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New decomposition rate functions based on volumetric soil water content for the ROMUL soil organic matter dynamics model

Tapio Linkosalo[a,*], Pasi Kolari[b], Jukka Pumpanen[b]

[a] Finnish Forest Research Institute, Jokiniemenkuja 1, Box 18, FI-01301 Vantaa, Finland
[b] University of Helsinki, Department of Forest Sciences, Finland

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ABSTRACT

The ROMUL decomposition model describes the flux of soil organic matter through the soil decomposition process separately for the organic layer (forest floor) and mineral soil. In the original ROMUL the effect of soil moisture on soil organic matter decomposition was described by using gravimetric soil water content, a parameter that is often difficult to obtain when the model is applied using continuous weather data. The new decomposition rate functions based on a soil respiration model use volumetric soil water content, replacing the original moisture functions with new ones. This paper also describes the development of a simple volumetric soil water model, with separate water storage compartments for the organic and mineral soil layers to better correspond with the structure of the ROMUL model. A volumetric soil model is also easier to adapt to various locations than the gravimetric soil water model.

The new decomposition rate functions employing the new soil water model were tested by re-simulating a previous application of the ROMUL model to a forest site representing conditions in Southern Finland. Whereas the original ROMUL model underestimated the steady-state soil carbon stock of the site, the new model structure considerably reduced the decomposition rates, and the revised soil carbon storage estimates are now in good agreement with the measured data.

1. Introduction

The ROMUL decomposition model (Chertov et al., 2001, 2007) is a practical tool for estimating the carbon storage of temperate and boreal forest soils. It handles the stores and flows of soil organic matter (SOM) as separate cohorts based on their origin (litter from needles/leaves, shoots, fine roots, coarse roots and stems). This structure is easy to link to stand growth models (for example Chertov and Komarov, 1995; Mäkipää et al., 2011). Nitrogen flux in the soil is described as an explicit mass flow, hence the model is also well-suited to describe the nitrogen flux between the soil and the stand. As nitrogen is the nutrient that mostly limits stand growth in the boreal zone (Helmisaari et al., 2011), this feature makes the ROMUL model well-suited to simulate forest nutrient balance in current and changing climate conditions, and under various management practices. See the appendix for a detailed description of the ROMUL model.

The decomposition of organic matter (i.e. heterotrophic soil respiration) produces a considerable part of CO_2 flux out of the soil. Therefore models of heterotrophic respiration can also be used as a proxy to describe the decomposition rate of soil organic matter. Skopp et al. (1990) describe how soil water content limits the decomposition rate: in dry soil the lack of water slows decomposition, while high soil water levels slow down the diffusion of oxygen into the soil, thereby also limiting the decomposition rate. Between these two there is an optimal point or range of soil water content around which the decomposition rate reaches a maximum; this roughly coincides with the field capacity. Pumpanen et al. (2003) used the model by Skopp et al. (1990) as a basis for estimating a function of *relative* decomposition rate for a boreal forest site in Southern Finland. Their model adds a constant limit of the maximum decomposition rate to the two limiting functions of Skopp et al. (1990). Both these models can also be used to describe how the decomposition rate depends on the soil water content.

Soil humidity as the driver of decomposition in the original ROMUL model is based on gravimetric water content. To calculate the gravimetric soil water content rather detailed soil data is required; in addition to soil depth and water retention capacity, soil density and the absolute values for soil water content at the field capacity and wilting point are also needed. Modelling the soil water content with volumetric measures would be much simpler, as basically the volume of the total water storage of the soil layer would suffice to describe the soil properties affecting the water dynamics.

Volumetric soil water models, in their simplest form, consider the soil as a single vessel holding the soil water. Such so-called "bucket" models, originally suggested by Manabe (1969), have been

* Corresponding author. Tel.: +358 40 8015583; fax: +358 10 211 2103.
E-mail address: tapio.linkosalo@metla.fi (T. Linkosalo).

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widely used in recent years to describe the soil water dynamics in models of evapotranspiration, used to test eddy-covariance measurements (Schwärzel et al., 2009; Scott, 2010; Naranjo et al., 2011). In these models, the soil water storage is filled with precipitation and depleted by evapotranspiration. As the soil water varies between the wilting point and field capacity (more refined models also take the water volume up to saturation into account), these two figures are the primary soil variables that are needed to be known. As a matter of fact, only the difference between the two suffices; the absolute water content of the soil is not needed.

The ROMUL model separately describes the pathways of SOM for the litter originating from aboveground sources, and for the underground litter. The former is proposed to decompose in the organic layer of the forest floor, and the latter in the mineral soil layer. The environmental conditions that drive the decomposition rate (primarily soil humidity and temperature) are rather different for the two layers, therefore in this paper we developed a slightly more complex model of the soil water dynamics, separately describing water content for the two soil layers and consecutive differences in the decomposition rates.

The aim of this study was to use the decomposition rate functions derived from Pumpanen et al. (2003) to describe the effect of soil water levels on the decomposition rate, and to develop a volumetric, two-storage soil water model that predicts the soil water dynamics in both the organic layer and mineral soil. Both the soil water model and the decomposition rate functions were then linked to the ROMUL decomposition model to replace the corresponding elements of the original decomposition model. Finally, the revised model was tested on a case study in a stand in Southern Finland to see how the modifications affect the model predictions of steady-state soil carbon stock.

2. Materials and methods

2.1. Decomposition rate functions

The decomposition rate modifier, f, which depends on the soil moisture content θ_v (in mm mm^{-1}), was taken from Pumpanen et al. (2003). The rate function consists of two parabolic sections and a maximum limit as follows:

$$f(\theta_v) = \min \begin{cases} \alpha \cdot \theta_v^d \\ \beta \cdot (P_0 - \theta_v)^g \\ 1 \end{cases} . \qquad (1)$$

Here P_0 is the total soil porosity (in mm mm^{-1}). The empirical parameters α, β, d, g were taken (as in Pumpanen et al., 2003) from Skopp et al. (1990). Variations in the value of total soil porosity, P_0, due to the varying composition of the soil matrix is reflected in the different patterns of the decomposition rate function, especially in the range of soil water contents where the decomposition rate function reaches the value of unity. This variation is most evident between the organic and mineral soil layers, the difference between the two in our case study is seen in Fig. 1, based on measurements from the SMEAR II site in Southern Finland (Ilvesniemi et al., 2010).

