parameterization of Konzelmann *et al.* (1994) as a function of air temperature, vapor pressure, and cloudiness.

2.5 Northern NOPEX Region: Sodankylä

During the second half of the CFEs, the experimental campaign in the northern part took place between 12th March and 19th April, 1997. The northern part of the NOPEX region is essentially northern part of Finland. The location of Sodankylä the winter season is the longest lasting around seven months (<http://worldfacts.us/Finland-geography.htm>). Finland is surrounded by the Atlantic Ocean in the west and the Eurasian continent in the east. These two bodies interact to influence the climate of the country. The warm Gulf Stream and North Atlantic Drift current help Finland to moderate the winter temperatures. Contradicting the maritime influence is continental high-pressure systems in the Eurasian Continent, which result in severe winter conditions and occasionally extreme warm conditions during the summer season. The landscape around the

northern NOPEX region is fairly flat with isolated hills reaching up to 500 m altitude. The forest region contains mainly coniferous trees, with pine being the most dominant of all species. Mixed/deciduous forests account for about 13% of the land cover and the rest consists of sparsely vegetated peat lands.

2.5.1 Sodankylä Meteorological Observatory

The Sodankylä Meteorological Observatory (SMO) is located in the sub-arctic region of northern Finland situated at 67°22' N, 26°39' E at an elevation of 179 m above sea level and is run by the Finnish Meteorological Institute (Tony Persson, 2004). Sodankylä usually has sub arctic climate and is 100 km north of the Arctic Circle. It has long and cold winters with short and warm summers. Characteristic of the arctic region, Sodankylä has low solar elevations and short days during the winter, resulting in negative radiation balance and stable atmospheric conditions. Snow data and micrometeorological data (Heikinheimo *et al.*, 2001) were measured as a part of the WINTEX campaign

2.5.2.1 Data Collection and Meteorological Data

The meteorological data collected at SMO, consisted of air temperature and relative humidity, net radiation at 2 m, downward long-wave radiation was measured at 16.8 m, reflected shortwave at 2 m and wind speed at 22 m. Various types of precipitation events were recorded along with cloud cover observations. A 20 m patch within a sparse coniferous forest was used to collect

observational data for vertical profiles of snow density and temperatures. The height of the trees surrounding the observational patch varied from 4-6 m to 6-15 m at a distance of about 50 m.



Figure 2.5.2.1.1

Radiometric and climatologically measurements at Sodankylä

The snow density profile was measured by weighing cylindrical snow samples having a 0.1 m vertical resolution at a distance of about 10-20 m from snow temperatures sensors. Measurements for snow temperatures were observed using eight thermocouples (Chromel-Constantan) which were installed at heights between 0.1 and 0.9 m above the ground. For measuring the temperature at the snow surface two infrared sensors (Everest Inter Sci. Inc., model 4000) were used and a Campbell 21X logger routinely recorded one-minute measurements

and stored hourly mean values. During precipitation event the snow depth, snow water equivalent and snow density profiles were manually measured.

CHAPTER 3

MODEL DESCRIPTION

3.1 History

Hydro-thermodynamic soil-vegetation scheme (HTSVS) is a land surface model used in this study. The model was first developed at the University of Frankfurt by Dr. G. Kramm (Kramm and Herbert 1984, Kramm 1987, 1995, Kramm *et al.* 1996). The model has undergone further development and changes. The changes include the effect of soil-frost and root effects (Mölders *et al.* 2003a) and snow metamorphism (Mölders and Walsh 2004) on the hydraulic cycle. Mölders and Walsh (2004) replaced the simple one-layer snow model that was used for the study by Mölders *et al.* (2003a) by a modified version of the multi-layer snow developed by Fröhlich and Mölders (2002); who used force restore method and implemented the snow model that was later modified for use in HTSVS. A multiple snow layer instead of a single layer was considered because it was found through various comparisons with other snow models that the number of snow-layers influences the performance of snow models (Slater *et al.*, 2000).

HTSVS coupled with an air chemistry model was applied for various air quality and air pollution studies (Kramm *et al.* 1994, 1996). An integrated modeling approach to simulate the water cycle and to study the atmospheric response to land-use change was performed with HTSVS incorporated in a non-hydrostatic meso-beta/gamma-scale meteorological model (Mölders and Rühaak 2002); the results from the integrated modeling technique could be used to

predict the supply of water under altered conditions. HTSVS including the process of soil freezing and thawing along with snow-metamorphism was integrated in the Penn State/NCAR Mesoscale Meteorological Model MM5 in a two-way coupled mode (Mölders and Walsh 2004) to examine the impact of permafrost and snow on weather.

3.2 Brief Description of the Model

HTSVS is a land surface model designed for use in numerical weather prediction (NWP) or General Climate models (GCM). HTSVS was developed initially to investigate and study the various fluxes mainly the water and energy fluxes at the biosphere-atmosphere interface. Various applications of HTSVS have been made (see previous section). HTSVS considers one canopy layer, multiple snow and soil layers. HTSVS includes 16 soil type including moss and lichen which are of great importance when considering the moisture distribution within the soil and for permafrost dynamics. HTSVS includes the following features: -

- The various fluxes at the vegetation-soil interface are included. The exchange of momentum, water vapor, and heat are the main fluxes that are considered.
- The mixture approach, also called the Deardorff-mixture approach, considers the heterogeneity at the microscale i.e. it considers that a grid cell can be partly covered with vegetation (Deardorff, 1978;Kramm *et al.* 1996).

- Cross-effects within the soil layers i.e. Ludwig-Soret-effect and Dufour effect (de Groot, 1995; Prigogine, 1961; Kramm *et al.* 1996, Mölders *et al.*, 2003a) are considered in the model. Ludwig-Soret-effect is a cross effect in which the temperature gradient in the soil results in the development of moisture gradient, this effect is important since it changes the soil volumetric content. The Dufour effect on the other hand refers to the moisture gradient in soil, which causes a temperature gradient to develop; this effect alters the soil temperature. These two effects can be of relevance during soil freezing and thawing process.
- Soil freezing and thawing are included in the model; these two processes have demonstrable impact on the thermal regime of the model (Flerchinger and Saxton, 1989; Mölders *et al.*, 2003a).
- Water uptake by roots and the vertical profile of the root is taken into account.
- Snow metamorphism and the temporal evaluation of the snow density and snow depth are incorporated in the model. The insulating effects of snow on the soil temperatures and water retention are considered.
- Variation of soil albedo, snow albedo and snow emissivity with time (Mölders *et al.*, 2003a).
- The three phases of water: vapor, ice, and liquid water.

