Author comment to Editor

Manuscript no. hess-2014-268, "Estimation of temporal and spatial variations in groundwater recharge in unconfined sand aquifers using Scots pine inventories" by Ala-aho, P. et al.

Dear Editor Dr. Stumpp,

Thank you for addressing the reviewer comments and providing your own valuable comments and suggestions to improve the manuscript. All the comments provided are considered and addressed and in our opinion organization and focus of the manuscript was greatly improved. Modifications in the manuscript can be examined from

1) point-by-point response of editor and reviewer comments (line numbers indicate lines in the FINAL manuscript without annotations) and
2) the "track changes" version of the manuscript which includes all the changes done, with comments upon whose request the change is made.

Both of the above are included in this file.

Sincerely,

Pertti Ala-aho

<u>Detailed response to Editor comments:</u>

Two reviewers thoroughly evaluated your manuscript. Main points were raised that concern:

(i) the length of the manuscript,

Length of the manuscript has been reduced significantly, main text (from intro to conclusions) has been reduced from ~9800 to ~8200 words. This has been done by removing two figures and related discussion, moving equation for ET processes to appendix and improving the focus of the paper by rearranging the text to avoid repetition.

(ii) consideration of water table in calculations,

Influence of water table on ET and thereby groundwater recharge for areas with unsaturated thickness < 1m included with a conceptual approach and explained (L395-410). Exclusion of water table for areas with unsaturated zone thickness > 1m is better justified in the discussion (L619-628).

(iii) terminology about soil or about evaporation, transpiration and evapotranspiration, and

Different evaporation processes are explicitly defined in the text (L367-370) and in caption of Fig. 5. Use of terminology is made coherent throughout the manuscript.

(iv) missing information about vertical depth profiles.

Vertical discretization of simulation profile is better described (L281-285). Depth of vertical profiles in the simulation domain correspond to the estimated unsaturated zone thickness (UZD) in Fig. 4.

You answered in detailed to the reviewers' comments. Generally, I agree to all answers and additionally have some important points I want to emphasize:
1) I agree to one of the reviewers that the specific objectives should be given in more detail.

Objectives are more concisely presented the introduction last paragraph (L102-108)

2) Give ranges of lichen layers that were used for the calculations.

Information for lichen parametrisation is given (L186-191) and limitations discussed (L586-596)

3) Don't forget the comments about the Figures that were not answered in the pdf-file.

Comments are responded to in the annotated PDF file

4) I agree that some of the basic equations in the M&M section can be given in an appendix as Supplement Information.

Equations for evapotranspiration processes are given as Annex 1

5) More details about the vertical profiles are required: It is mentioned in the text that the unsaturated zone varies between 1 and 15m. Is it really all soil? I reckon you only have mineralized soil horizons in the upper meter(s) followed by sediments or (weathered)

rocks/geological material. Are these deeper parts of the unsaturated zone porous sediments? If fractured, the model is not suitable actually. Besides improving chapter 2.1.3, please also change its header as indicated by one of the reviewers.

The conceptual geological model is better explained by adding a cross-section (Fig. 2) and explained in writing (L112-119). Use of the term 'soil' is kept, but explicitly defined (L113).

Author comment to Reviewer #1

The paper seems excessively long. I recommend reducing the text, such as in the Discussion section.

We appreciate this comment to improve the readability of the paper and will shorten the manuscript in following ways:

- Results of the water flow at different depths (Fig. 7) and related discussion will be removed from the manuscript. We reconsidered that this result is not essential for the paper and can be removed in order improve the focus of the paper.
- The comparison of measured stream baseflow to different simulated recharge will be simplified (Table 3).
- Materials and methods will be shortened by removing example of the spatial distribution of model results (page 12 lines 11-23) and not explaining the technicalities (page 14, 20-26).
- In materials and methods section, equations for different evaporation components could be presented as additional material / annex if this in line with the journal formatting.
- some sections of the discussion will be removed (e.g. page 28 lines 2-6 and lines 10-16) or reorganized.