The physical parameters of soil water content were estimated with the model presented by Saxton and Rawls (2006). The equations of the model give the values of water storage (volumetric values for field capacity and wilting point) for the soil as a function of sand, clay and organic matter content. The soil water storage measures are as follows:

$$\theta_{fc} = 0.2576 \pm 0.002 C_s + 0.0036 C_c + 0.0299 C_o \qquad (2)$$

$$\theta_{wp} = 0.026 + 0.005 C_c + 0.0158 C_o \qquad (3)$$

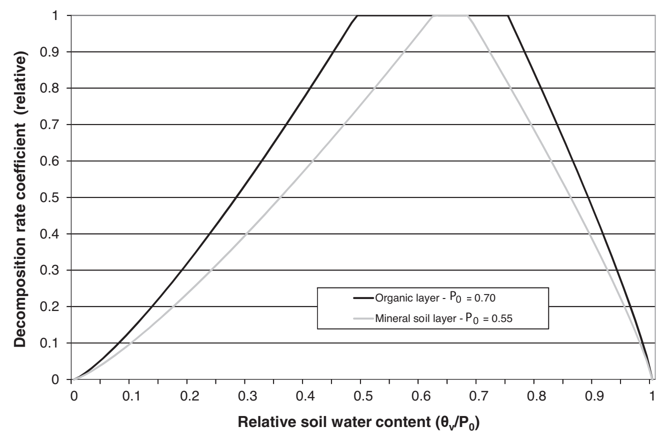

Fig. 1. The decomposition rate functions as function of the ratio of soil water content, θ_v, to the total soil porosity, P_0. The values for P_0 are from measurements at the SMEAR site. Black line represents the organic layer and grey the mineral soil layer. Value of 1 indicates no water limitation for decomposition, and value of 0 stops the decomposition altogether.

where θ_{fc} is the soil volumetric water content (cm^3 cm^{-3}) for a potential of −0.33 bar (field capacity) and θ_{wp} is the water content for a potential of −15 bar (wilting point). C_s, C_c and C_o are the weight percentages of sand, clay and organic matter in the soil, respectively. The forest floor/organic layer was assumed to consist of only organic matter.

2.2. Soil water model

The key features of the soil water model presented here include the following: (1) separate water storage compartments for the organic soil layer and the mineral soil layer and (2) the description of soil water content with volumetric measures. The first feature is intended to reflect the structure of the ROMUL decomposition model, where SOM decomposition takes place explicitly in the two separate soil layers. The second feature allows to make soil water modelling easier; with a volumetric soil water model only the water storage capacity is needed to describe the soil characteristics, and this can be quite easily estimated, for example, from the soil texture and the depth of soil layers.

The basic structure of the soil water model is similar to a simple "open bucket" type model where soil is considered as a vessel holding the soil water. This vessel is filled by precipitation and emptied by evapotranspiration, with excess water draining out of the system. In the present model there are two separate water storage compartments, one for each soil layer, and therefore the water dynamics between the two stores also need to be considered. In the forest the organic layer lies on top of the mineral soil, and this structure is also reflected in the model structure. The primary principles of the two-storage model are as follows:

1. There are separate water storage compartments for the organic layer and the mineral soil layer.
2. Soil water content is presented in volumetric units.
3. Precipitation fills the organic layer.
4. Drainage from the organic layer fills the mineral soil layer.
5. Drainage from the mineral soil layer is drainage out of the system.
6. For both layers there is a parameterised time constant for the drainage; this determines the rate at which water in excess of the field capacity is drained.
7. Water is released from both soil layers; plant root water uptake from both layers, and in addition evapotranspiration from the organic layer. The balance between the two components is

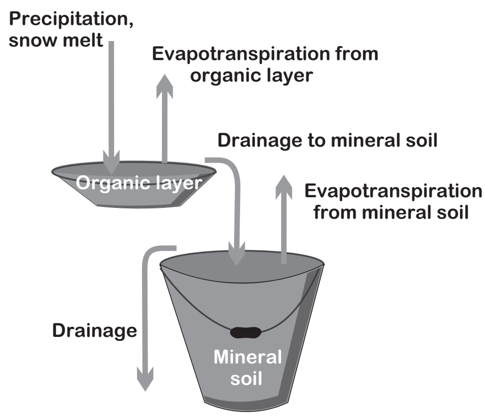

Fig. 2. Water flow in the 2-storage soil water model.

parameterised. If the required water release from the organic layer is higher than the remaining soil water content in the layer, then the remainder is taken from the mineral soil layer.

8. The time step of the model is one day in order to respond to rapid variations in the soil water content and corresponding changes in the decomposition rate.

   The following was also considered, but turned out not to be needed:

9. There is a capillary flux between the two storages, from higher relative moisture to lower, and the rate of the flux depends on the difference between the relative water contents.

The model does not explicitly define the depth of the soil layers, but these relate to the water holding capacity. Therefore the mineral soil layer of the model is implicitly assumed to cover the rooting layer of the plants. The water flows of the model are presented in Fig. 2.

The following equations describe the water fluxes in the soil water model. First, the change in the organic layer storage, W_O, is:

$$\Delta W_O = P - E_O - D_O \tag{4}$$

where P is the precipitation, E_O is evapotranspiration and D_O drainage from the organic layer. All these measures are in volumetric units, in other words the water column thickness is in mm. Evapotranspiration from the organic layer is a fixed ratio of the total evapotranspiration, E_{Tot}, but cannot exceed the total water storage in the organic layer. It is represented by the following equation:

$$E_O = \max(a \cdot E_{Tot}, W_O) \tag{5}$$

where a is a model parameter (evapotranspiration ratio) that splits the overall evapotranspiration between the organic layer and mineral soil.

Drainage out of the organic layer, D_O, is defined as follows:

$$D_O = \begin{cases} 0, & W_O < FC_O \\ \dfrac{W_O - FC_O}{\tau_O}, & W_O \geq FC_O \end{cases}. \tag{6}$$

Here, τ_O is the time constant for drainage in days. All the other variables are of volumetric storage, given in mm. The time constant parameter can be interpreted as the time it takes for water in excess of the field capacity to drain from the organic layer. In the implementation of the model drainage was only calculated after evapotranspiration had occurred. An additional parameter for saturation, beyond which the water immediately drains, was also tested but found unnecessary for the model.

Equally, the change in water storage for the mineral soil layer, W_M, is:

$$\Delta W_M = D_O - E_M - D_M \tag{7}$$

where D_M is the drainage from the mineral soil layer. Plant water uptake from the mineral soil, E_M, is:

$$E_M = \max(E_{Tot} - E_O, W_M) \tag{8}$$

In other words, evapotranspiration cannot exceed the remaining water storage. Drainage from the mineral soil, D_M, is:

$$D_M = \begin{cases} 0, & W_M < FC_M \\ \dfrac{W_M - FC_M}{\tau_M}, & W_M \geq FC_M \end{cases}. \tag{9}$$

Here, τ_M is the time constant for drainage from the mineral soil layer (in days). All the other variables represent water storage, in mm. Furthermore, the occurrence of drainage from the mineral soil water storage was calculated after first deducting the evapotranspiration from the storage.