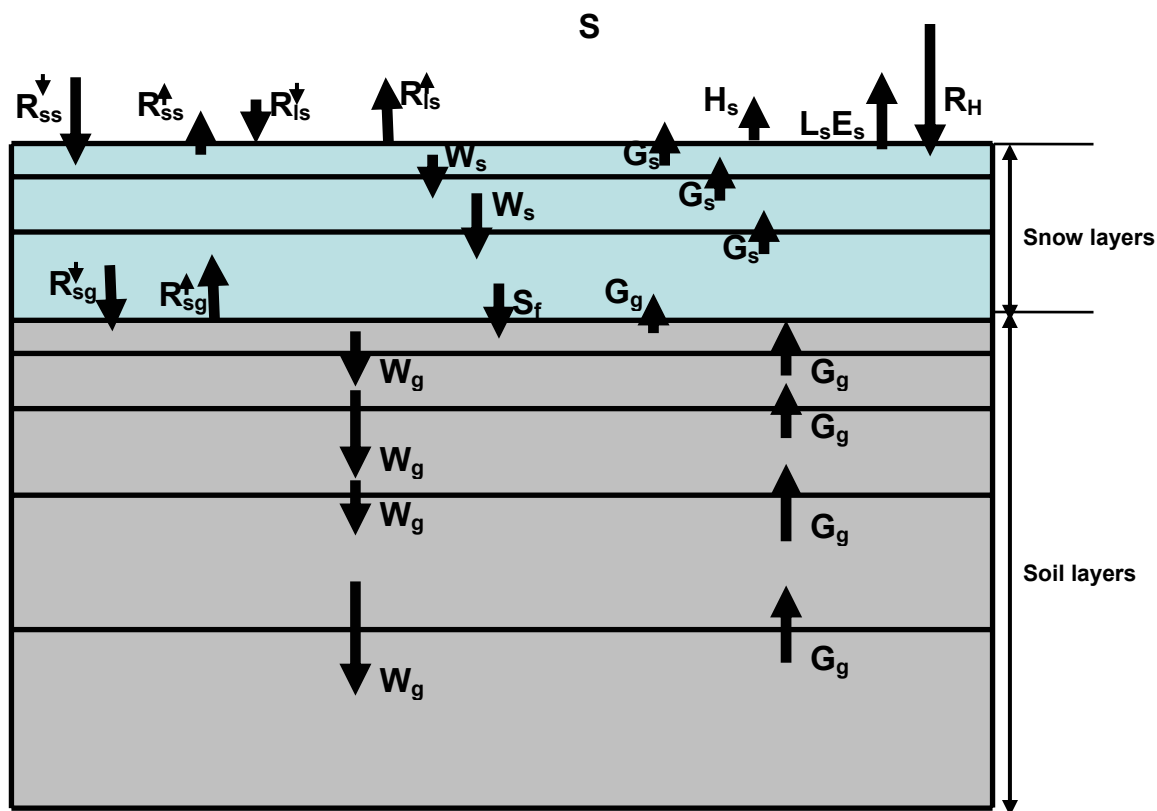


Figure 3.2.1
Schematic view of the design of the snow and soil frost module of HTSVS
<http://www.gi.alaska.edu/~molders>

3.3 Equations and Assumptions

3.3.1 Main equations for the soil-vegetation scheme

In HTSVS, the equations of energy and mass fluxes relate the exchange at atmosphere-vegetation interface. The equations are parameterized using a resistance approach, which is analogous to Kirchhoff's law of electrostatics (Kramm 1995; Mölders 1999). The resistance network analogy is schematically shown in figure 3.2. This approach helps in determining the soil moisture, soil temperature, and snow temperature along with the fluxes.

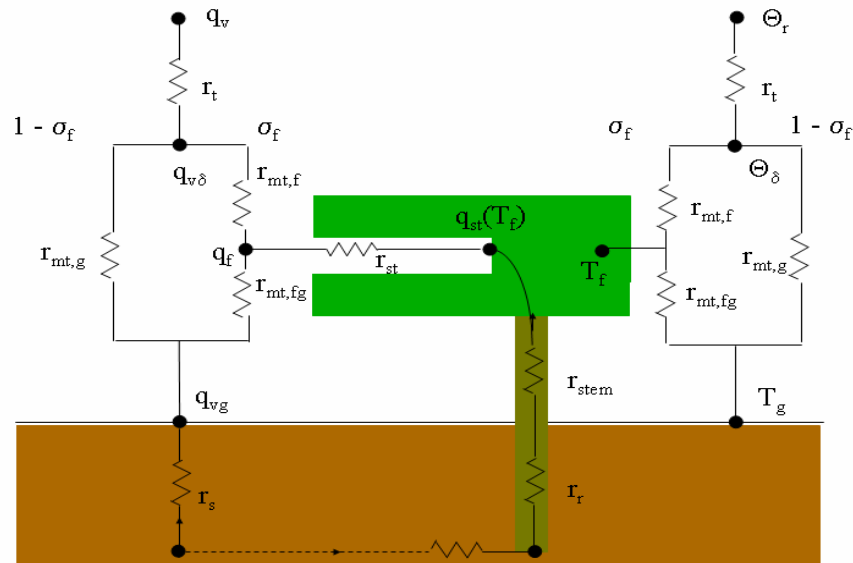


Figure 3.3.1.1
Schematic of the resistance approach

- Sensible fluxes (H) at the surface of the foliage (f) and soil (g) (Kramm *et al.* 1996) are listed from equation 3.3.1 to equation 3.3.4

$$H_f = -\sigma_f c_p \rho_a \left\{ \frac{1}{r_{mt,f}} (\Theta_\delta - T_f) - \frac{1}{r_{mt,f}} (T_f - T_g) \right\} \quad (3.3.1.1)$$