After revision:

- *The manuscript was shortened with the above mentioned ways*
- *Some additional parts were eliminated (e.g. Fig. 3 in the old manuscript and related discussion in the text).*
- *Manuscript was also re-arranged to improve readability and avoid repetition.*
- *All changes and eliminations are presented in the annotated version of the manuscript.*

Throughout the paper, please change the word "depth" to "thickness" in reference to the thickness of the unsaturated zone. The unsaturated zone is the region between land surface and the water table and thus is not a "depth".

This is a good specification, and will be addressed throughout the manuscript

After revision:

- *done*

Page 18: Not simulating the water table "for computational efficiency" is not a valid justification in my opinion. I recommend that the water table be included in the model to accurately simulate hydrologic processes such as ET.

An important comment to ensure models ability to produce realistic ET rates. Presence of water table is acknowledged in the simulations indirectly for cells where the interpolated water table is less than one meter from the ground surface. This is done with the water balance approach described in the paper (page 18). When the simulations were performed with water table fixed at 1m, the annual average ET rates were 5,4%, 2,3 % and 6,5 % higher for LAI values of 0.5, 1.5 and 3 than without the water table, respectively. For deeper water table configuration (2m) the increase in ET was trivial for LAI values of 0.5 and 1.5, and 3,5 % higher for LAI values of 3. We assume that for deeper water table configuration the water table influence on ET would be insignificant.

Therefore we assume that neglecting the water table influence below depth of 1m can produce minor overestimation in areas where the water table is in the region on 1-2 m from ground surface with high LAI values. However we argue that in aquifer scale the impacts will be minimal, because 8% of model surface is within this groundwater table configuration (Fig. 2), and model cells with high LAI are not very common (Fig. 2). This justification will be more clearly incorporated in the manuscript and the text on page 18 better organized to convey the point.

After revision:

- *approach to include water table for areas with unsaturated thickness < 1m is explained (L395-410)*
- *exclusion of water table when unsaturated zone thickness > 1m is justified in the discussion (L619-628)*

Page 20, last paragraph: I don't agree that the land surface is a reasonable representation of the water table "in the transition zone between recharge and discharge areas". Please modify accordingly.

This concept of groundwater table being close to land surface in the recharge-discharge area transition zone is obtained from the work of Rossi (2014). This is better explained with a cross-section which will replace Fig. 3. The cross-section shows the water table sloping towards the discharge zone which demonstrates the assumption of GW-table near ground surface. We assume similar water table configuration around the aquifer.

After revision:

- *A cross-section added (Fig. 2) which demonstrated the assumption of water table near the ground surface in the transition zone between recharge and discharge areas*
- *The issue is also explicitly explained in the text (L231-240)*

Page 23: In comparing the model recharge estimates to that from the baseflow method I recommend that the authors acknowledge that streamflow estimates are (at best) accurate to within 5% based on USGS data. Modify the text accordingly in relation to this qualifier.

A valid notification, and the uncertainty related to baseflow determination will be included in the revised manuscript.

After revision:

- *uncertainty in baseflow is acknowledged and a reference added (L480-482)*

References:

Rossi, P.M.: Integrated management of groundwater and dependent ecosystems in a Finnish esker. PhD thesis, University of Oulu, Finland. Available: http://jultika.oulu.fi/Record/isbn978-952-62-0478-9. 2014.

<u>Detailed response to reviewer #2 comments:</u>

Moreover, I agree with Reviewer1 that the paper should be reduced in the text (the M&M section is rather long)

We appreciate this comment to improve the readability of the paper and will shorten the manuscript in following ways:

- Results of the water flow at different depths (Fig. 7) and related discussion will be removed from the manuscript. We reconsidered that this result is not essential for the paper and can be removed in order improve the focus of the paper.
- The comparison of measured stream baseflow to different simulated recharge will be simplified (Table 3).
- Materials and methods will be shortened by removing example of the spatial distribution of model results (page 12 lines 11-23) and not explaining the technicalities (page 14, 20-26).
- In materials and methods section, equations for different evaporation components could be presented as additional material / annex if this in line with the journal formatting.
- some sections of the discussion will be removed (e.g. page 28 lines 2-6 and lines 10-16) or reorganized.