2.3. Evapotranspiration model

The stand evapotranspiration was estimated with an empirical model using meteorological variables. The model parameters were estimated with evapotranspiration data from the SMEAR II eddy covariance site at the Hyytiälä Forestry Field Station in Southern Finland (Ilvesniemi et al., 2010). The evapotranspiration, E_{Tot} (in mm), is calculated from meteorological variables as follows:

$$E_{Tot} = E(\Phi) \cdot f_T(T) \cdot f_{VPD}(D) + a_1. \tag{10}$$

The $E(\Phi)$ function defines evapotranspiration rate depending on photosynthetically active irradiation (PAR), Φ (in $\mu mol\, m^{-2}\, s^{-1}$). We estimated this with a linear equation:

$$E(\Phi) = E_0 \cdot \Phi \tag{11}$$

where E_0 is a constant. This evapotranspiration is then scaled with temperature and water vapour pressure deficit factors. The temperature function, f_T, is a moving average of temperature with a lower limit:

$$f_T(k+1) = \max \begin{cases} f_T(k) + \dfrac{T_{Air} - f_T(k)}{\tau_{ET}} \\ 0 \end{cases}. \tag{12}$$

Here k refers to the time step, T_{Air} is the air temperature (°C) and τ_{ET} the time constant (in days) for the f_T function. The model calculation was started on 1st January, with the temperature coefficient initialised to −0.25 °C for the organic layer, and 1.22 °C for the mineral soil layer.

The evapotranspiration model also considers the effect of water vapour pressure deficit (D, in Pa) on stomatal conductance, to describe how plants optimise their carbon gain per amount of water lost in evapotranspiration. This control is described by the following exponential function:

$$f_{VPD}(D) = e^{\kappa_{ET} \cdot D} \tag{13}$$

where κ_{ET} is a constant. Irradiance and temperature also affect stomatal conductance, but in our model these effects are accounted for in the corresponding factors ($E(\Phi)$ and f_T).

The parameter a_1 adds a small base level of transpiration, which is seen in the Eddy covariance measurements during winter months, when our model would due to low temperature and/or irradiance otherwise show zero transpiration.

2.4. Soil temperature model

We used a daily time-step soil temperature model to simulate soil temperature as a driver of the decomposition rate in ROMUL. This model is a simplification of the heat flux models commonly used for soil temperature simulations (e.g. Jansson and Moon, 2001). We estimated the soil temperature with a model of the moving average of the air temperature, separately for the two soil layers, as follows:

$$\Delta T_{\mathrm{Soil}} = \max \begin{cases} \dfrac{T_{\mathrm{Air}} - T_{\mathrm{Soil}}}{\tau} \\ T_{\min} - T_{\mathrm{Soil}} \end{cases} \tag{14}$$

where T_{Air} is the measured air temperature, T_{Soil} is the predicted soil temperature, τ is a time constant and $T_{\min}$ is the lower limit of the soil temperature. All temperatures are in (°C), and the time constant in (days). Unlike the heat flux models, we used the air temperature as an independent variable for both soil layers.

2.5. Measured data, model parameterisation and sensitivity analysis

Data collected at the SMEAR II station in Hyytiälä, Southern Finland (Vesala et al., 2005) in 1997–2011 was used to estimate the parameters of the various submodels. As in Rantakari et al. (2012), the SMEAR II station was also used as the representative site for the Southern Finland for the steady-state soil carbon balance calculations.

For the evapotranspiration submodel, daily means of air temperature and relative humidity, as well as the daily sum of PAR and precipitation were used as the independent variables. The predicted values of evapotranspiration were compared to data measured with the Eddy covariance method from a measuring tower over the study area, with the data quality checked and missing data gap-filled with standard procedures (Vesala et al., 2005; Mammarella et al., 2009). The parameters were estimated by minimising the sum of squared model residuals. We also conducted a Monte Carlo analysis on the parameters to estimate the model parameter sensitivity.

The empirical values for soil water content were measured with a time-domain reflectrometry (TDR) method using unbalanced steel probes installed in all soil horizons in five locations. The permanent probes installed in the illuvial horizon (B horizon) were used to represent the soil water content in the mineral soil layer, and measurements in the organic layer (O-horizon) were used for modelling that layer. Each daily measurement of soil water content was an average of the TDR readings from 4 to 5 locations and over the 48 daily measurements; with obvious outliers being omitted. Temporally missing observations were linearly interpolated from adjacent valid measurements. TDR measurements in freezing conditions were considered unreliable, as the dielectric properties of ice and water are highly different. For this reason, the model was only evaluated on days when the measured soil temperature was above 0.5 °C.

The soil water holding capacity was estimated with the Saxton and Rawls (2006) model, using values for the soil matrix composition of the SMEAR site published by Ilvesniemi et al. (2010). Thus for the actual soil water model we only had to estimate the values of 3 parameters, namely the saturation drainage time constants for organic and mineral soil layers, τ_H (day) and τ_M (day) correspondingly, and the evapotranspiration ratio, a (i.e. the split of root water uptake between the organic and mineral soil layers).

The values of the parameters were estimated with a Monte Carlo simulation, testing 65,000 randomised triplets of the parameters and calculating the corresponding model root mean squared error (RMSE) as the difference between the daily values of observed and modelled soil water contents for both soil layers. The RMSE values were calculated as the difference in absolute soil water content (in mm). As the water storage in the organic layer is much smaller than in the mineral layer, weights of 5 and 1 were given to the model prediction errors of the two layers, respectively; otherwise the mineral soil water compartment would have dominated the model fitting procedure.

Finally, empirical parameters for the decomposition rate functions were not estimated, but were taken from Skopp et al. (1990).

2.6. Soil carbon simulations for Southern Finland

In a recent paper by Rantakari et al. (2012), the soil decomposition models Yasso (Liski et al., 2005), Yasso07 (Tuomi et al., 2011) and ROMUL (Chertov et al., 2001, 2007) were compared and tested against empirical soil carbon data measured from Biosoil sites in Southern Finland (Mäkipää and Heikkinen, 2003). Rantakari et al. (2012) used the ROMUL model as presented by Chertov et al. (2001, 2007), and found that the model underestimated the steady-state soil carbon storage compared to the results of the Yasso models, as well as the Biosoil measurements. In this paper we re-ran the ROMUL model to a steady state condition using the same soil characteristics, weather conditions and litter input as given in the paper by Rantakari et al. (2012). However, in this simulation we used the new decomposition rate functions with the new soil water model, and compared the results to those of Rantakari et al. (2012).

3. Results

3.1. Model parameter estimation

The values of the parameters for the evapotranspiration model were estimated by fitting the models to the Eddy-covariance measurements from the SMEAR II station for the period 1997–2011 (Vesala et al., 2005). We estimated the model sensitivity with a Monte Carlo analysis, producing 32,000 parameter sets and calculating the standard deviation of the 200 parameter combinations producing the smallest model RMSE. The values and relative standard deviations were $E_0 = 0.000416$ (mm), 1.6%; $\tau_{ET} = 7.22$ (days), 9.2%; $x_0 = -0.86$ (°C), 2.1%; $\kappa = -0.000394$ (Pa^{-1}), 7.1% and $a_1 = 0.24$ (mm), 5.5%. The R^2 of the evapotranspiration model was 0.86. *XY*-plot of modelled vs. measured evapotranspiration data indicate that the model has practically no bias (Fig. 3). We also tested the Penman–Monteith (1965) evapotranspiration model, which gave a R^2 value of 0.68, but chose to use our simpler model as it gave slightly better results.