H = Sensible heat flux

f = subscript stands for foliage

g = subscript stands for ground

σ_f = Shielding factor

C_p = Heat capacity

ρ_a = air density

r = resistance and the subscript mt stands for molecular turbulence and fg stands for foliage and soil surface

Θ_δ = Potential temperature and the suffix δ stand for height close above the foliage

T = Temperature

$$Hg = -c_p \rho_a \left\{ \frac{1-\sigma_f}{r_{mt,g}} (\Theta_\delta - T_g) + \frac{\sigma_f}{r_{mt,fg}} (T_f - T_g) \right\} \quad (3.3.1.2)$$

- Water vapor flux at the foliage and the ground

$$E_f = -\sigma_f \rho_a \left\{ \frac{1}{r_{mt,f}} (q_\delta - q_f) - \frac{1}{r_{mt,fg}} (q_f - q_g) \right\} \quad (3.3.1.3)$$

E = Water vapor flux

q = specific humidity

$$E_g = -\rho_a \left\{ \frac{1-\sigma_f}{r_{mt,g}} (q_\delta - q_g) + \frac{\sigma_f}{r_{mt,fg}} (q_f - q_g) \right\} \quad (3.3.1.4)$$

- Coupled energy-and water budget equations used to calculate the values of temperature and moisture at the surface of the foliage and soil

$$R_{ss} \downarrow - R_{ss} \uparrow + R_{ls} \downarrow - R_{ls} \uparrow - H_s - L_s E_s + G_{snow} + P_H = 0 \quad (3.3.1.5)$$

R_{ss} = Shortwave radiation downward

R_{ls} = Longwave radiation downward

G_{snow} = Heat flux inside snow

$L_s E_s$ = Latent heat of sublimation

P_H = Input of heat into the snow pack due to rain

$$P + S - E_s = 0 \quad (3.3.1.6)$$

P =Precipitation

S =Precipitation in solid form

$$R_{ss} \uparrow = \alpha_{snow} R_{ss} \downarrow \quad (3.3.1.7)$$

α_{snow} = albedo

$$R_{ls} \uparrow = \epsilon_{snow} \sigma T_{snow, surf}^4 + (1 - \epsilon_{snow}) R_{ls} \downarrow \quad (3.3.1.8)$$

ϵ_{snow} =Emissivity

3.3.2 Snow-atmospheric interaction

Analogously the equation for the latent and sensible heat fluxes at the snow surface are given by

$$H_s = - \frac{c_p \rho_a}{r_{mt, snow} + r_t} (\Theta_R - T_{snow, surf}) \quad (3.3.2.1)$$

$$E_s = - \frac{\rho_a}{r_{mt, snow} + r_t} (q_R - q_{snow, surf}) \quad (3.3.2.2)$$

- The snow heat flux is

$$G_{snow} = -\lambda \frac{\partial T_{snow}}{\partial Z_{snow}} - L_v \rho_w k_v \frac{\partial q_{snow}}{\partial z_{snow}} \quad (3.3.2.3)$$

λ = Heat conductivity

Z_{snow} =Height of snow

L_v =Latent heat of vaporization

K_v =Molecular diffusion coefficient of water vapor within air-filled pores of the snow-pack

- The equation to calculate the temperature and moisture at the surface of snow and soil is

$$R_{sg} \downarrow - R_{sg} \uparrow + G_g - G_{snow,g} = 0 \quad (3.3.2.4)$$

$$S_F + W_{soil} = 0 \quad (3.3.2.5)$$

S_F = Fluxes in soil

W_{soil} = Vertical transfer of water vapor fluxes across the soil layers

$$R_{sg} = R_{ss} \exp(-k_{ext} z_{snow}) \quad (3.3.2.6)$$

k_{ext} = Extinction coefficient

$$R_{sg} = R_{sg} \alpha R_{sg} \quad (3.3.2.7)$$

3.3.3 The Soil Model

HTSVS calculates the soil flux (soil moisture flux and the energy flux coming in and out of soil) in the soil. Various methods are available which compute the temporal variation of soil moisture (e.g. Philip 1957); these methods have been incorporated in various models in the past for e.g. Sasamori (1970) and Garrett (1978). Recent research has demonstrated the importance of accurate soil moisture values in the numerical weather prediction especially the effect it has on the water balance. The equation derived for the water flux derived below is based

on principles of the linear thermodynamics of irreversible processes including the Richards-equation (Philip and De Vries 1957, Philip 1957, de Vries 1958, Kramm 1987, 1995, Kramm *et al.* 1994, 1996, Mölders 2003a). The local time rate of change of the volumetric moisture content is given as:

$$\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z} \left((D_{\eta,v} + D_{\eta,w}) \frac{\partial \eta}{\partial z} + D_{T,v} \frac{\partial T}{\partial z} + K_w \right) - \frac{\chi}{\rho_w} - \frac{\rho_{ice}}{\rho_w} \frac{\partial \eta_{ice}}{\partial t} \quad (3.3.3.1)$$

η = moisture content

$D_{\eta,v}$ = Transfer coefficients with respect to water vapor

$D_{\eta,w}$ = Transfer coefficients with respect to water

$D_{T,v}$ = Transfer coefficients with respect vapor

χ = Water uptake by roots per soil volume

$$\chi = \frac{X}{\Delta z} \quad (3.3.3.2)$$

$$C \frac{\partial T_s}{\partial t} = - \frac{\partial}{\partial z} \left(-(\lambda + L_v \rho_w D_{T,v}) \frac{\partial T_s}{\partial z} - L_v \rho_w D_{\eta} \frac{\partial \eta_{ice}}{\partial z} \right) + L_{ice} \frac{\partial \eta_{ice}}{\partial t} \quad (3.3.3.3)$$

$$C = (1 - \eta_s) \rho_s c_s + \eta \rho_w c_w + \eta_{ice} \rho_{ice} c_{ice} + (\eta_s - \eta - \eta_{ice}) \rho_a c_p \quad (3.3.3.4)$$

