A simple model for simulation of water content, soil frost, and soil temperatures in boreal mixed mires

G. Granberg, H. Grip, M. Ottosson Löfvenius, I. Sundh, B. H. Svensson, and M. Nilsson

Abstract. In this paper we present a model that can reconstruct water table position and soil temperature profiles to 3 m depth in boreal mixed mire systems using the readily available climate data on air temperature and precipitation as driving variables. The model simulates complete, multiple annual cycles including winter conditions and freeze-thaw processes. The major requisite for an accurate description of the soil heat flux in a mire is an accurate description of the water content of the profile because of the high porosity of the soil and the thermal properties of water. The soil moisture profile in this model is described as a function of water table position derived from empirical published data of soil moisture profile measurements at different water table positions. The parameters of the model were set and tested for a lawn community, one of the most dominant plant communities occurring in boreal mixed mires. The model is optimized for the period 1995–1997 at the mixed mire Degerö Stormyr and was validated for the period 1997–1998 at the same location and for 1991–1993 at Storåmyran, 60 km away. The mean deviation of simulated to measured soil temperature from 10 cm to 3 m depth in the validation data sets was <1.6°C with a maximal standard deviation of 1.8°C. A response analysis of the amount of winter precipitation showed the influence of snow cover depth on soil frost fluctuations. A simulated precipitation corresponding to 25% of the measured precipitation during winter 1995–1996, with all other factors unchanged, prolonged the period of frozen ground in the model from June 8 to August 1. With a doubling of the winter precipitation the modeled period of frozen ground lasted only until May 6.

1. Introduction

The hydrology, soil temperature, and depth of soil frost are the main abiotic factors influencing microbial activity and subsequent production of greenhouse gases in northern wetlands. There are few existing long-term records of soil temperature and water content from boreal mixed mires, and simulated data will make a valuable contribution to the evaluation of interannual variation in trace gas emission from these environments.

The pronounced climatic change at high latitudes predicted to occur as a result of global warming [Manabe and Wetherald, 1980, 1986; Ramanathan, 1988] is expected to alter the cycling characteristics of the large pools of carbon stored in boreal wetlands and thereby create a significant feedback effect on the global climate. One way that a changed climate might affect the system is via the strong relationships between water table position, soil temperatures, and emissions of methane in the wetlands [Dise et al., 1993; Granberg et al., 1997; Bubier et al., 1993; Moore and Roulet, 1993].

Numerical soil models describing heat fluxes, moisture regimes, and freezing in soils have been frequently applied to mineral soils for many years [Jansson and Halldin, 1979]. However, even though many hydrological models have been developed for mires [e.g., Guertin, 1987], the model presented by Froliking and Crill [1994] is to our knowledge the only one that describes heat fluxes, moisture regimes, and soil frost in a boreal mire. Their model is based on a linear function to describe the water content above the water table and uses daily air temperature, daily precipitation, and total daily net radiation data as driving variables.

The necessity of correctly describing water content over time in modeling heat flows and ice formation in arctic soils was discussed by Waeldbroeck [1993]. In peat-forming mires it is even more important to have a correct water content description for calculating heat fluxes because of the high water content in the saturated zone and the large variation of the water content in the unsaturated zone. Accurate water content determination of the unsaturated zone in mires is difficult with standard techniques, so accurate published data are rare [Ingram, 1983]. Hayward and Clymo [1982] used gamma ray attenuation to measure change in water content of intact peat profiles accompanying changed water table positions. Their measurements are probably still the most reliable soil moisture profiles pertaining to mire plant communities.

The fewer climatic variables needed to drive a model, the easier it will be to find long-term records with only a small amount of missing data. The variables most readily available from long time series and from widely spread locations are air temperature and precipitation. Thus the purpose of this study was to construct a model driven only by air temperature and...
precipitation, enabling the reconstruction of historical soil temperature, water content, and soil frost data from mixed mires. We also intended to model the complete annual cycle including winter conditions with freeze-thaw processes over multiple annual periods. To achieve these goals and to test the generality of the model, parameters were set with data from a 3-year period for one mire, and the model was validated for another time period at the same mire and for another mire situated 60 km away.

2. Material and Methods

2.1. Site Description

2.1.1. The parameterization site. Degerö Stormyr (64°11'N, 19°33'E, altitude 270 meters above sea level (masl) area 6.5 km²) is a mixed acid mire system situated at the Vindeln Experimental Forests, in the research park Kulbückslden, ~70 km from the Gulf of Bothnia in the County of Västerbotten, Sweden (Figure 1). The mire is positioned on highland between two major rivers, the Umeälven and the Vindelälven. The mire area consists of a quite complex system of small, interconnected mires divided by islets and ridges of glacial till. The total catchment area including Degerö Stormyr, the total annual precipitation is 523 mm; the mean annual temperature is +1.2°C; the July temperature is 14.7°C; and the January temperature is −12.4°C. At the Urmeälven climate station, ~20 km north of the Storåmyr test site, the total annual precipitation is 662 mm; the mean annual temperature is +2.7°C; the July temperature is 15.5°C; and the January temperature is −8.7°C (normal period 1961–1990).

2.1.2. The validation site. Storåmyran (63°44'N, 20°06'E, altitude 35 masl, area 1 km²) is a Calluna Sphagnum fuscum bog [Eurola et al., 1984]. It is a mixed acid mire, situated 7 km from the coast of the Gulf of Bothnia, on a flat coastal flood plain with small glacial till formations. In the center of the mire there is a 12 ha ombrogenic feature [Waddington and Roulet, 1997]. All measurements referred to in this study were taken in this ombrogenic area. The two temperature profile measurements used in the study are from a carpet community dominated by E. vaginatum, S. palustris, and S. majus, located on the southern and northern edges of a small ridge, extending in a north–south direction. The northern and southern sites will be referred to as NS and SS, respectively. Water table data from an adjacent lawn community dominated by E. vaginatum and S. balticum were also used in this study. The depth of the peat is 3 m.

2.2. Climate

The climate of both sites is characterized as cold temperate humid. At the Kulbückslden climate station, close to Degerö Stormyr, the total annual precipitation is 523 mm; the mean annual temperature is +1.2°C; the July temperature is 14.7°C; and the January temperature is −12.4°C. At the Umeälven climate station, ~20 km north of the Storåmyr test site, the total annual precipitation is 662 mm; the mean annual temperature is +2.7°C; the July temperature is 15.5°C; and the January temperature is −8.7°C (normal period 1961–1990).