The soil water storage measurements for the SMEAR II site, as derived with the Saxton and Rawls (2006) model, were as follows:

- field capacity of the organic layer, $FC_O = 14.4$ mm
- wilting point of the organic layer, $WP_O = 2.1$ mm
- field capacity of the mineral soil layer, $FC_M = 128.0$ mm
- wilting point of the mineral soil layer, $WP_M = 13.1$ mm.

The measured values for the water capacity of the whole soil layer are 200 mm (in saturation) and 50 mm (at the wilting point) (Ilvesniemi et al., 2010). Assuming that the field capacity is roughly 70% of the total porosity of the soil, these figures would yield a water storage of 105 mm (0.7 times the difference between field capacity and wilting point) in the soil at the SMEAR II station, which corresponds well with the capacity calculated by the Saxton and Rawls (2006) model.

The parameters for the soil temperature model were estimated by fitting the models to the soil and air temperature measurements

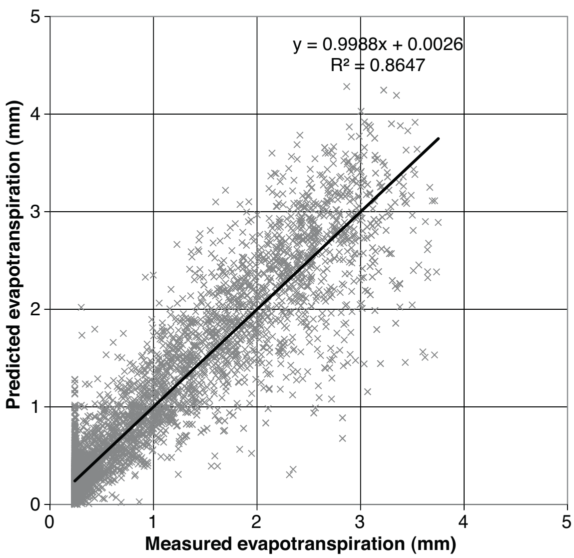

Fig. 3. *XY*-plot of the measured vs. modelled evapotranspiration (in mm).

from the SMEARII station. The values for the soil temperature model parameters, τ and T_{min}, were 14.9 days and −0.13 °C, respectively for the organic layer, and 10.5 days and 0.24 °C, respectively, for the mineral soil layer. Our soil temperature model is much simpler than the common heat flux models, yet the R^2 values between the observed and predicted values were 0.93 and 0.97 for the organic and mineral soil layers, respectively. One should bear in mind, however, that such empirical model may not work in much different conditions than where it is parameterised to.

3.2. Simulated soil water levels

The values for the three estimated parameters of the soil water model, giving the smallest RMSE of model fitting, were: $\tau_O = 0.90$ (days), $\tau_M = 7.97$ (days), $a = 0.25$. RMSE values for the organic and mineral soil layers were 4.34 and 3.67, respectively, and the corresponding R^2 values 0.09 and 0.51. This is a considerable improvement over the CLISS model (Chertov et al., 2001) used in the original ROMUL model, which for the SMEAR site reached RMSE values of 12.0 and 26.6 days, respectively. The R^2 values for CLISS were of similar magnitude as our new model, 0.05 and 0.49, suggesting that the model is able to reproduce the daily variation in almost similar accuracy, but shows a systematic bias in the overall level of soil water content.

The small storage and rapid dynamics of the thin (50 mm) organic layer turned out to be difficult to reproduce, while the water model performed much better for the mineral soil layer, where the major water storage resides. The small time constant value for the organic layer means that the layer immediately drains of water beyond the field capacity (with the time step of the model being 1 day). This suggests that the value of the organic layer field capacity, FC_O, may be a slight underestimate. In the mineral layer it takes 8 days to fully drain the water in excess of the field capacity. The evapotranspiration ratio, a, suggests that the water abstraction from the mineral soil layer is about 3 times larger than the water abstraction from the organic layer.

Fig. 4 shows the simulated soil water level pattern for period from 2003 to 2011, together with the measured values derived from TDR readings. Especially for the organic layer, the model shows a higher variation than the measured data. This may be partially due to difficulties in measuring the organic layer, where the depth and quality of the soil matrix is more heterogeneous than that of the mineral soil. The 15-day moving average of the model predictions shows a much better correspondence to the measured pattern of dry and moist periods. In the mineral soil layer, the variability in daily values is much lower, which is as expected: the water storage in that layer is 10 times larger than in the organic layer. Also, the surface area of the roots is highest in the organic layer (Ilvesniemi and Liu, 2001), probably resulting in higher fluctuations in water content due to the transpiration of the trees and small water storage. For the mineral soil layer, in most years the modelled soil water content fit well the observations. There is a pronounced difference in the overall level of values for some of the years; however even in those years the pattern of variation in the mineral soil water levels seems similar. Grayson and Western (1998) found that TDR probes placed at different locations have rather different temporal patters.

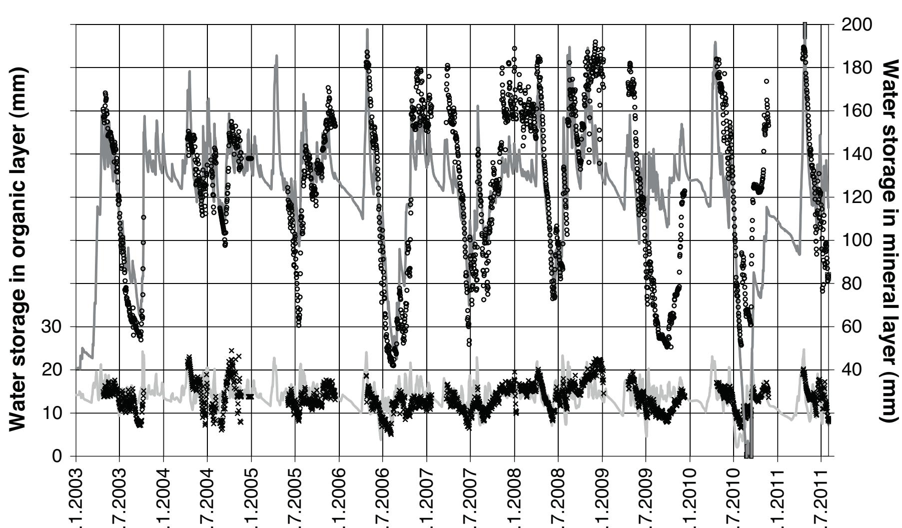

Fig. 4. The modelled and measured soil water values for the period 2003 to 2011. Dark and light grey lines are the 15 day moving averages for the modelled soil water content for the mineral (scale to the right) and organic (scale to the left) layers, respectively. Black circles and crosses are measured values for mineral and organic soil layers, respectively. Measured data from days when soil temperature was below 0.5 °C was omitted.