C = Volumetric heat capacity of moist soil

$$\psi = \psi_s \left(\frac{\eta_s}{\eta} \right)^B = \psi_s W^{-B} \quad (3.3.3.5)$$

ψ = Water potential

ψ_s = Water potential at saturation

3.4 Main Equations of the Snow Model

There is an increase in snow height due to precipitation in formation of snow, the decrease in snow height is caused by various processes i.e. sublimation, increase of snow density by windbreak, compaction and freezing (Mölders and Walsh 2004). Fröhlich and Mölders (2002) did not consider sublimation, as it is not very relevant in the midlatitude. The redistribution of snow depth by blowing snow was also neglected, assuming that the associated variation in snow depth will average out over the size of a model grid-cell. Snow density depends on the volumetric water content and the porosity of the snow (Dingman, 1994) and is given as:

$$\rho_{snow} = (1 - \phi)\rho_{ice} + \theta\rho_w \quad (3.4.1)$$

The water equivalent of snow which is nothing but the depth of water that would result after complete melting of the snow-cover (Dingman, 1994):

$$h_w = h_{snow} \frac{\rho_{snow}}{\rho_w} \quad (3.4.2)$$

The effective porosity of the snow (Dunne *et al.* 1976) is:

$$\Pi = \frac{\rho_{snow} - \rho_{ice}}{\theta_{ret}\rho_w - \rho_{ice}} \quad (3.4.3)$$

Rate of change of snow depth due to sublimation is:

$$\frac{\partial h_{snow}}{\partial t} = \frac{E_s}{\rho_{snow}} \quad (3.4.4)$$

Rate of change in snow density by compaction is calculated by (Anderson, 1976):

$$\frac{1}{\rho_{snow}} \frac{\partial \rho_{snow}}{\partial t} = C_1 \exp(-0.08(T_0 - T_{snow})) W_{snow} \times \exp(-C_2 \rho_{snow}) \quad (3.4.5)$$

The equation for destructive metamorphism is given as:

$$\frac{1}{\rho_{snow}} \frac{\partial \rho_{snow}}{\partial t} = C_3 \exp(C_4(T_0 - T_{snow})) \quad (3.4.6)$$

Percolation through matured snow layer is:

$$J = \rho_w \theta k_w \quad (3.4.7)$$

CHAPTER 4

RESULT AND DISCUSSION

4.1 Introduction

The simulation was carried out by using HTSVS for the two sites of the WINTEX project. WINTEX took place in the Boreal/Arctic regions in 1996-98. These studies enhanced our knowledge of the various water and energy fluxes of the region around Sodankylä (Finland) and Marsta (Sweden).

The hydro-thermodynamic soil-vegetation model HTSVS was initialized using observed soil temperatures, snow temperature and densities. As typically precipitation refills the soils reservoir in fall, and since there are hardly sinks for soil moisture after a snow pack is built up, the soil is assumed to be saturated. This means that all soil pores are filled. The partitioning between the liquid and soil phase of the total soil volumetric water follows Mölders and Walsh (2004), i.e. frozen ground is taken into account when initializing HTSVS.

Hourly data of observed shortwave downward radiation, air temperature, specific water vapor, wind speed, and pressure as well as daily precipitation data served to drive HTSVS (see Tab. 4.1). Observed data of downward long wave radiation data were not used in this study as they turned out to be erroneous due to recording problems. Therefore, long-wave downward radiation was alternatively parameterized in accord with: (a) Croley by

$$H_{lp\downarrow} = 6.1078 \exp(17.269 * (T_{2M} - 273.15) / (T_{2M} - 35.86)) \quad , \quad (4.1)$$

In the past equation by Bolz and Falkenberg by have given appropriate result

(b) Bolz and Falkenberg by

$$R_{lf\downarrow} = R_{ls\downarrow} = \sigma T_a^4 (0.52 + 0.065 e_a^{0.5}) (1 + 0.22 c^2) \quad , \quad (4.1)$$

and Idso and Jackson (1969) by

$$R_{ls\downarrow} = \frac{\sigma}{\epsilon_s} \left\{ (\epsilon_s - 1) T_s^4 + (\epsilon_a (1 - c) + c) T_a^4 \right\} \quad , \quad (4.2)$$

with

$$\epsilon_a = 1 - 0.261 \exp\left(-7.77 \cdot 10^{-4} (273 - T_a)^2\right) \quad , \quad (4.3)$$

In nature, soil-type usually changes with depth. In principle, HTSVS can consider different soil-types with depth (see Mölders et al. 2003a) if the soil-type profiles with depth are available. For these two sites, no information on the vertical soil profile was available. Since typically numerical weather prediction (NWP) models rely on soil-type data that assign one representative soil-type for an area of several square kilometers without consideration of any variation of soil-type with depth, performing an evaluation without vertical variation of soil-type allows an evaluation closer to NWP reality than using site specific soil data. Note that no three dimensional data sets of soil-type data exist at the typical resolution of NWP models. Sensitivity analysis was carried out by varying various parameters to see the difference in the results.

Table 4.1. Detailed information on the forcing data for the two stations

Site	Latitude, Longitude	Forcing data	Data used for evaluation
Sodankylä, northern Finland	67° 18'N 24° 24' E	Long wave downward radiation (parameterized), short-wave downward radiation, wind speed, air temperature, specific humidity, air pressure, and ground temperature at 2m depth, precipitation (observed)	Soil temperature, snow height
Marsta, central Sweden	60°08'N 17 °47' E	Long wave downward radiation, short-wave downward radiation, wind speed, air temperature, specific humidity, air pressure, and ground temperature at 2m depth, precipitation (observed)	Soil temperature

4.2 CASE 1: Sodankylä

The first case analyzed the region of Sodankylä for 75 days starting March and ending on May. The model was initialized by using the initial values given in figure 4.2.2 for this particular case. In a first step to evaluate HTSVS the simulated and observed snow height were compared. In the second one, HTSVS's ability to simulate soil temperature was evaluated by running simulations wherein alternatively the soil-types were changed. Since observations existed for soil depths of 5 cm, 10 cm, 20 cm, 30 cm, 50 cm, 70 cm, 100 cm, 150 cm, 200 cm and 300 cm and HTSVS provides soil temperatures at 50cm, 126 cm, 316 cm, 796 cm and boundary condition is 200 cm, following

procedure was applied. The observed soil temperature values were assumed to be representative for the soil model-layer that had the closest distance between the level, for which soil temperature was predicted and the level of observations i.e. if, for instance, the depth, for which soil temperature was simulated, was 12.6 cm, the simulated value was compared to the soil temperature observed at 10 cm, Since the difference in depth was not large, it was assumed that the temperature will not vary much over this short distance.