After revision:

- *The manuscript was shortened with the above mentioned ways*
- *Some additional parts were eliminated (e.g. Fig. 3 in the old manuscript and related discussion in the text).*
- *Manuscript was also re-arranged to improve readability and avoid repetition.*
- *All changes and eliminations are presented in the annotated version of the manuscript.*

There is a confusion about the use of the terms evaporation, transpiration, evapotranspiration. In my opinion evaporation is the process when water leave in gaseous form the bare soil. No plant or crop should be involved in this process. Transpiration is, obviously, the same type of process involving only crop/plant system. The process from the understorey depend if the soil is bare or covered (partially) by vegetation. If the latter applies, it is an evapotranspiration. If everywhere under the forest there are lichens, we can assume this floor as an evaporating surface, assuming no transpiration from the lichens. This is not a semantic question, because through the paper (i.e. in the M&M and Results section) it is not clear at which process the Authors refer.

We appreciate this comment to clarify the different evaporation conceptualizations. The different evaporation and naming conventions are presented in Fig. 4. Their definition will be elaborated in the figure caption.

After revision:

- *different evaporation processes are explicitly defined in the text (L367-370) and in caption of Fig. 5. The use of different terms is made coherent throughout the manuscript.*

2. The soils. This is a problem of the manuscript. It seems to me that the Authors mix soil with the rock/geological material underlaying the soil. The Authors tend to call "soil" all the material between surface and groundwater. This isn’t correct.

The mineral geological material in manuscript is referred to as “soil” throughout the manuscript, which will be explicitly defined in the manuscript to avoid confusion. This naming convention is typically found in the literature. Lichen constituted and organic layer on top of the mineral soil, which is treated as an organic soil type, with specified Brooks and Corey parameter ranges.

After revision

- *the use of the word ‘soil’ for the geological unconsolidated sediment is maintained for simplicity, but use of the term is better defined (L113-114)*

Moreover, just at the end of the discussion they speak about homogeneity of the simulation domain.They do not support this statement with any analysis/observation. So, I was not able to understand the reasons and evidence of homogeneity of the simulation domain. They should better clarify this.

Homogeneity is assumed only in the vertical direction in the soil column for a given model run (page 31, line 21) justification of the assumption and the justification of the assumption is presented (page 31 lines 22-25). Spatially distributed heterogeneity in the model domain is introduced by hydraulic parameters (Section 2.1.3, table 2) varied in the Monte Carlo process.

After revision:

- *assumption of homogeneity in the vertical direction is stated (L296-297) and validity of the assumption is discussed (L609-618)*

And what about the lichens? Till which depth they occur?

Height of the lichen layer based on the samples will be added to section 2.1.2 The height is however mentioned when the model discretization is provided, (page 12 line 1)

After revision:

- *Vertical discretization of simulation profile better described (L281-285), and the height of lichen layer stated (L282)*

Summarizing, the Authors should review the simulation domain, reporting a scheme of it or at least they should clearly report in the text or in a table the different depths of the simulation domain.

This will be improved by replacing Fig. 3 with a more informative cross-section describing the model domain. This will aid the understanding of the model domain, as already explained in section 2.2 and Fig. 1. Different thicknesses of the unsaturated zone are given in detail in Fig. 5.