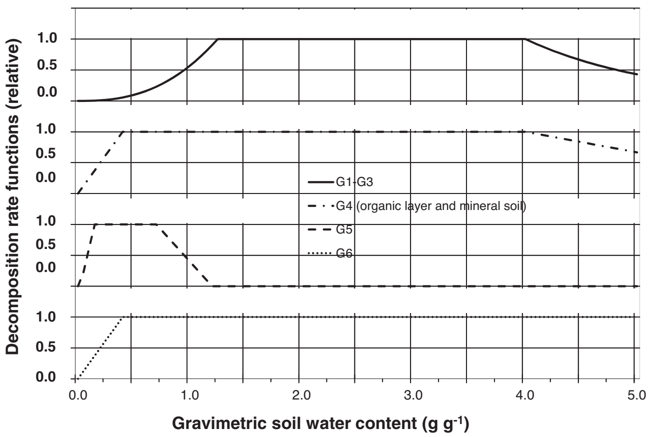

Fig. 5. Decomposition rate functions, used to scale down decomposition, depending on soil water content for the ROMUL model; the "new" rate functions as published by Chertov et al. (2007). Functions G1 to G3 (solid line) are for the transition from fresh litter to partly decomposed SOM, G4 (alternating line) and G5 (dashed line) from partly decomposed litter to (semi)stable humus, and G6 (dotted line) portrays the decomposion of the stable humus layer.

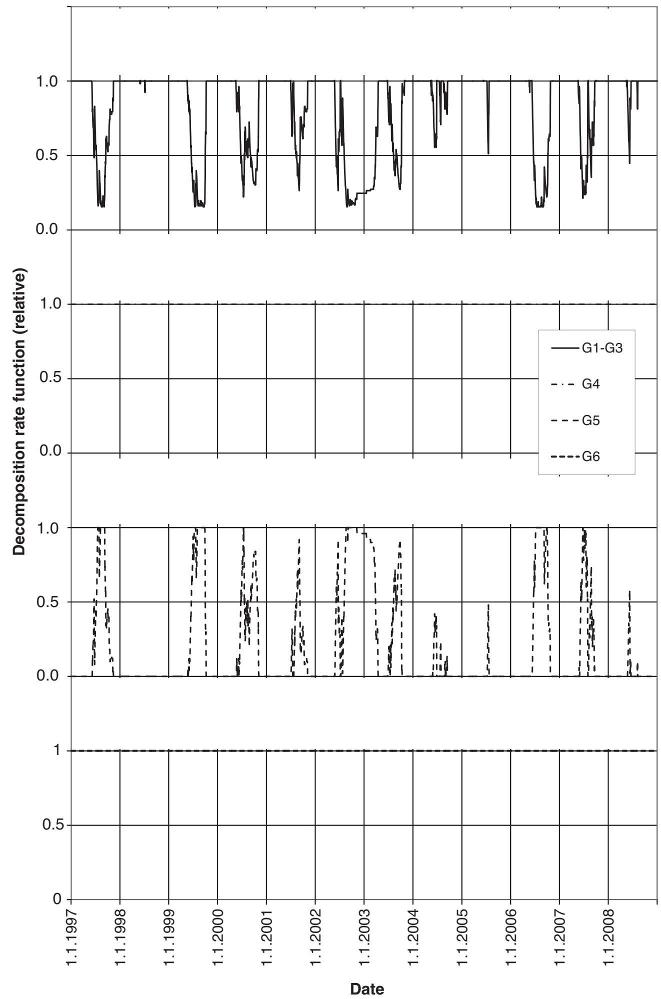

Fig. 6. Decomposition rates as a function of time, scaling down the decomposition due to soil water constrains, calculated with meteorological data and using the original ROMUL equations (Chertov et al., 2007). Functions G1 to G3 (solid line) are for the transition from fresh litter to partly decomposed SOM, G4 (alternating line) and G5 (dashed line) from partly decomposed litter to (semi)stable humus, and G6 (dotted line) portrays the decomposion of the stable humus layer.

It may be that the difference in overall level between different years in the SMEAR II TDR data also reflects similar shifts in level because of spatial variations in the root distribution, microtopography of the soil, as well as the maintenance of the instrumentation. If these level differences were accounted for, the R^2 value of the soil water model for the mineral soil layer would probably have improved from the current value of 0.51.

3.3. Decomposition rate functions

The original ROMUL model uses decomposition rate functions that are defined separately for each SOM flux, from raw litter to fermented litter and later to stable humus, as well as out of each storage (Chertov et al., 2001, 2007). These functions are further modified with rate functions based on soil temperature and water content (Fig. 5). Even though the pathways of organic matter are different for litter originating from above- and below-ground, the original ROMUL model uses the same decomposition functions for both organic and mineral soil layer litter. The only potential difference in decomposition rates would be a result of the different soil water content of the organic and mineral soil layers. As a matter of fact, even the water dynamics for different layers are similar in the original ROMUL water model (Chertov et al., 2001). In the present study, the rate functions that modify the transition rates depending on soil moisture were different for the organic and mineral soil layers due to the different water dynamics of the two soil layers. Only the final transition to the stable humus layer and its decomposition are considered to take place in the mineral soil layer for litter of all origins, just as in the original ROMUL model.

The soil water dynamics and the corresponding decomposition rate functions derived in this paper show a rather different temporal behaviour compared to the original one, when calculated using data from the SMEAR II station. Fig. 6 shows the values of the 6 rate functions of the original ROMUL model calculated over the simulation period. The rate functions G1–G3 (decomposition of fresh litter and partly decomposed litter) are all identical, and show some slowing down during the summer drought periods. During moist summers there is minimal reduction in the decomposition rate. However, the rate functions for the further decomposed organic matter, G4 and G6, have a practically constant value of unity, indicating no limitation in the decomposition rate in the later phases of decomposition due to soil water content.

For the new decomposition functions presented in this paper, the picture is quite different. Calculated separately for the organic and mineral soil layer, the decomposition modifier for the organic layer shows a considerable day-to-day variation, as can be expected from the shallow water storage and thus extreme water dynamics in that layer (Fig. 7A). The variation of the decomposition rate modifier in the mineral soil layer is much lower, but the summer period with a lower level of soil water slows down the decomposition rate considerably (Fig. 7B). In contrast, in the original ROMUL model the rates G4 and G6 in particular show no drought effect at all (Fig. 6). The new decomposition rate functions developed in this paper do not differentiate the fluxes between different stores, but are the same for all the rates, only differing for the rates between organic and mineral soil layers.

3.4. Soil carbon storage for Southern Finland

Rantakari et al. (2012) calculated the steady state carbon storage for Southern Finland with different decomposition models. The two versions of the Yasso model yielded steady-state storage exceeding 60 Mg[C] ha^{-1}, which corresponds well with the estimate of

Table 1
Soil carbon content for average Southern Finland forest stand.