2.3. Field Measurements

The meteorological data used as input data were monitored at climate stations located on the actual mires, and complementary data were obtained from the climate station at Kulbückslden (2 km away) for Degerö and from the meteorological station at Umeälven airport (20 km away) for Storåmyr. At Degerö Stormyr, data were collected during the snow-free period in 1995 and continuously from April 11, 1996. Data from Storåmyr were only collected during the snow-free period.

Air temperature and relative humidity were measured at standard screen height, 1.5–2 m above the ground, using an unventilated radiation shield equipped with a Rotronic Air-Probe YA-100-Hygrometer (Rotronic ag, Zürich, Switzerland). The wind speed was measured with a cup anemometer at 5 m, and the precipitation was recorded with a tipping bucket (ARG-100, Campbell SCI) set 1 m above the vegetation surface (except for the period June 8–15, 1996, when the tipping bucket was placed on the vegetation surface).

The soil temperature was continuously monitored at Degerö Stormyr at 2, 10, 18, 26, 34, and 42 cm depth using thermistors. In April 1996, additional thermistors were installed at 0.5, 1, 2, and 3 m depth. Furthermore, the water table and peat surface levels were continuously monitored with a potentiometric water level recorder [Roulet et al., 1991]. Soil temperature data from Storåmyr were collected with probes at 5, 10, 20, and 45 cm depths. Data from two profiles are used for 1991 and 1992. From 1993, only data from NS were available. Water table positions from Storåmyr were manually measured, at 2 week intervals during summer 1991 and 1992, from stationary perforated plastic (PVC) tubes placed close to the temperature profile measurement points. All sensors were queried at 10 min intervals and stored as hourly averages on a Campbell Scientific CR-10 logger. During the winter period the query interval was prolonged to 40 min, and data were stored on a 2 hourly
basis. Snow depth was measured at the Kulbücksiden climate station as part of a reference monitoring program and also at Degerö Stormyr during winter 1997-1998. Snow depth data were also occasionally available from Storfimyran.

3. Model Description

3.1. General Approach

The Mixed Mire Water and Heat model (MMWH) is based on fundamental geophysical relationships and is designed to be as general as possible for mixed mires in fairly flat locations. One important objective was to reduce driving variables to the readily available climatic data on air temperature and precipitation, allowing reconstruction of historical data sets of soil temperature and water content profiles. The MMWH model in this study was tested, and the parameters were set in a lawn community, one of the most dominant plant communities occurring in boreal mixed mires.

MMWH was developed and is run in the MATLAB module SIMULINK (Math-works Inc.). It consists of three submodels: (1) a hydrological bucket model describing the change in water content of a unit area, (2) a box flow model describing the heat flux in the profile, and (3) a snow model describing snow cover fluctuations. The description of the water content profile is based on soft gamma radiation measurements of intact peat cores conducted by Hayward and Clymo [1982]. The snow model is based on the snow model used in the SOIL model [Jansson and Halldin, 1979; Jansson, 1996].

3.2. Porosity

The variation in organic fraction (by volume) of the profile in MMWH is described using mean values representing two layers, 2% in the upper 30 cm and 8% in the underlying peat profile. The value for the upper layer is based on mean bulk density measurements gathered from profiles collected at Degerö Stormyr. The measured bulk density was mostly around 0.02 g cm\(^{-3}\) in the upper 20 cm and increased to a mean value of 0.05 g cm\(^{-3}\) in the 20-40 cm section. On visual examination of the profiles a marked boundary around 30 cm deep was detected, which was used as a separator between deep peat, with relatively high bulk density, and the less dense upper zone. The soil water characteristics in the upper peat are considered to be linear in the suction interval 0-0.1 m and constant in the interval 0.1-0.3 m. The gradient \(a_z\) in the linearly decreasing interval is given by (2) (see text). The solid area to the right represents the proportion of the profile occupied by organic material (as dry matter). The other portions of the profile show the water content at the different water table positions.

\[
\theta_{sw}(z) = \min \left[ \phi, \theta_s + (\phi - \theta_s \left( \frac{z}{z_{sw}} \right)^2 \right]
\] (1)

The soil water characteristics in the upper peat are considered to be linear in the suction interval 0-0.1 m and constant in the interval 0.1-0.3 m. The gradient \(a_z\) in the linearly decreasing interval is given by

\[
a_z = \frac{\phi - \theta_{sw}}{z_{sw}} \]

where \(z_{sw} = 0.1\) and the minimum value \(\theta_{sw}\) is set to 0.25, which approximately corresponds to the amount of water still held by the capillarity of the Sphagnum mosses at the lowest range of the water tables considered in this study [Clymo and Hayward, 1982]. The functions are chosen to resemble water content profiles in mires adapted from Hayward and Clymo [1982].

The total volume of water in the profile from the vegetation surface to depth \(z_a\) would be

\[
\theta = \max \left[ \theta_{sw}, \phi - (a_z z_a) \right]
\]

These properties make it possible to consider the system as a simple bucket model, where the water content in the profile is dependent only on the position of the actual water table. To describe accurately the water table position, the water-holding capacity of the system is mimicked by a constructed function.
Figure 3. Functions for relative transmissivity and reduction of potential evapotranspiration. The transmissivity function is a direct exponential curve fitted to empirically measured (solid diamonds) transmissivity for a Sphagnum-Eriophorum-sedge microtope from Ivanov [1981]. The actual evapotranspiration (bold line) decreases according to a square root function approaching zero when the water table reaches 0.28 m.