After revision:

- *Description of the simulation domain is improved by adding a cross-section (Fig. 2) and providing more information about the study site geological conceptual model (L112-119)*
- *Vertical discretization of simulation profile better described (L281-285)*

Estimation of temporal and spatial variations in groundwater recharge in unconfined sand aquifers using Scots pine inventories

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Abstract

Climate change and land use are rapidly changing the amount and temporal distribution of recharge in northern aquifers. This paper presents a novel method for distributing Monte Carlo simulations of 1-D soil profile spatially to estimate transient recharge in an unconfined esker aquifer. The modeling approach uses data-based estimates for the most important parameters controlling the total amount (canopy cover) and timing (~~depth~~ thickness of the unsaturated zone) of groundwater recharge. Scots pine canopy was parameterized to leaf area index (LAI) using forestry inventory data. Uncertainty in the parameters controlling soil hydraulic properties and evapotranspiration was carried over from the Monte Carlo runs to the final recharge estimates. Different mechanisms for lake, soil, and snow evaporation and transpiration were used in the model set-up. Finally, the model output was validated with independent recharge estimates using the water table fluctuation method and baseflow estimation. The results indicated that LAI is important in controlling total recharge amount~~, and the modeling approach successfully reduced model uncertainty by allocating the LAI parameter spatially in the model~~. Soil evaporation compensated for transpiration for areas with low LAI values, which may be significant in optimal management of forestry and recharge. Different forest management scenarios tested with the model showed differences in annual recharge of up to 100 mm. The uncertainty in recharge estimates arising from the simulation parameters was lower than the interannual variation caused by climate conditions. It proved important to take unsaturated ~~depth~~ thickness and vegetation cover into account when estimating spatially and temporally distributed recharge in sandy unconfined aquifers.

Commented [PA1]: Rev#1 com, depth changed to thickness thoughout the manuscript, when referring to the unsaturated zone thickness

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1 Introduction

Eskers are permeable, unconfined sand and gravel aquifers (Banerjee, 1975). In addition to water supply, they support groundwater-dependent ecosystems and provide recreational services (Kløve et al., 2011). Esker hydrology is important as eskers and other glaciofluvial aquifer types cover large areas of the North and are among the dominant aquifer types in the boreal zone. Management of these complex aquifers has gained recent attention (Bolduc et al., 2005, Karjalainen et al., 2013, Koundouri et al., 2012, Kurki et al., 2013). The European Groundwater Directive requires such systems to be characterized in order to determine their quality status, so knowledge of how to estimate groundwater recharge in esker aquifers is becoming increasingly important (EC, 2006). Esker aquifers are commonly covered with managed pine forests, where the forest canopy is likely to influence recharge amounts. The soil surface profile of eskers is complex and highly variable, consisting of kettle holes and sand dunes, resulting in variable ~~depth~~ thickness of the unsaturated zone (Aartolahti, 1973), a factor which also needs to be accounted for in recharge estimation.

Computational methods to estimate groundwater recharge vary from simple water balance models, where water stores and fluxes are represented conceptually and related with adjustable parameters (Jyrkama et al., 2002), to physically-based models using the Richards equation (Assefa and Woodbury, 2013, Okkonen and Kløve, 2011) to solve water fluxes through unsaturated zone. Computational methods solving the Richards equation are often limited to small-scale areal simulations (Scanlon et al., 2002a) and shallow unsaturated zones, and they commonly lack the soil freeze, thaw, and snow storage sub-routines relevant at higher northerly latitudes (Okkonen, 2011). However, computational approaches can be employed to produce the values on spatial and temporal variability in recharge often needed in groundwater modeling (Dripps and Bradbury, 2010). The methods ~~developed so far~~ commonly rely on a GIS platform for spatial representation and calculation approaches based on water balance to create the temporal dimension of recharge (Croteau et al., 2010, Dripps and Bradbury, 2007, Jyrkama et al., 2002, Sophocleous, 2000, Westenbroeck et al., 2010). Neglecting variations in ~~depth~~ thickness of the unsaturated zone is common practice in many water balance models used in recharge estimations. However, the residence time in the unsaturated zone may play an important role, especially in the timing of recharge in deep unsaturated zones (Hunt et al., 2008), as acknowledged in recent work (Assefa and Woodbury, 2013, Jyrkama and Sykes, 2007, Scibek and Allen, 2006, Smerdon et al., 2008).