Model/measurement	Average soil carbon Mg[C] ha^{-1}	95% confidence interval
Biosoil	66.9	[31.2–102.6]
Yasso	65.6	–
Yasso07	66.5	[53.5–79.5]
Original ROMUL	36.1	–
ROMUL with new water model and decomposition rate functions	58.3	–
ROMUL with new decomposition rate functions	56.7	–

From Rantakari et al. (2012).

carbon storage measured at the Biosoil project (Table 1; for details see Rantakari et al., 2012). On the other hand, the ROMUL model with the original decomposition rate functions, gave a considerably lower value for the storage, of only 36.1 Mg[C] ha^{-1}. A repeat of this study, using the new soil water model and decomposition rate functions, changed the outcome, with a new steady-state value of 58.3 Mg[C] ha^{-1}. The new result is well within the confidence limits of the empirical Biosoil measurements, and in a good agreement with the Yasso model results. We also repeated the simulation with a simplified version of the soil water model, in which the decomposition rate of the organic layer was calculated from the water content of the mineral soil layer, in other words the decomposition rate was simulated as if there was only one water storage. The resulting steady-state value from this simulation, of 56.7 Mg[C] ha^{-1}, suggests that the majority of the increase in the steady-state carbon storage in the ROMUL model simulation is due to the new decomposition rate functions. The increase due to simulating the water dynamics separately for the two layers is not so significant.

4. Discussion

In some earlier model comparisons, the original ROMUL model has considerably underestimated the steady-state soil carbon stores compared to measured data (Palosuo et al., 2012; Rantakari et al., 2012). Our results suggest that the reason for this may lie in the rate functions that adjust the decomposition rate based on the soil water content. With the modifications presented, the more restricting rate functions slowed down the decomposition, which in turn increased the storage in the steady state situation, to a level that fits well with the measured data and also agrees with other model results.

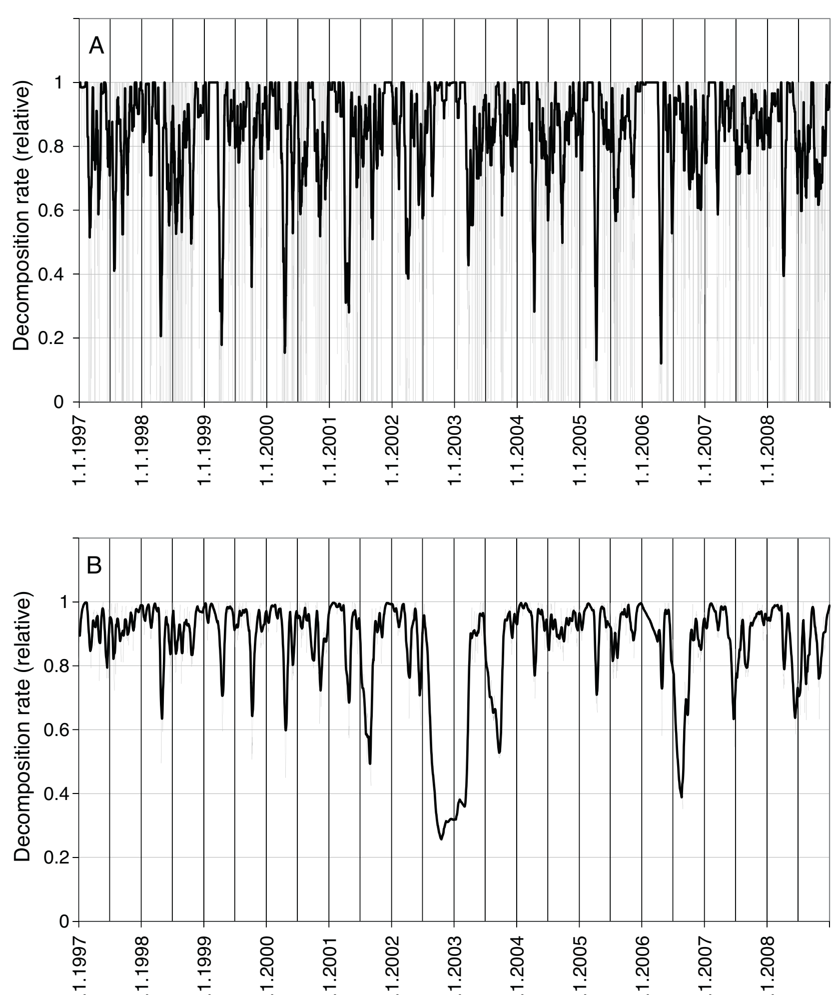

Fig. 7. Decomposition rates as a function of time, scaling down the decomposition due to soil water constrains, calculated with the new rate functions and soil water model, for both the organic layer (A) and the mineral soil (B). The grey lines show the daily values and the black lines a 15-day sliding average of the decomposition rate.

Decomposition in the soil is hard to evaluate quantitatively. The most common experimental method for evaluating decomposition is the so-called litter bag method, where the matter loss from litter material contained in mesh bags placed in the soil is weighed after the bags have been placed in the soil for various periods of time. The interpretation of these results is somewhat problematic, mainly because distinguishing between the decomposing material, surrounding soil medium and, for example, decomposers grown into the bags, adds to the uncertainty of the results (Cotrufo et al., 2009). However, the decomposition process taking place in the soil also produces CO_2, the flux of which has been measured in a number of experiments and several models of soil respiration have been produced based on the flux measurements. These models of CO_2 flux from the soil can therefore be used to describe the rate of decomposition in the soil. This method also has its shortcomings, since decomposition is not the sole source of CO_2 from the soil, but root respiration also provides its share (Högberg and Read, 2006). Moreover, the ratio of root respiration and decomposition activity is not constant, but varies over time (Pumpanen et al., 2008). The CO_2 flux due to decomposition can be described with theoretical models of the soil environment, and these models have been calibrated with empirical experiments where the soil has been cleared of plant roots, leaving only the decomposition processes to produce CO_2.

The decomposition rate models based on CO_2 flux alone produce the rate functions for the processes releasing carbon as gaseous CO_2. Therefore, when using CO_2 flux as a proxy we assume that the relationships between CO_2 release and mass flows from one C storage depends in a similar manner on the environmental conditions. Using such an assumption, as we did in this study, is a simplification compared to the original ROMUL model where the coefficients modifying the rate functions depending on soil water content are separately estimated for the different phases of the decomposition process. On the other hand, our simplified approach seems to work well for the case studies presented here for the SMEAR II station in Southern Finland. Based on these results, the use of similar rate functions for all the fluxes seems applicable.

Soil hydrology in a boreal or temperate forest can be described with a variety of models, from single storage models, such as the one presented by Kellomäki et al. (2010) or the soil water part of the original ROMUL model (Chertov et al., 2001), to complicated models with multiple soil layers and a complex structure of water fluxes between them, such as the Coup (Jansson and Moon, 2001), LPJ-Guess (Hickler et al., 2008) and ORCHIDEE/SECHIBA (Ducoudé et al., 1993) models. Models with only a single water storage compartment cannot distinguish between the different soil water characteristics of the organic layer and the mineral soil. On the other hand, adding more layers and fluxes to the model quickly adds to its complexity, which in turn increases the calculation time and complicates the parameterisation of the model. The model proposed in this paper is only slightly more complex than the single storage model and has only a few parameters. Yet the 2-storage model is more compatible with the ROMUL decomposition model, which distinguishes between the decomposition taking place in the organic and mineral soil layers. In an earlier version of the soil water model we also tested capillary flux between the two storages, from higher to lower moisture, but it turned out to be an unnecessary feature in the model (with a very low rate coefficient of the flux), and was therefore omitted from the final version of the model.