3.4. Change in Water Storage

The contribution of terrestrial inflow to the water balance of a mixed mire system with a large mire/total watershed area ratio, situated in flat terrain, is small compared to the contribution from the precipitation on the mire surface. The modeled inflows can therefore be restricted to the actual precipitation \( P \) and snowmelt from the snow pack \( S_m \) on the mire. The outflows are evapotranspiration \( E \), and runoff \( Q_{out} \) due to the slope of the water table. The water balance gives the change in storage \((\text{m}^3\text{d}^{-1})\):

\[
\Delta S = P + S_m - E - Q_{out}.
\]  

3.5. Precipitation

Precipitation is separated into rain and snow using air temperature \( T_a \) as a separator. The precipitation was considered to be all snow if \( T_a \leq -2^\circ\text{C} \) and all rain if \( T_a \geq 2^\circ\text{C} \) and to follow a linear relationship in-between these values. The measured precipitation was initially corrected for gauge inefficiency by adding 10% if rain and 18% if snow, according to Eriksson [1979]. An extra 10% percent rain precipitation correction was added to the input data for the period following June 15, 1996, to compensate for the movement of the rain gauge from the surface to the more wind-exposed position at 1 m elevation.

3.6. Snowmelt Runoff

When a snow layer was present, all precipitation entered that pool. If the amount of free water in the snow pack exceeded the retention capacity \( c_{ret} \) (which is assumed to be 7% of the total water equivalent of the snow pack according to Jansson [1996]) and if the water table was lower than the vegetation surface, the snow melt runoff entered the water balance submodel.

3.7. Evapotranspiration

Potential evapotranspiration \( E_p \) is calculated using a modified version of the Blaney-Criddle temperature equation as proposed by Doorenbos and Pruitt [1977], which accounts for variations of the local climate \((\text{m}^3\text{d}^{-1})\):

\[
E_p = \frac{a_e p b_e p + c_e p (0.46 T_a + 8)}{d_w} \quad (\text{m}^3\text{d}^{-1})
\]

Here \( d_w \) is daily percentage of total annual daylight hours, \( T_a \) is mean diurnal air temperature \((^\circ\text{C})\), \( a_{ep} \) is an adjustment factor used for latitudes of more than 55°, where radiation, and therefore \( E_p \), is weaker than at low-latitude and midlatitude areas having the same day length, and \( b_{ep} \) and \( c_{ep} \) are parameters determined from local data of minimum relative humidity, ratio of actual to maximal sunshine hours, and daytime wind estimates [Doorenbos and Pruitt, 1977]. The actual evapotranspiration was described by adjusting the potential evapotranspiration to the availability of water determined by the position of the water table.

The ability of the Sphagnum mosses to wick water decreases rapidly with a receding water table because of the increased path length for capillary rise and decreased hydraulic conductivity as water is withdrawn from the larger pores [Ingram, 1983]. The transpiration from vascular plants is only affected by a drop in the water table during extremely dry periods when it drops below their root system. The actual evapotranspiration \( E_a \) is assumed to decrease by a square root function, approaching zero when the water table reaches the transition zone between the acrotelm and the catotelm (Figure 3). The actual evapotranspiration is given by \((\text{m}^3\text{d}^{-1})\):

\[
E_a = \begin{cases} 
E_p & z \leq a_{ea} \\
E_p \sqrt{(b_{ea} - z_{wa})/(b_{ea} - a_{ea})} & z > a_{ea}
\end{cases}
\]  

where \( a_{ea} \) is the maximum depth to where \( E_a = E_p \) (m) and \( b_{ea} \) is the depth where \( E_a = 0 \) (m). All evapotranspiration was prohibited in the presence of snow cover.

3.8. Outflow

The net outflow from a unit area \( A \) is considered horizontal and therefore related to the local slope of the water table \((dh/dl)\) and the transmissivity \( T_r \) of the active zone \((\text{m}^3\text{d}^{-1})\):

\[
Q_{out} = -T_r w \frac{dh}{dl} / A,
\]  

where \( w \) is the width of the layer (m). The transmissivity is a function of the position of the water table. In the upper layer,
conductivity decreases with depth, resulting in an exponential decrease in transmissivity when the water table drops below the soil surface, and will asymptotically approach zero when it reaches the catotelm. The transmissivity function is a direct exponential curve fitted to empirical data of transmissivity for a Sphagnum-Eriophorum-sedge microtope [Ivanov, 1981] (Figure 3) (m² d⁻¹):

\[ T_r = a e^{b \omega_r} \]

where \( a \) and \( b \) are constants in the fitted equation. When the mass fraction of frozen water exceeded 0.7 in the top 0.08 m of the profile (\( f_{\text{freez}} \)), no runoff occurred.

### 3.9. Soil Temperature

Heat transfer in peat soils occurs predominantly by conduction because of the high water content [Hillel, 1982] and is described by the Fourier equation:

\[ C_h \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2} \]

where \( C_h \) is the volumetric heat capacity of the peat (J m⁻³ °C⁻¹), \( K \) is the soil thermal conductivity (W m⁻² °C⁻¹), \( z \) is the vertical distance (m), and \( t \) is time (s). The soil temperatures are modeled with a box flow model working in one dimension (vertical), calculating the heat balance in each compartment from inputs, outputs, and change in temperature using a forward difference of explicit method. The soil profile was divided into 15 compartments, thin near the surface for high resolution and thicker with depth, to make long-term simulations possible [Campbell, 1985]. The profile was divided into elements 2, 4, 4, 4, 4, 4, 8, 8, 8, 25, 25, 50, 50, 100, and 100 cm thick, downward from the vegetation surface.

### 3.10. Thermal Properties

Vertically orientated Sphagnum stems form most of the structure of the unsaturated zone of mixed mires. The soil water in this zone, which is the major component responsible for the heat transport, is closely associated with the Sphagnum mosses, and consequently, the soil volume occupied by air is dispersed between the vertically orientated, saturated Sphagnum mosses. When the ratio between the solid and fluid thermal conductivity is low, as in this system, parallel heat flow is expected to predominate [Farouki, 1986], and a weighted arithmetic mean equation can be used to describe thermal conductivity. To integrate the effect of formation of ice on the thermal conductivity in this system, we use a model where the water, ice, and organic volumetric content is treated as a separate subvolume parallel to the soil air volume. This subvolume is considered to be fully saturated, and its thermal conductivity is calculated using the geometric mean equation [Farouki, 1986] and then weighted by its volumetric fraction. The soil water fraction bound below the wilting point (\( f_{\text{wp}} \)) is assumed to remain unfrozen and is therefore treated as a separate constituent in the equations describing heat capacity. The peat thermal conductivity (\( \lambda_{\text{total}} \)) will be