In numerical recharge models, actual evapotranspiration (ET) is a difficult variable to estimate accurately from climate, soil, and land use data. The vegetation is commonly parameterized from land use or land cover maps (Assefa and Woodbury, 2013, Jyrkama et al., 2002, Jyrkama and Sykes, 2007, Keese et al., 2005), where the vegetation characteristics and leaf area index (LAI) are estimated based solely on vegetation type. In addition to tree canopy transpiration, ~~understorey~~ soil evaporation, i.e. evaporation from the pores of soil matrix, can constitute a large proportion of total ET. Soil evaporation from the forest floor is generally reported to range from 3 to 40% of total ET (Kelliher et al., 1993), although values as high as 92% have been recorded (Kelliher et al., 1998). For conifer forest canopies, ~~understorey~~ soil evaporation can largely compensate for low transpiration in areas with lower LAI (Ohta et al., 2001, Vesala et al., 2005). Data on canopy-scale evaporation rates at latitudes above 60°N are rare (Kelliher et al., 1993). A few studies have estimated ET from pine tree stands at patch scale (Kelliher et al., 1998, Lindroth, 1985), but none has extended this analysis to spatially distributed groundwater recharge. Forest management practices have the potential to affect the transpiration characteristics of coniferous forests, which typically leads to increased groundwater recharge (Bent, 2001, Lagergren et al., 2008, Rothacher, 1970).

Commented [PA3]: Rev#2 general comment; soil evaporation defined here, and explicit definitions used in thie study follow in section 2.2.2

The overall aim of the study was to provide novel information on groundwater recharge rates and factors contributing to the amount, timing, and uncertainty of groundwater recharge in unconfined sandy eskers aquifers. ~~This study sought to~~ Study expands the application of physically-based 1-D unsaturated water flow modeling ~~to simulate spatial and temporal variations in~~for groundwater recharge, while taking into account detailed information on vegetation (pine, lichen), unsaturated ~~soil depth~~layer thickness, cold climate, and simulation parameter uncertainty. ~~CoupModel (Jansson and Karlberg, 2004) was used in simulations because of its ability to represent the full soil-plant-atmosphere continuum adequately and to include snow processes in the simulations (Okkonen and Kløve, 2011). The modeling set-up developed here uses spatially detailed information on tree canopy properties and concentrates on simulating different components of evapotranspiration.~~ Furthermore, ~~it~~ this study considers the effect that forestry land use has on vegetation parameters and how this is reflected in groundwater recharge. ~~The simulation approach takes into account the variability in the unsaturated depth throughout the model domain. Parameter uncertainty, often neglected in recharge simulations, is considered by using multiple random Monte Carlo simulation runs in the process of distributing the 1-D simulations spatially.~~

Commented [PA4]: Rev#2com; overall aim moved up front.

Commented [PA5]: moved to section 2.2

Commented [PA6]: EdCom and Rev#2com; this chapter is reduces to better bring out the aims of the study

~~The overall aim of the study was to provide novel information on groundwater recharge rates and factors contributing to the amount, timing, and uncertainty of groundwater recharge in unconfined sandy eskers aquifers.~~

2 Materials and Methods

2.1 Study site

Groundwater recharge was estimated for the case of the Rokua esker aquifer in northern Finland (Fig. 1). Rokua is an unconfined aquifer consisting of unconsolidated sandy sediments (from here on referred to as soil) underlain by chrystalline bedrock (Fig. 2). Aquifer was formed during previous deglaciation when rivers under the melting ice sheet deposited sandy sediments in the river bed (Aartolahti 1973). The Rokua esker has a rolling surface topography in the aquifer recharge area rising about 60 m above the flat peatland areas surrounding the esker. In the groundwater discharge areas, the aquifer is locally confined by peat soil with low hydraulic conductivity (Rossi et al. 2012).