Many soil decomposition models, for example Yasso (Liski et al., 2005) and Yasso07 (Tuomi et al., 2011), utilise a monthly time step. Even though this makes the simulation of the decomposition process easier from a calculation time point of view and provides many more datasets with which to test the models, it also poses the risk of losing some essential characteristics of the decomposition process. The dependence of the decomposition rate on soil water content is non-linear, with the maximum rate of decomposition taking place in the middle of the water content range. Therefore, averaging the soil water content over longer time periods tends to underestimate the limiting effect of the hydrological conditions on the decomposition, which may result in overestimating the decomposition rate and, consequently, underestimating the steady state carbon storage in the soil. This behaviour was also reflected in a recent study by Rantakari et al. (2012), where the simulations of the ROMUL model were tested with weather data averaged over several years. The averaging of data over longer time periods evens out the precipitation patterns, as a consequence of which the limitation set by the soil water on the decomposition may be reduced, and this results in a reduced steady state storage of soil carbon.

5. Conclusion

The decomposition rate functions and the corresponding soil water model presented here considerably improved the soil carbon storage estimates. In addition, their application makes the modified ROMUL model much easier to apply, as a less detailed and a more general characterisation of the soil water holding properties is sufficient to be able to run the model. This makes it possible to use the model even for sites where the soil characteristics have not been measured in detail.

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Appendix. Description of the ROMUL decomposition model

The ROMUL model describes separate fluxes of soil organic matter (SOM) based on its origin (litter from leaves/needles, shoots, fine roots, coarse roots and stems). The model also deals with the flux of nitrogen, and for this purpose the model actually simulates a parallel mass flow for the nitrogen, with a similar structure of storage and flux as that of the SOM. The nitrogen content, calculated as the ratio of corresponding stores of nitrogen and SOM, affects the decomposition rate in many phases of the decomposition process (Fig. A1).

The fresh litter originating from the trees enters the model as stores of undecomposed litter ("$L_{C,i}$" for SOM and "$L_{N,i}$" for the corresponding N, in Fig. A1; "i" refers to the separate cohorts of litter from different origins). Part of this litter is metabolised by decomposing organisms, and the corresponding carbon is released as gaseous CO_2 (flows $C_{1,i}$ in Fig. A1). The majority of the SOM proceeds to a more slowly decomposing complex of partly decomposed organic matter and humus (flows $C_{2,i}$ and stores "$F_{C,i}$" in Fig. A1). Some of the material in the "$F_{C,i}$" store is also metabolised by the decomposing organisms, releasing a further amount of gaseous CO_2

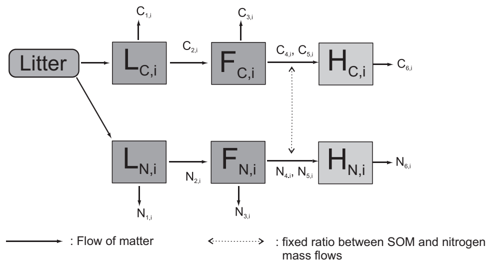

Fig. A1. A schematic presentation of the organic mass flows in the ROMUL decomposition model.

(flows $C_{3,i}$ in Fig. 3). Parallel to the flux of SOM, nitrogen in the litter is stored in storage $L_{N,i}$, and it flows, as a result of decomposition, into storage $F_{N,i}$. The nitrogen fluxes out of the system ($N_{1,i}$, $N_{3,i}$ and N_6), corresponding to the release as gaseous CO_2 from the decomposition of SOM, are considered mineralised nitrogen.

The contents of the $F_{C,i}$ stores eventually ends up in the immobilised humus store ("H_C" in Fig. A1). Storage "H_C" is common for all the fluxes of SOM, and, independent of the origin of the litter, resides in the mineral soil layer. This is because the SOM in this storage (and the corresponding N storage) is considered to be produced by the dead biomass of the decomposing organisms, namely *Bacteria*, *Arthtropoda* and *Lumbricidae*.

All the mass flows are considered to be ratios of the originating storage content, depending on the characteristics (nitrogen and ash content) of the originating storage. These mass flows are scaled down with rate functions depending on temperature and soil water content, if conditions are sub-optimal for the decomposition. The first three coefficients (k_1, k_2, k_3) for the original ROMUL model were obtained by statistical methods from laboratory experiments with different litter cohorts. The corresponding k coefficients for the nitrogen flux are identical to those of the SOM flux, hence the ratio of the SOM flux to N flux between the compartments are constant. The transformation of organic matter from $F_{i,j}$ compartments into the H_C compartment, i.e. mass flows $C_{4,j}$ and $C_{5,j}$, is based on the metabolism of the decomposing organisms (*Bacteria*, *Arthtropoda* and *Lumbricidae*), and a C/N ratio used for these groups is obtained from soil biology experiments. Nitrogen has a special role in this phase: first, the rate of nitrogen moving from compartments $F_{N,i}$ into compartment H_N is calculated, then a corresponding amount of organic matter, typical for the type of decomposers and depending on the C/N ratio of produced humus (24.0 for *Bacteria* and *Arthtropoda* and 12.8 for *Lumbricidae*), moves from the compartments $F_{C,i}$ to compartment H_C.

The immobilised humus pool, H_C, is also decomposing, at a rather slow rate, modified by the soil temperature and moisture conditions. The value of the decomposition rate, k_6, has a range of an annual minimum of 1–2% up to 15%, depending on the soil texture and the clay content. A maximum rate of H decomposition may be observed for arable soils. On the other hand, as roughly half of the SOM in the boreal zone is in compartment H, the value of rate factor k_6 has a significant effect on the total storage of organic matter in the soil.

References

Chertov, O.G., Komarov, A.S., 1995. Dynamic modelling of Scots pine, Norway spruce and Silver birch ecosystems in European boreal forests. EFI Research Report, Project 519, 152 pp.

Chertov, O.G., Komarov, A.S., Nadporozhskaya, M., Bykhovets, S.S., Zudin, S.L., 2001. ROMUL – a model of forest soil organic matter dynamics as a substantial tool for forest ecosystem modeling. Ecological Modelling 138, 289–308.

Chertov, O.G., Bykhovets, S.S., Nadporozhkaya, M.A., Komarov, A.S., Larionova, A.A., 2007. Evaluation of the rates of transformation of soil organic matter in the ROMUL model. In: Kudeyarov, V.N. (Ed.), Modelling of Organic Matter Dynamics in Forest Ecosystems. Nauka Publishers, Moscow, pp. 83–99 (in Russian).