\[ \lambda_{\text{total}} = (\phi - \theta) \lambda_{\text{air}} + (\theta + f_{\text{org}}) f_{\text{org}} \lambda_{\text{org}}(\phi - f_{\text{org}})/(\theta + f_{\text{org}}) \lambda_{\text{water}}(\theta + f_{\text{org}}), \]

where \( \phi \) is porosity, \( \theta \) is total water content, \( f_{\text{org}} \) is volumetric fraction of organic matter, and \( f_{\text{ice}} \) is volumetric fraction of ice. The effect of freeze-thaw processes on the heat capacity is considered by adding a component including the latent heat of fusion (\( L_f \)). The overall heat capacity \( C_{\text{total}} \) was derived as the sum of the heat capacities of the constituents weighted by their volumetric fractions:

\[ C_{\text{total}} = (\phi - \theta) C_{\text{air}} + (\theta + f_{\text{org}}) C_{\text{org}} + (\theta - f_{\text{ice}}) C_{\text{water}} \]

\[ + f_{\text{ice}} C_{\text{ice}} + f(\theta - f_{\text{wp}}) \frac{L_f}{(\Delta T)^2}, \]

with capacity values obtained from Hillel [1982]. The soil water is assumed to freeze continuously and uniformly over a small finite temperature range \( \Delta_T \), here set to 1°C, around 0°C, and \( f(\theta - f_{\text{wp}}) \) is a step function that is 1 during freezing or thawing and 0 otherwise. To allow continuous movement of the soil frost border in the profile, the temperature in the layers that are freezing or thawing is halted at 0°C. A change in water content in a compartment with a proportion of the water occurring as ice will affect the proportion of ice to total water content, and the fraction of ice will be adjusted to fit the new total water content.

### 3.11. Snow Cover

The fluctuation of the snow cover over the winter season was modeled according to the snow model in the SOIL model [Jansson, 1996]. The entire snow cover is considered as total water equivalent \( S \) with a separated fraction of unfrozen water \( S_{\text{uf}} \). A melting-freezing function combines the two different pools. The daily amount of snow melt (\( Q_{\text{melt}} \)) is derived from an air temperature function (\( Q_{\text{a}} \)) and the soil surface heat flow (\( q_{\text{ho}} \)):

\[ Q_{\text{melt}} = T_r Q_{\text{a}} + \frac{q_{\text{ho}}}{L_f} \]

The function linking air temperature and snow melt is empirical and based on the day of the year as input, following the work of E. A. Andersson, as discussed by Bras [1990]:

\[ Q_{\text{a}} = 0.5 T_s (\text{MF}_{\text{max}} + \text{MF}_{\text{min}}) \]

\[ + (\text{MF}_{\text{max}} - \text{MF}_{\text{min}}) \sin \left[ 2\pi (d - 79)/366 \right], \]

where \( Q_{\text{a}} \) is snow melt per hour, \( d \) is day number (January 1 = 1), and \( \text{MF}_{\text{max}} \) and \( \text{MF}_{\text{min}} \) are snow melt parameters. An adjustment suggested by E. A. Andersson for Fairbanks, Alaska, which is situated at a similar latitude to the mires in this study, is used to decrease the temperature snowmelt function in midwinter:

\[ Q_{\text{a}} = (Q_{\text{a}} - \text{MF}_{\text{min}}) F + \text{MF}_{\text{min}} \]

where \( F \) is an adjustment given by

\[ F = \begin{cases} 0 & X \leq 0.48 \\ \frac{(X - 0.48)/0.22}{0.48 < X < 0.7} & X \geq 0.48 < X < 0.7 \\ 1 & X \leq 0.48 \end{cases} \]

and \( X \) is the proportion of the time between December 21 and June 21. During periods with unfrozen soils, snowmelt will also occur through heat flux from the upper soil layer. Refreezing is modeled by the same function, but while melting affects the whole snowpack, refreezing affects only a fraction of it near the surface defined by

\[ Q_{\text{refreeze}} = Q_{\text{melt}} \min \left( 1, \frac{RF}{\Delta T_{\text{snow}}} \right), \]
Table 1. Parameters of the Mixed Mire Water and Heat Model (MMWH)

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Value</th>
<th>Optimized*</th>
<th>Units</th>
<th>Description</th>
<th>Reference</th>
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<td>...</td>
<td>...</td>
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<td>Doorenbos and Pruitt [1977]</td>
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<td>...</td>
<td>...</td>
<td>potential evapotranspiration parameter</td>
<td>Doorenbos and Pruitt [1977]</td>
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<td>x m</td>
<td>...</td>
<td>upper limit for maximum evaporation</td>
<td>Hayward and Clymo [1982]</td>
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<tr>
<td>b ea</td>
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<td>...</td>
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<td>x</td>
<td>...</td>
<td>fraction of ice in top 8 cm to block runoff</td>
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**Hydraulic**

<table>
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<tr>
<th>Symbol</th>
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<th>Description</th>
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<td>m</td>
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<td></td>
</tr>
<tr>
<td>( C_{c} )</td>
<td>2.5 \times 10^{6}</td>
<td>J m (^{-3}) °C (^{-1})</td>
<td>heat capacity organic material</td>
<td>Hillel [1982]</td>
<td></td>
</tr>
<tr>
<td>( C_{c} )</td>
<td>4.2 \times 10^{5}</td>
<td>J m (^{-3}) °C (^{-1})</td>
<td>heat capacity water</td>
<td>Hillel [1982]</td>
<td></td>
</tr>
<tr>
<td>( C_{c} )</td>
<td>1.25 \times 10^{4}</td>
<td>J m (^{-3}) °C (^{-1})</td>
<td>heat capacity water</td>
<td>Hillel [1982]</td>
<td></td>
</tr>
<tr>
<td>( \Delta T )</td>
<td>1</td>
<td>°C</td>
<td>temperature interval for freezing/thawing</td>
<td>set*</td>
<td></td>
</tr>
<tr>
<td>( L_{f} )</td>
<td>3.3 \times 10^{6}</td>
<td>J m (^{-3})</td>
<td>latent heat fusion</td>
<td>Jury [1991]</td>
<td></td>
</tr>
<tr>
<td>( f_{wp} )</td>
<td>0.05</td>
<td>...</td>
<td>wilting point fraction</td>
<td>Pålsvåken [1973]</td>
<td></td>
</tr>
</tbody>
</table>