As pointed out above, for this station the long wave radiation was parameterized. The simulation results were found to be very sensitive to parameterized long-wave radiation Simulation results also were found to be very sensitive to initialization, especially of soil volumetric water content and snow density.

Figure 4.2.2 is a comparison of simulated and observed snow height in meters. The simulated snow height does not perfectly match the observations except in few points. In the first part of the simulation episode it underestimates snow height slightly. This means that the snow model predicts a slightly too strong reduction in snow height by snow metamorphism. However, predicted snow height compares favorably with the observed variations. One of the reasons for a seemingly under prediction of the snow height may also be that HTSVS was driven by observed precipitation data. It is well known that snowfall is hard to measure as catch, leading to lower measured values then what really happened. They may exceed 30% (Dingman 1994). Another reason is that

longwave downward radiation may be also affected by cloud cover. Since no cloud cover data were available, this effect had not been included in the parameterization of longwave downward radiation. Wind can also reduce snow height and this process is included in the model. However, snow advected horizontally by wind (snowblow) in nature which could lead to higher accumulation of snow in reality. This process is not included in the model. In nature, this process can increase snow height if more snow is transported towards the site than transported away.

The snowmelt rates are over estimated by the model. The snow model uses a parameterization for snow albedo that is based on albedo data in the fifties. Since fewer filters were used at time of observations, the model may predict a higher albedo value than occurred in nature. Consequently, the snow pack tends to be persist longer than in the reality. Since the snow pack is thicker it tends to have stronger insulation. Since soil albedo depends on soil volumetric water content, incorrect predictions of soil volumetric water content or errors in initializing soil volumetric water content, will lead to errors in soil albedo that affect the simulated snowmelt.

After all snow is melted in the model world, there is a sudden change in the ground surface from snow to grass and in the model the change is faster than in reality indirectly also affecting the energy balance which can accelerate the melting process in the soil. In nature, it takes some time until grass greens up. Thus, immediately after snowmelt, HTSVS like all state-of-the-art soil models

uses an unrealistic albedo value. The root mean square value for the snow height was calculated (Figure 4.2.2) and the percent error was around 8%. Thus, the snow height predictions must be considered as good.

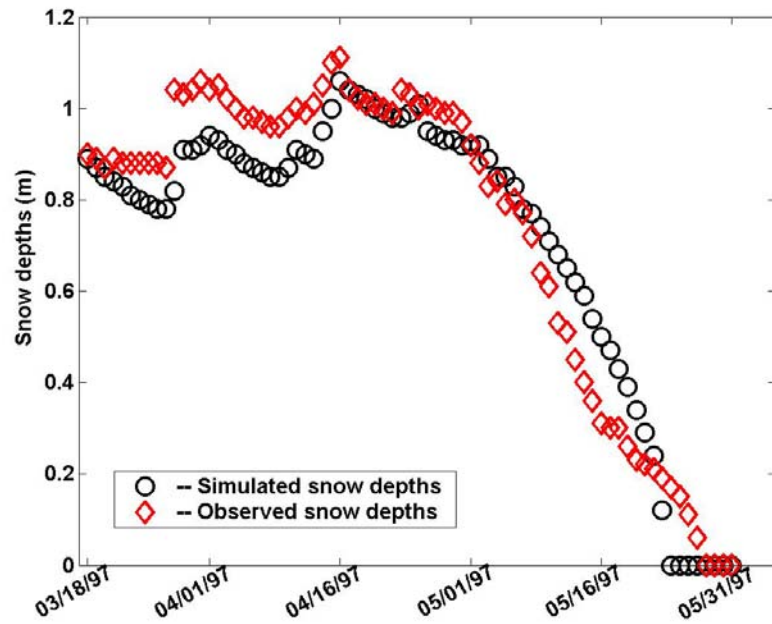


Figure 4.2.1
Graph of simulated vs. observed snow height for the region of Sodankylä

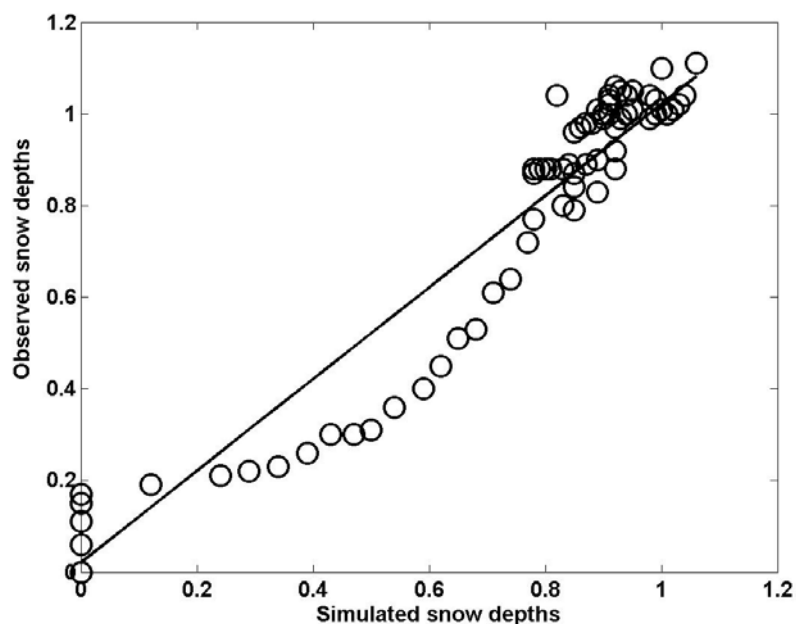


Figure 4.2.2

Scatter plot for snow height for the region of Sodankylä

The soil temperature was compared for various layers as mentioned earlier. The soil type for Sodankylä was sandy podzol. The results obtained indicated that different soil types gave the best fit for different layers of the soil. The RMS for all the different soil types was investigated. The soil type with the minimum RMS error was considered. The RMS error shows a decreasing trend from the upper layer going down to the lower layer for all the soil types. The RMS table is shown with the graphs and the yellow highlight shows the soil type chosen for each depth. These results emphasize the need for a three-dimensional data set for soil type data for use in NWP models. Note that HTSVS when used in an atmospheric model can run 3-D simulations with different soil types in the various