Cotrufo, M.F., Del Galdo, I., Piermatteo, D., 2009. Litter decomposition: concepts, methods and future perspectives. In: Kutsch, W.L., Bahn, M., Heinemeyer, A. (Eds.), Soil Carbon Dynamics An Integrated Methodology. Cambridge University Press, pp. 76–90 (Chapter 5).

Ducoudé, N., Laval, K., Perrier, A., 1993. SECHIBA, a new set of parametrizations of the hydrologic exchanges at the land/atmosphere interface within the LMD atmospheric general circulation model. Journal of Climate 2 (9), 248–273.

Grayson, R.B., Western, A.W., 1998. Towards areal estimation of soil water content from point measurements: time and space stability of mean response. Journal of Hydrology 207, 68–82.

Helmisaari, H.S., Hanssen, K.H., Jacobson, S., Kukkola, M., Luiro, J., Saarsalmi, A., Tamminen, P., Tveite, B., 2011. Logging residue removal after thinning in Nordic boreal forests: long-term impact on tree growth. Forest Ecology and Management 261, 1919–1927.

Hickler, T., Smith, B., Prentice, C.I., Mjöfors, K., Miller, P., Arneth, A., Sykes, M.T., 2008. CO_2 fertilization in temperate FACE experiments not representative of boreal and tropical forests. Global Change Biology 14, 1531–1542.

Högberg, P., Read, D.J., 2006. Towards a more plant physiological perspective on soil ecology. Trends in Ecology & Evolution 21 (10), 548–554.

Ilvesniemi, H., Liu, C., 2001. Biomass distribution in a young Scots pine stand. Boreal Environment Research 6, 3–8.

Ilvesniemi, H., Pumpanen, J., Duursma, R., Hari, P., Keronen, P., Kolari, P., Kulmala, M., Mammarella, I., Nikinmaa, E., Rannik, Ü., Pohja, T., Siivola, E., Vesala, T., 2010. Water balance of a boreal Scots pine forest. Boreal Environment Research 15, 375–396.

Jansson, P.E., Moon, D.S., 2001. A coupled model of water, heat and mass transfer using object orientation to improve flexibility and functionality. Environmental Modelling & Software 16, 37–46.

Kellomäki, S., Maajärvi, M., Strandman, H., Kilpeläinen, A., Peltola, H., 2010. Model computations on the climate change effects on snow cover, soil moisture and soil frost in the boreal conditions over Finland. Silva Fennica 44 (2), 213–233.

Liski, J., Palosuo, T., Peltoniemi, M., Sievänen, R., 2005. Carbon and decomposition model Yasso for forest soils. Ecological Modelling 189, 168–182.

Mäkipää, R., Heikkinen, J., 2003. Large-scale changes in abundance of terricolous bryophytes and macrolichens in Finland. Journal of Vegetable Science 14, 497–508.

Mäkipää, R., Linkosalo, T., Niinimäki, S., Komarov, A.S., Bykhovets, S.S., Tahvonen, O., Mäkelä, A., 2011. How forest management and climate change affect the carbon sequestration of a Norway spruce stand. Journal of Forest Planning 16, 107–120.

Mammarella, I., Launiainen, S., Grönholm, T., Keronen, P., Pumpanen, J., Rannik, Ü., Vesala, T., 2009. Relative humidity effect on the high frequency attenuation of water vapour flux measured by a closed-path eddy covariance system. Journal of Atmospheric and Oceanic Technology 26 (9), 1856–1866.

Manabe, S., 1969. Climate and the ocean circulation. I. The atmospheric circulation and the hydrology of the earth's surface. Monthly Weather Review 97, 739–774.

Monteith, J.L., 1965. Evaporation and environment. Symposia of the Society for Experimental Biology 19, 205–224.

Naranjo, J.A.N., Weiler, M., Stahl, K., 2011. Sensitivity of a data-driven soil water balance model to estimate summer evapotranspiration along a forest chronosequence. Hydrology and Earth System Sciences 15, 3461–3473.

Palosuo, T., Foereid, B., Svensson, M., Shurpali, N., Lehtonen, A., Herbst, M., Linkosalo, T., Ortiz, C., Todorovic, G.R., Marcinkonis, S., Li, C., Jandl, R., 2012. A multi-model comparison of soil carbon assessment of a coniferous forest stand. Environmental Modelling & Software 35, 38–49.

Pumpanen, J., Ilvesniemi, H., Hari, P., 2003. A process-based model for predicting soil carbon dioxide efflux and concentration. Soil Science Society of America Journal 67, 402–413.

Pumpanen, J., Ilvesniemi, H., Kulmala, L., Siivola, E., Laakso, H., Kolari, P., Helenelund, C., Laakso, M., Uusimaa, M., Hari, P., 2008. Respiration in boreal forest soil as determined from carbon dioxide concentration profile. Soil Science Society of America Journal 72, 1187–1196.

Rantakari, M., Lehtonen, A., Linkosalo, T., Tuomi, M., Tamminen, P., Heikkinen, J., Liski, J., Mäkipää, R., Ilvesniemi, H., Sievänen, R., 2012. The Yasso07 soil carbon model – testing against repeated soil carbon inventory. Forest Ecology and Management 286, 137–147.

Saxton, K.E., Rawls, W.J., 2006. Soil water characteristics estimated by texture and organic matter for hydrologic solutions. Soil Science Society of America Journal 70, 1569–1578.

Schwärzel, K., Menzer, A., Clausnitzer, F., Spank, U., Häntzschel, J., Grünwald, T., Köstner, B., Bernhofer, C., Feger, K.H., 2009. Soil water content measurements deliver reliable estimates of water fluxes: a comparative study in a beech and a spruce stand in the Tharandt forest (Saxony, Germany). Agricultural and Forest Meteorology 149, 1994–2006.

Scott, R.L., 2010. Using watershed water balance to evaluate the accuracy of eddy covariance evaporation measurements for three semiarid ecosystems. Agricultural and Forest Meteorology 150, 219–225.

Skopp, J., Jawson, M.D., Doran, J.W., 1990. Steady-state aerobic microbial activity as a function of soil water content. Soil Science Society of America Journal 54, 1619–1625.
Tuomi, M., Laiho, R., Repo, A., Liski, J., 2011. Wood decomposition model for boreal forests. Ecological Modelling 222, 709–718.
Vesala, T., Suni, T., Rannik, Ü., Keronen, P., Markkanen, T., Sevanto, S., Grönholm, T., Smolander, S., Kulmala, M., Ilvesniemi, H., Ojansuu, R., Uotila, A., Levula, J., Mäkelä, A., Pumpanen, J., Kolari, P., Kulmala, L., Altimir, N., Berninger, F., Nikinmaa, E., Hari, P., 2005. Effect of thinning on surface fluxes in a boreal forest. Global Biogeochemical Cycles 19, GB2001, http://dx.doi.org/10.1029/2004GB002316.