**Snow**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Value</th>
<th>Optimized*</th>
<th>Units</th>
<th>Description</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>MF(_{max})</td>
<td>0.004</td>
<td>m d (^{-1})</td>
<td>snow melt maximum</td>
<td>Bråm [1990]</td>
<td></td>
</tr>
<tr>
<td>MF(_{min})</td>
<td>0.002</td>
<td>m d (^{-1})</td>
<td>snow melt minimum</td>
<td>Bråm [1990]</td>
<td></td>
</tr>
<tr>
<td>RF</td>
<td>0.1</td>
<td>m</td>
<td>snow melt refreezing efficiency</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( S_{ret} )</td>
<td>0.07</td>
<td>...</td>
<td>retention capacity of the snowpack</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( S_{w} )</td>
<td>0.003</td>
<td>m</td>
<td>minimum amount of water in snow to set ( T_0 = 0 )</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( \rho_{s} )</td>
<td>100</td>
<td>kg m (^{-3})</td>
<td>minimum snow density</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( \Delta S_{c} )</td>
<td>0.07</td>
<td>x m</td>
<td>depth of snowpack complete snow cover</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( S_{m} )</td>
<td>200</td>
<td>m</td>
<td>snow density parameter, free water</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( S_{w} )</td>
<td>0.5</td>
<td>m</td>
<td>snow density parameter, total water</td>
<td>Jansson [1996] (SOIL)</td>
<td></td>
</tr>
<tr>
<td>( k_{snow} )</td>
<td>( 2.86 \times 10^{-6} )</td>
<td>W m (^{-3}) kg (^{-1})</td>
<td>thermal conductivity parameter for snow</td>
<td>U.S. Army Corps of Engineers [1956]</td>
<td></td>
</tr>
</tbody>
</table>

*The parameters marked with x in this column are optimized for Degerö Stormyr.

*According to section 3.1.2.

*According to section 3.10.

where \( \Delta z_{snow} \) is snow depth (meters) and RF is a critical value at which the refreezing becomes inversely dependent on the snow depth (meters) [Jansson, 1996]. A completely frozen precipitation is assumed to have a constant minimum density \( \rho_{s_{min}} \). Mixed precipitation density \( \rho_{prec} \) depends on the ratio between rain and total precipitation according to

\[
\rho_{prec} = \rho_{s_{min}} + \left( \rho_{water} - \rho_{s_{min}} \right) \frac{P_{r}}{P},
\]

where \( P_{r} \) is rain and \( P \) is total precipitation. The snow density used to calculate heat flow through the snowpack is a weighted average of the density of the old snowpack (i.e., the density of the snow remaining from the previous time step) and precipitation density. The density of the old snowpack increases with the relative amount of free water in the pack and overburden water, i.e., with increasing water equivalent. The age dependency of the old snowpack is accounted for by updating snow density as the maximum density of the previous time step or

\[
\rho_{old} = \rho_{s_{min}} + s_{a} \frac{S_{a}}{C_{ret} S_{res}} = s_{dw} S_{res}
\]

where \( s_{a} \) and \( s_{dw} \) are parameters, listed in Table 1, and \( S_{res} \) is the water equivalent of the snowpack from the previous time step.

3.12. Boundary and Initial Conditions

The upper boundary condition is commonly determined using the assumption that the daily mean temperature of the first node is the same as the daily mean air temperature. During the summer period, there is a relatively large difference between the duration of the near-ground daytime lapse temperature profile and the nighttime inversion because of the high latitude of the mires investigated in this study. Thus, if the daily mean air temperature at standard screen height is substituted for the soil surface temperature, there will be an underestimation of the soil surface temperature. In a simulation where the air
temperature was used directly as the first node temperature this underestimation of the soil temperatures was correlated with day length and was maximal at the summer solstice (Figure 4). A correction term was therefore used when relating the soil surface temperature \( T_s \) to the air temperature:

\[
T_s = T_a + f(d_t) \left( \frac{d_t - 12}{d_{t,\text{max}} - 12} \right) a_{\text{sT}},
\]

where \( d_t \) is day length in hours, \( d_{t,\text{max}} \) is day length at solstice, \( a_{\text{sT}} \) is a surface temperature correction parameter, and \( f(d_t) \) is a step function that is 1 when \( d_t \) is larger than 12 and 0 otherwise. During periods of snow cover the soil surface temperature is derived by assuming a steady state heat flow between the soil and a homogenous snowpack. The conductivity of the snow cover is determined from the density of the snow [U.S. Army Corps of Engineers, 1956]:

\[
\lambda_{\text{snow}} = k_{\text{water}} \rho_{\text{snow}}^2.
\]

where \( \rho_{\text{snow}} \) is snow density (kg m\(^{-3}\)). The surface temperature \( T_{ss} \) is then calculated as a weighted average derived from the temperatures and conductivities of the snow and soil:

\[
T_{ss} = T_1 + a T_s \left( \frac{\Delta z_0}{2} \right) \frac{1}{\lambda_{\text{soil}} \lambda_{\text{snow}}},
\]

where \( T_1 \) is the temperature at the first node, \( \Delta z_0 \) is the thickness of the first element, and \( \lambda_{\text{soil}} \) is the conductivity of the first element. If the free water content in the snow exceeds a constant threshold \( S_{\text{swt, min}} \), the \( T_{ss} \) is set to 0°C. To account for the effects of incomplete snow cover and radiation reaching the ground during periods with thin snow cover, a complete snow cover is assumed to occur only when the depth of the snowpack is above a certain value (\( \Delta z_{\text{cov}} \)). During conditions with snowpack depths below \( \Delta z_{\text{cov}} \), the soil surface temperature \( T_s(\Delta z_{\text{cov}}) \) is calculated as a weighted sum of the calculated soil surface temperature \( T_{ss} \) (from (23)) and the air temperature \( T_a \), which is assumed to represent the bare soil surface temperature:

\[
T_s(\Delta z_{\text{cov}}) = \left( 1 - \frac{\Delta z_{\text{snow}}}{\Delta z_{\text{cov}}} \right) T_s + \frac{\Delta z_{\text{snow}}}{\Delta z_{\text{cov}}} T_a.
\]

When water is pooled above the vegetation cover, \( T_{ss} \) is calculated by the same procedure as is used for complete snow cover. The lower boundary condition is set to 4 m, corresponding to the average peat thickness, and a constant temperature of 4°C, corresponding to the mean annual temperature measured at 3 m depth at Degerö Stormyr. The measured annual amplitude at 3 m is 1°C, and for the purpose of this study the approximation that the temperature at 4 m is constant is accurate enough.