soil layers but there is a lack of dataset. The trend in the graphs can be briefly stated as:

- The first soil layer the best fit soil type was loam which is close to sandy podzol. The soil type sandy podzol cannot be directly translated to USGS soil types. The simulated data shows under prediction for first phase of the simulation and towards the end of the episode the temperature is over predicted. The reason for under prediction could be related to under prediction of snow pack. Less snow pack would mean lesser insulation compared to observation and the cool air temperature penetrates more readily in the soil resulting in cooler temperatures. Towards the end of the episode when the snowmelt season sets in, the temperatures are over predicted. The over prediction could be due to more insulation of snow which prevent insulates the soil from exposure to the air temperature.
- The second layer type is sand. The same trend is observed for the sand layer too, under prediction in the first 45 days and over prediction during snow melt.
- The third layer was found to be sandy clay loam. The soil layer follows the same trend as the upper layer in the first few days there is under prediction and then there is over prediction towards the end of the episode.
- The fourth layer, investigated as clay loam, the soil layer tends to over predict almost throughout the episode.
- The soil temperatures for the fifth layer were kept constant for through out the simulation period (Fixed boundary condition).

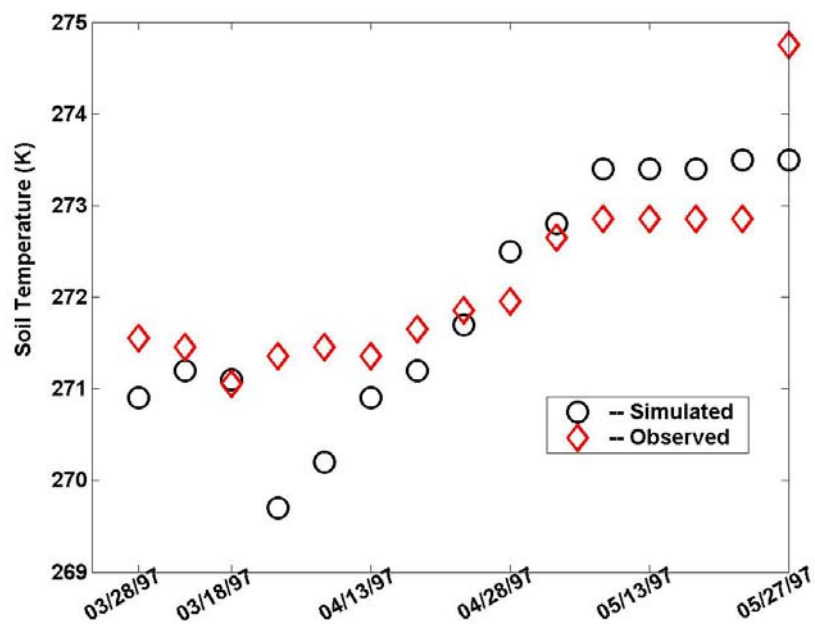


Figure 4.2.3

Graph of simulated vs. observed soil temperature at depth of 0.05 m for the region of Sodankylä

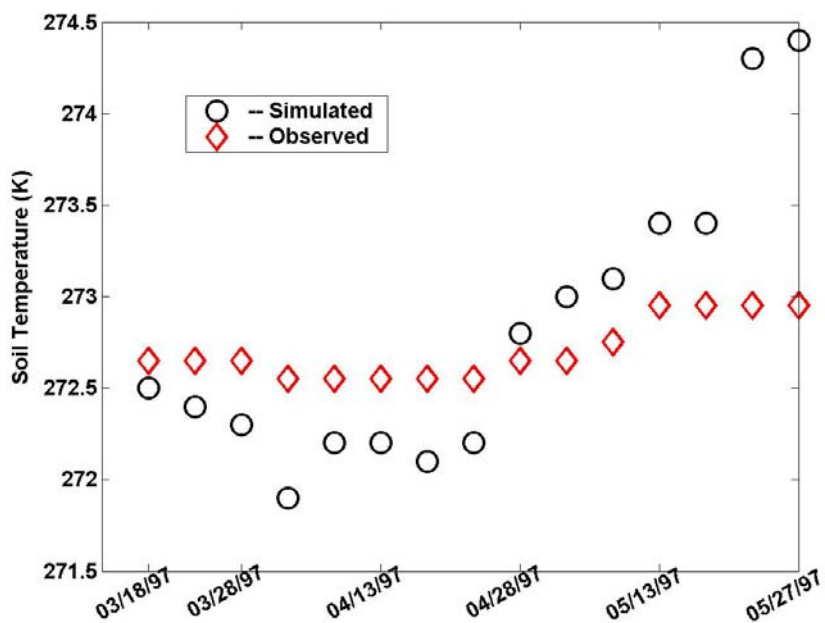


Figure 4.2.4

Graph of simulated vs. observed soil temperature at depth of 0.126 m for the region of Sodankylä