The profile at Degerö Stormyr was initiated on June 8, 1995, setting the simulated water table and temperature according to water table and soil temperatures measured (for the upper 42 cm) on June 8, 1995, and (for the deeper zone) temperatures measured during the corresponding date in 1996. At the Storåmyran test site, values measured from June 20, 1991, were used for the upper zone, and since no deep profile measurements were taken, a profile from a corresponding date from Degerö Stormyr was used.

### 3.1.3. Numerical Techniques

The iteration technique chosen for solving the differential equations was the variable time step model Runge-Kutta 5. A stable model performance was achieved with a minimum time step size of 1.44 min and when the tolerance parameter, which controls the relative error of the integration at each step of the integration, was set to \( 1 \times 10^{-8} \).

### 4. Results

#### 4.1. Calibration and Validation

The MMWH model was constructed using data from Degerö Stormyr in 1995–1997 as calibration material. The generality of the model was tested by using the model to predict the soil temperature and water table positions at Degerö Stormyr in 1998 and at Storåmyran in 1991–1993. The calibration was first made for the hydrological submodel by trial and error optimization of evapotranspiration parameters \( a_{\text{ea}} \) and \( b_{\text{eat}} \), the run-off parameter \( f_{\text{runoff}} \) and the snowcover parameter \( \Delta z_{\text{cov}} \) until the simulated water table agreed with the measured water table. Then the surface temperature correction parameter \( a_{\text{sT}} \) was optimized until the modeled upper soil temperatures conformed to the measured temperatures.

#### 4.2. Model Performance

##### 4.2.1. Hydrology

The simulations of the water table using the calibration data from Degerö Stormyr were in good
agreement with the measured positions, with a maximal mean seasonal deviation of 2.1 ± 1.2 cm occurring during summer 1996 (Table 2). During the summer 1998 validation period at Degerö Stormyr the mean seasonal deviation was 1.3 ± 1.0 cm. The model successfully simulated the initiation of the spring runoff at Degerö Stormyr (Figure 5). The model also reproduced the general seasonal pattern of fluctuation of the water table in the validation set from Storåmyran. The simulation of the water table was generally lower than the observed values at the carpet plant community (Figure 6). The largest difference appeared at the beginning of the season in 1991, when the deviation was as high as 10 cm. The deviation was less great for the lawn plant community. The deviation between simulated and measured water table position was evenly distributed over time for the summer seasons 1995-1998 (Figure 7).

4.2.2. Soil temperatures. The model reproduced the seasonal and interseasonal soil temperature variations very satisfactorily. The mean deviation of the soil temperature in the calibration set 1995-1997 at Degerö Stormyr was never more than 1.3°C, with a maximal standard deviation of 1.1°C, which occurred at the surface layer during summer 1995 (Table 2). Farther down in the profile, the deviations between simulated and measured temperatures were smaller. For instance, at 26 cm and lower in the peat profile the maximum mean deviation was 0.5°C with a standard deviation of 0.4°C. At the deepest levels investigated the simulated temperatures approached the measured values with time (Figure 5). The mean deviation of the soil temperature during the summer 1998 validation period at Degerö Stormyr was slightly higher, with a maximum deviation of 1.3°C and a standard deviation of 0.5°C at 26 cm depth. The deviation between simulated and measured soil temperatures at the 10 cm level was evenly distributed over time for the summer seasons 1995-1998 (Figure 8).

In the validation set from Storåmyran the largest deviations were obtained at the 5 cm level, where the mean deviation was 1.6°C with a standard deviation of 1.8°C (Table 3). As for the validation data, the deviation between simulated and measured soil temperatures at the 10 cm level was evenly distributed over time (Figure 8). For levels deeper in the profile the simulated temperature during 1991 and 1992 was higher than the measured temperature at NS and slightly lower than the measured temperature at SS (Figure 6). After an overestimation in the beginning of the 1993 summer season the simulated temperature followed the NS temperature profile well.

4.2.3. Frost penetration and snow depth. Frost penetration in the profile was identified from the temperature profile

Table 2. Fit of Simulated Versus Measured Water Table (wt) and Soil Temperature Data From Degerö Stormyr 1995-1998

<table>
<thead>
<tr>
<th>Period</th>
<th>Deviation Simulated-Measured Variable</th>
<th>wt, cm Mean s.d.</th>
<th>T 02 Mean s.d.</th>
<th>T 10 Mean s.d.</th>
<th>T 26 Mean s.d.</th>
<th>T 42 Mean s.d.</th>
<th>T 200 Mean s.d.</th>
</tr>
</thead>
<tbody>
<tr>
<td>June 8 to Nov. 95</td>
<td>0.8 ± 1.7</td>
<td>1.3 ± 1.1</td>
<td>0.7 ± 0.3</td>
<td>0.3 ± 0.3</td>
<td>0.5 ± 0.2</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>April 14 to 3, 96</td>
<td>2.1 ± 1.2</td>
<td>0.8 ± 0.7</td>
<td>0.5 ± 0.5</td>
<td>0.4 ± 0.4</td>
<td>0.5 ± 0.4</td>
<td>0.1 ± 0.1</td>
<td>---</td>
</tr>
<tr>
<td>May 22 to Sept. 97</td>
<td>1.2 ± 0.7</td>
<td>0.9 ± 0.7</td>
<td>0.6 ± 0.3</td>
<td>0.5 ± 0.4</td>
<td>0.3 ± 0.2</td>
<td>0.2 ± 0.1</td>
<td>---</td>
</tr>
<tr>
<td>May 16 to Aug. 25</td>
<td>1.3 ± 1.0</td>
<td>1.0 ± 0.7</td>
<td>1.2 ± 0.6</td>
<td>1.3 ± 0.5</td>
<td>1.2 ± 0.7</td>
<td>0.2 ± 0.1</td>
<td>---</td>
</tr>
<tr>
<td>Nov. 7, 96, to May 21, 97</td>
<td>0.2 ± 0.7</td>
<td>0.3 ± 0.3</td>
<td>0.2 ± 0.5</td>
<td>0.1 ± 0.2</td>
<td>0.2 ± 0.1</td>
<td>---</td>
<td>---</td>
</tr>
</tbody>
</table>

*Measured and simulated soil temperatures from specific depths (centimeters below vegetation surface).

b T 200 not installed.

Table 2. Fit of Simulated Versus Measured Water Table (wt) and Soil Temperature Data From Degerö Stormyr 1995-1998

Figure 5. Simulated and measured snow depth, soil frost, water table, and soil temperatures from Degerö Stormyr, 1995-1998. (a) Measured snow depth from Kulbäcksliden climate station (open circles) and from Degerö Stormyr (solid circles) and measured soil frost boundaries (solid circles) obtained from soil temperature profiles from Degerö Stormyr. Simulated snow depth and frost borders are represented by shaded lines. (b) Measured water table (solid line) and simulated water table (shaded line). (c)-(f) Measured soil temperatures (solid lines) and simulated soil temperatures (shaded lines).
the season was within the soil frost boundaries inferred from the temperature measurements during winter 1995–1996 at Degerö Stormyr. The simulated soil frost disappeared 2 days earlier in spring 1996 than the soil frost in the mire, as inferred from the temperature measurements. During winter 1996–1997 the mire was almost free from soil frost, a pattern that the MMWH model successfully reproduced. During winter 1997–1998 the MMWH overestimated the depth of the soil frost slightly, and the simulated soil frost disappeared 12 days later than the soil frost according to the temperature measurements.

For the winter 1991–1992 the simulated soil frost for the validation site at Storfimyran closely followed the field measurements at the NS and SS sites. The simulated soil frost disappeared 7 and 15 days earlier than the measured soil frost from sites SS and NS, respectively. The simulated snow depth at Degerö Stormyr followed that measured at the meteorological station at Kulbäcksliden, which is in a relatively small sheltered field at the top of an adjacent hill (274 m asl, 4 m above Degerö Stormyr), during the winter of 1995–1996. In the following winter the simulated cover was somewhat deeper than that observed in the field (Figure 5). During winter 1996–1997, when snow depth measurements were also available from Degerö Stormyr, the simulated snow depth agreed well with measurements of snow depth available both from the actual mire and from the Kulbäcksliden meteorological station.

5. Discussion
5.1. Model Performance

The small difference in the model's performance between the calibration set and the validation sets is encouraging, especially since the driving data for the validation data set from Storåmyran had more gaps and the supplementary data had to be collected from a climate station farther away from the actual mire than was the case for the calibration site at Degerö Stormyr. It is also encouraging that the simulated deep temperatures are close to the measured values toward the end of the simulation, though it is impossible to obtain an accurate initial temperature profile when reconstructing historical long-term soil temperature and water table data series. The monitoring facilities at Storåmyran were not set up for model validation, and there was no soil temperature equipment monitoring the same vegetation type as the one studied at the Degerö Stormyr experiment site. Therefore we chose two profiles at the border of a carpet area to evaluate the MMWH model. This is a system dominated by floating S. majus mats, which within a certain range, move together with changes in the water table position. In order to evaluate the water table simulations, water table data from a lawn area, which more closely resembles the study area at Degerö Stormyr, were also included. Because of the difference in behavior between the two systems, the simulated water table was ~10 cm lower than the position recorded at the SS and NS sites during the driest periods. However, the MMWH model faithfully reproduces...
the general seasonal pattern of fluctuation of the water table observed at all three sites.

The large difference in soil temperature between the NS and SS sites is an effect of the difference in exposure between southern and northern microrelief slopes of high-latitude mires. The MMWH model, which simulates average conditions, conforms better with the behavior of the south side profile during the first part of the summer and with the north side profile during the later part.

In an initial simulation (not shown here) where the 10% correction term for rain precipitation was used over the whole simulation period the model underestimated the water table during the second half of the summer in 1996 and for the entire summer in 1997 and 1998 at Degerö Stormyr. This was assumed to be due to an increased error in the precipitation measurement caused by the movement of the tipping bucket to the higher and windier 1 m elevation. The wind-induced measurement error is commonly the largest component in the measurement error of precipitation [Bras, 1990]. An additional 10% correction term was added to the precipitation record after the date when the rain gauge was moved, which improved the simulation of the water table significantly. The exposed wind conditions at open mires tend to cause large spatial and temporal variations in measurement errors for precipitation. It is therefore critical for the performance of a hydrological model to evaluate thoroughly the precipitation record used if it is measured at the mire.

The main emphasis for this modeling exercise was to describe the water content of peat profiles as accurately as possible and thereby improve heat flux modeling. The ability of the derived water content function to simulate water content profiles closely resembling those measured by Hayward and Clymo [1982] is the basis on which this model is constructed. A correct water content description is necessary to be able to model heat flows and ice formation. The water content function also determines the position of the water table, which affects all the major contributing hydrological factors in the model. Frohling and Crill [1994] point out that their model had problems in simulating the water table position in their study at Sallies Fen, New Hampshire, because of a lack of data concerning the hydraulic properties of the peat. Most probably, their water table simulations would be improved by introducing the transmissivity and soil water content functions used in our modeling approach.

The MMWH model tends to be self-regulating; so, after events of overestimation or underestimation of the precipitation data inputs the model adjusts from the incorrect prediction toward the measured water table values. This self-regulating property of the model corresponds to the behavior of the real mire, where it helps the mire plant community maintain long-term hydrological stability. The property responsible for this behavior is the steep decline in hydraulic conductivity in the upper horizons [Ivanov, 1991]. The transmissivity is high when water tables are high since the conductivity changes by up to 4 orders of magnitude through the acrotelm. This allows the system to rapidly dispose of episodic large inputs of water following, for instance, snow melt or rainstorms, without causing significantly raised water levels in the long term. The low hydrological buffering capacity of the mire, caused by the thinness of the acrotelm to which the water table movement is confined [Ingram, 1983], is probably the reason why the mire water table can be accurately modeled without taking runoff from the surrounding watershed into consideration.