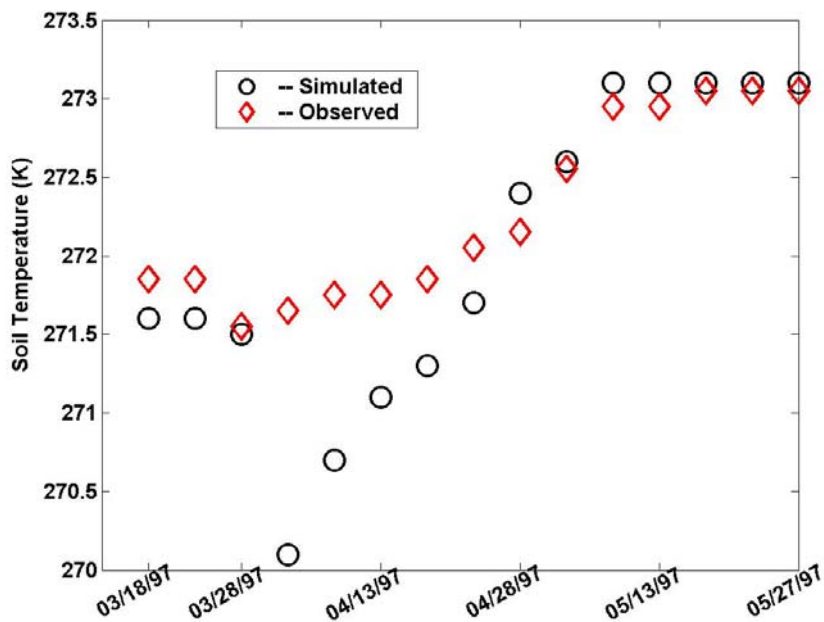


Figure 4.2.5

Graph of simulated vs. observed soil temperature at depth of 0.316 m for the region of Sodankylä

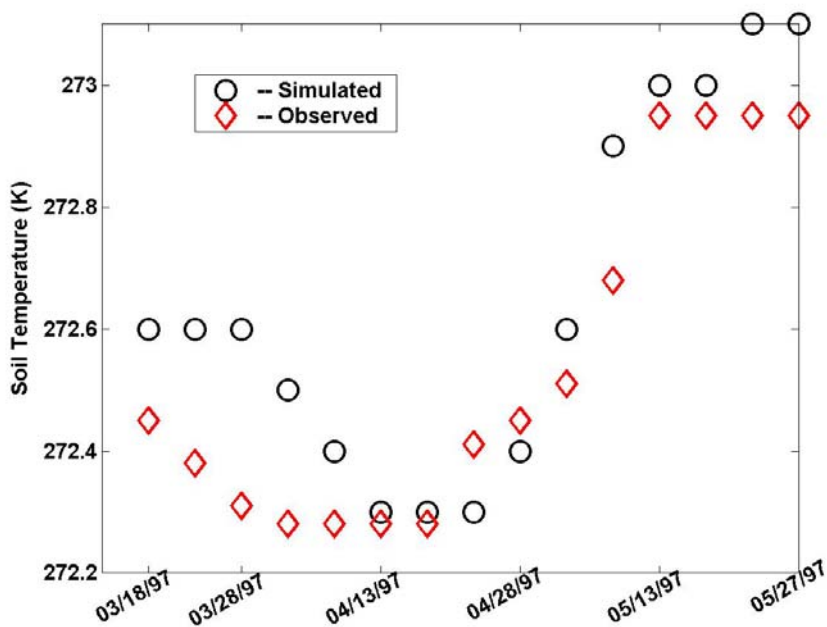


Figure 4.2.6

Graph of simulated vs. observed soil temperature at depth of 0.795 m for the region of Sodankylä

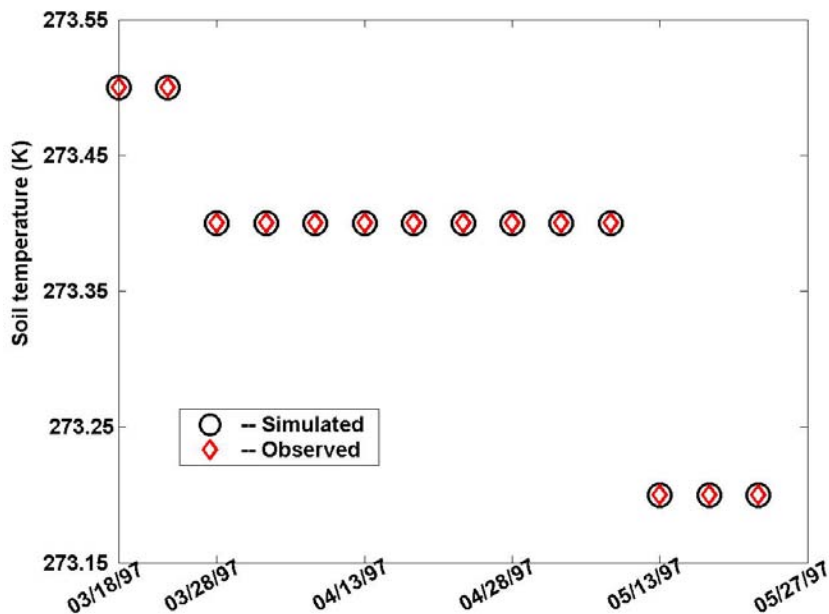


Figure 4.2.7

Graph of simulated vs. observed soil temperature at depth of 2 m for the region of Sodankylä

Table 4.2.1 RMS errors for different soil types

	1	2	3	4	5	6
Depth	Sand	Loam sand	sandy loam	Silt loam	Silt	Loam
0.05	0.86	4.26	0.76	13.25	13.25	0.732
0.126	0.62	3.83	1.22	10.43	10.43	1.32
0.316	0.6702	2.43	0.55	6.5	6.49	0.55
0.795	0.43	0.26	0.28	1.88	1.88	0.319

	7	8	9	10	11	12
Depth	Sand clay loam	Silty clay loam	Clay loam	Sandy clay	Silty clay	Clay
0.05	0.749	3.93	1.77	4.55	4.55	2.5
0.126	1.26	3.21	1.83	3.8	3.8	2.32
0.316	0.55	1.55	0.99	2.19	2.19	1.3
0.795	0.276	0.18	0.15	0.2	0.2	0.188