The close resemblance of the simulated water table positions to the levels measured at Degerö Stormyr when the floater was thawed and activated in the spring verifies the assumption that runoff during the spring from the mire is governed by the

### Table 3. Fit of Simulated Versus Measured Soil Temperature From the Validation Data Set From Storämyran 1991–1993

<table>
<thead>
<tr>
<th>Period</th>
<th>Deviation Simulated-Measured Variable$^*$</th>
<th>T 05 (06)$^b$</th>
<th>T 10 (10)</th>
<th>T 20 (20)</th>
<th>T 45 (45)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean s.d.</td>
<td>Mean s.d.</td>
<td>Mean s.d.</td>
<td>Mean s.d.</td>
<td>Mean s.d.</td>
</tr>
<tr>
<td>June 20 to Nov. 25, 1991</td>
<td>1.0 ± 1.0</td>
<td>0.8 ± 0.9</td>
<td>0.8 ± 0.7</td>
<td>0.7 ± 0.6</td>
<td></td>
</tr>
<tr>
<td>April 22, 1992, to Oct. 11, 1992</td>
<td>1.6 ± 1.8</td>
<td>1.5 ± 1.4</td>
<td>0.7 ± 0.6</td>
<td>1.5 ± 1.7</td>
<td></td>
</tr>
<tr>
<td>May 20 to Nov. 23, 1993</td>
<td>0.8 ± 0.7</td>
<td>0.7 ± 0.6</td>
<td>0.5 ± 0.5</td>
<td>0.4 ± 0.7</td>
<td></td>
</tr>
</tbody>
</table>

$^a$Measured temperature data from the northern site (NS).

$^b$Depth of temperature probe unit (centimeters); modeled depth within parentheses.
amount of ice in the upper part of the acrotelm. The MMWH model is able to simulate the behavior of the water table for spring 1996, when there was still an ice layer deeper down in the profile. Because of the lack of continuous measurements at the Storåmyran test site, the generality of the assumption above could not be tested. It is difficult to describe how runoff is affected by ice formation in the peat. However, the large amount of water that has to be drained from the mire at snowmelt, together with the rapid drop in transmissivity with depth in the profile, as discussed above, makes an exact description of the start of the runoff less important.

During development of the model several different approaches to calculating the thermal conductivity of the peat were tested. The commonly used geometric mean equation was tested, but it caused an underestimation of the heat flow when all constituents were included in the equation. Better model performance was achieved when a mixed model was applied, where the geometric mean equation was used to calculate the thermal conductivity for the subvolume consisting of the water, ice, and organic volumetric content. This subvolume was then introduced in a weighted arithmetic mean equation together with the air component. (For further discussion, see section 3.10.) Further development of an algorithm that describes the thermal conductivity in the unsaturated part would improve modeling of soil temperatures in mires.

5.2. Effect of Snow Pack on Soil Frost Penetration and Soil Temperatures

During the last part of the 1995–1996 winter and throughout the 1996–1997 and 1997–1998 winters at Degerö Stormyr, for which measurements of soil temperatures were available, the simulated temperatures below the ice closely followed those measured. This is encouraging for the possibilities of modeling breakdown of organic material and the accompanying accumulation of greenhouse gases under the ice during the long winter period at northern latitudes. With decreasing soil temperature the degradation of the peat decreases and is finally totally blocked in the frozen parts of the profile. Thus deep penetration of soil frost increases the net carbon storage in the mire by decreasing the production of carbon dioxide and methane during the winter period. At Degerö Stormyr the date of the disappearance of the ground frost varied between the 3 years by as much as 35 days, and at Storåmyran a variation of 9 days was found between the adjacent SS and NS sites in the same spring period. The large natural variation in soil frost thawing probably has a significant impact on the net primary production at the mire by delaying the development of the vascular plants because of the lower soil temperature caused by the presence of soil frost in or close to the rooting zone. The MMWH model simulated the date of disappearance of the soil frost at Degerö Stormyr for spring 1996 and 1997 with an accuracy of 12 days. This fit of the model can be compared with the 9 day difference between the modeled and actual dates of disappearance of the soil frost for the SS and NS sites at Storåmyran in spring 1992. To improve calibration and further develop the functions used to describe soil temperature and ice formation during the winter period, more detailed measurements of snow and soil frost development have to be obtained from the field sites.

To investigate the importance of the local climate during the long winter periods at high latitudes, the MMWH model was used in a response analysis where the amount of winter precipitation was varied. This gave rise to a remarkable change in the simulated ground frost pattern. By decreasing the simulated winter precipitation during the 1995–1996 winter at Degerö Stormyr by 75% the period with frozen ground is prolonged from June 8 to August 1 (Figure 9). With a doubling of the winter precipitation the soil frost disappears by May 6. There is also a large effect on how deep the soil frost penetrates the mire. Neither of these situations is unrealistic: ice lenses are sometimes found in parts of the mires at the end of July, and thawed mires can be found in May. The soil temperature at 10 cm depth in the scenario with 25% of the measured precipitation is clearly lower than in the normal precipitation scenario. The MMWH model is able to simulate the behavior of the water table for spring 1996, when there was still an ice layer deeper down in the profile. Because of the lack of continuous measurements at the Storåmyran test site, the generality of the assumption above could not be tested. It is difficult to describe how runoff is affected by ice formation in the peat. However, the large amount of water that has to be drained from the mire at snowmelt, together with the rapid drop in transmissivity with depth in the profile, as discussed above, makes an exact description of the start of the runoff less important.

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6. Conclusions

A physically based model for the soil thermal regime and water content of mixed mires, driven by air temperature and precipitation, has been developed and validated with 3 year long data sets from two different mires in northern Sweden.
The model was able to simulate water table position, soil temperature profiles to 3 m depth, and soil frost development with acceptable agreement to measured data over several annual periods with varying winter conditions. The results presented here show that it is possible to reconstruct soil temperatures and water table data from readily available standard meteorological data. The reconstructed data series can be used as driving variables in models for methane emission to evaluate factors such as the effect of climate on methane emission from boreal mixed mires. Using climatic data spanning 30–40 years, which probably cover most of the current natural climatic variability, would enable the maximal variation in annual methane emission from specific mire types to be evaluated. This study also shows the strong effects of snow conditions during the winter on thermal characteristics of the soil during the subsequent vegetation period and that it is possible to predict these effects satisfactorily using only air temperature and precipitation as driving variables.

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