Table 4.2.2 Initial values for Sodankylä

Parameters	Values
	Assigned
Soil Type	Silt Loam
Vegetation Type	Grass
Vegetation Fraction	30%
Total Hours	594
Evaluation episode	March
Wind Height	2 m
Time since last snow fall event	0.00s
Initial height of snow at start of snowfall	0.9m
Ground temperature	257.72 K
Foliage temperature	257.72 K
Air temperature	257.72 K
Soil Temperature at 5 cm	273.55K
Soil Temperature at 12.6 cm	272.45K
Soil Temperature at 31.6 cm	271.85K
Soil Temperature at 79.5 cm	272.65K
Soil Temperature at 200 cm	271.55K
Total (liquid and solid) relative Soil volumetric water content at 5 cm	6.68%
Total (liquid and solid) relative Soil Volumetric water content at 12.6 cm	37.22%
Total (liquid and solid) relative Soil Volumetric water content at 31.6 cm	87.4%
Total (liquid and solid) relative Soil Volumetric water content at 89.5 cm	56.9%

Total (liquid and solid) relative Soil Volumetric water content at 200 cm	48.4%
Density of snow at layer 1	205.5 kg/m ³
Density of snow at layer 2	260.27 kg/m ³
Density of snow at layer 3	243.46 kg/m ³
Temperature of snow layer 1	260.1 K
Temperature of snow layer 2	265.4 K
Temperature of snow layer 3	268.6 K

4.3 Case 2: *Marsta*

The second case studies, analyzed the region of Marsta (Sweden) for just four days from 03/14/1997 to 03/18/1997. The model was initialized by using the initial values given in figure 4.3.1 for this particular case. This case was to test and confirm the insulating property of snow. As the two graphs showing the comparison of simulated and observed soil temperature for four day indicate just a straight flat line showing no diurnal effect for the depth of 30 cm and 10 cm. There is a good match between the observed and the simulated.

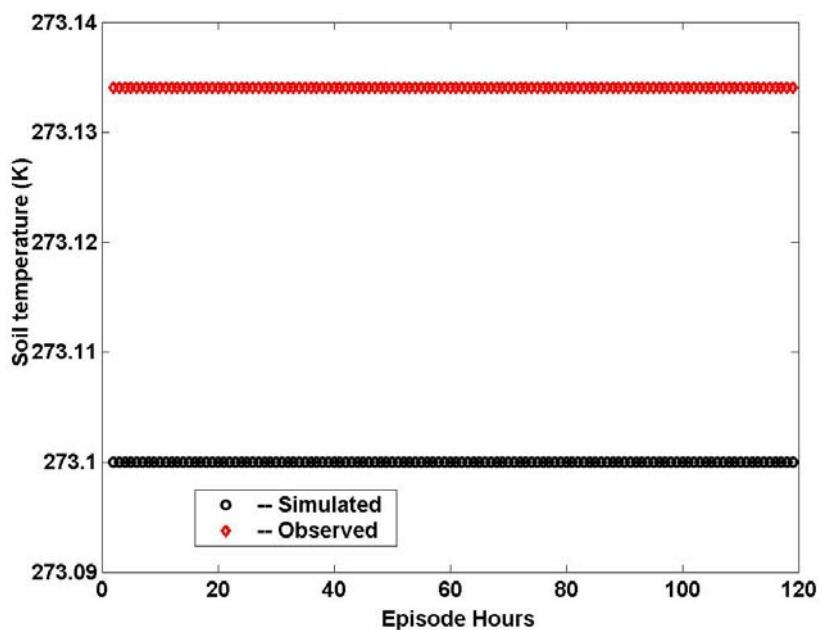


Figure 4.3.1

Graph of simulated vs. observed soil temperature at depth of 0.0126 m for the region of Marsta

Table 4.3.1 Initial values for Marsta

Parameters	Values Assigned
Soil Type	Sand
Vegetation Type	Grass
Vegetation Fraction	70%
Total Hours	119
Month	March
Wind Height	1 m
Time since last snow fall event	0.00seconds
Initial height of snow at start of snowfall	0.048 m
Ground temperature	271.81K

Foliage temperature	268.81 K
Air temperature	275.81 K
Soil Temperature at 5 cm	275.21 K
Soil Temperature at 12.6 cm	274.81 K
Soil Temperature at 31.6 cm	271.93 K
Soil Temperature at 79.5 cm	273.37 K
Soil Temperature at 200 cm	271.55 K
Soil Volumetric water content at 5 cm	86%
Soil Volumetric water content at 12.6 cm	85%
Soil Volumetric water content at 31.6 cm	80%
Soil Volumetric water content at 89.5 cm	87%
Soil Volumetric water content at 200 cm	85%
Density of snow at layer 1	400 kg/m ³
Density of snow at layer 2	460.27 kg/m ³
Density of snow at layer 3	443.46 kg/m ³
Temperature of snow layer 1	270 K
Temperature of snow layer 2	269 K
Temperature of snow layer 3	280 K

CHAPTER 5

CONCLUSION AND SCOPE FOR FUTURE WORK

- *5.1 Concluding Remarks*
- The soil temperature showed a slight decrease and increase with time showing the response to the rise in temperature in the spring.
- Greatest discrepancies in observed and simulated soil temperatures during snow-melt and spin-up.
- Different soil types were found to fit various depth of the soil, i.e. meteorological models should use a vertically variable soil type distribution
- The snow model tends to over-estimate snow height until the melting season sets on.
- The snow model has a slight offset for melt-up.
- The snow model is very sensitive to proper initialization of snow density and snow temperature.

- Snow simulations were sensitive to cloud cover, snow albedo and parameterization of long-wave downward radiation.
- HTSVS provides reasonable results so it can be considered as suitable for application in weather or climate models.

5.2 Scope for future work

- Further investigations to test HTSVS under various other conditions, add more parameters.
- More offline evaluations using other sites around the globe too test the performance of HTSVS
- Discrepancies related to snow model should be investigated
- Horizontal fluxes within the soil and snow should be studied
- Horizontal advection of wind should be incorporated
- Discrepancies related to albedo can be investigated

- Study of HTSVS with NWP models can be researched
- More soil layers should be added and the effect of it should be studied

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