-The climate at the Rokua aquifer is characterized by precipitation exceeding evapotranspiration on an annual basis and statistics of the annual climate for the study period 1961 - 2010 in terms of precipitation, air temperature and FAO reference evapotranspiration according to Allen et al. (1998) is presented in Table 1. Another important feature of the climate is annually recurring winter periods when most precipitation is accumulated as snow.

~~Groundwater recharge was estimated for a model domain of 82.3 km^2, 3.6 % of which is covered by lakes.~~

2.1.1 ~~Quantifying l~~Leaf area index from forestry inventories

Forestry inventory data from the Finnish Forest Administration (Metsähallitus, MH) and Finnish Forest Centre (Metsäkeskus, MK) were used to estimate LAI for the Rokua esker groundwater recharge area. The available data consisted of 2786 individual plots covering an area of 52.4 km^2 (62.4% of the model domain). The forestry inventories, performed mainly during 2000-2011, showed that Scots pine (*Pinus sylvestris*) is the dominant tree in the model area (94.2% of plots). The forest inventory data include a number of data attributes and the following data fields, included in both the MH and MK datasets, were used in the analysis:

- Plot area (p_A); [ha]
- Main canopy type

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Commented [PA7]: Ed Com, Rev#2 com; specified that the concept of soil is used to refer to the unconsolidated sandy sediments.

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Commented [PA10]: Rev#2 spec com

- Average tree stand height (h); [m]
- Average stand diameter at breast height (d_{bh}); [cm]
- Number of stems (n_{stm}); [1 ha^{-1}]
- Stand base area (b_A); [m^2 ha^{-1}]
- Stand total volume (V); [m^3]

Inventory plots were excluded from the analysis if: (1) main canopy type was not pine forest, (2) data were missing for d_{bh} and h or n_{stm}, or (3) the MH and MK datasets overlapped, in which case MH was retained. However, several plots in the MH dataset were lacking n_{stm} data, which would have created a large gap in data coverage. Therefore the n_{stm} variable was estimated with a log-transformed regression equation using data on d_{bh} , p_A, and V as independent variables. This regression equation was built from 280 plots (R^2 = 0.88) and used to estimate n_{stm} for 288 plots. LAI was estimated as described by Koivusalo et al. (2008). Needle mass for an average tree in stand/plot was estimated from h and d_{bh} using empirical equations presented by Repola et al. (2007). LAI for a stand was calculated as:

$$LAI = N_t * n_{stm} * S_{LA} \quad (1)$$

where N_t = needle mass per average tree in stand [kg], n_{stm} = number of stems per hectare [1 ha^{-1}], and S_{LA} = specific leaf area = 4.43 m^2 kg^{-1} = $4.43*10^{-4}$ ha kg^{-1} (Xiao et al., 2006).

Detailed information on LAI was used to obtain an estimate of how different ~~land use~~ forest management options, already actively in operation in the area, could potentially affect groundwater recharge. Three scenarios were simulated testing the potential impact of forestry operations on groundwater recharge:

1) The first “baseline” scenario simulated the current situation by using LAI pattern at the site (Fig. 3) estimated with Eq. (1). ~~Clear-cutting is an intensive land use form in which the entire tree stand is removed, and it is carried out in some parts of the study area.~~

2) The ~~first~~second scenario simulated the impact of intensive forestry operations as clear-cutting of the tree stand.~~ by not resorting to the estimated LAI pattern at the site (Fig. 2), but~~ Clear-cutting is an intensive land use form where almost the entire tree stand is

removed, and it is carried out in some parts of the study area. ~~by using an~~Low LAI values of 0-0.2 for the whole study site ~~for the whole simulated area~~were used in simulating this scenario.

~~1)~~3) The third scenario~~, which was the opposite of clear-cutting~~ simulated the impact of no forestry operations, i.e. absence of forestry cuttings. ~~the~~The hypothetical mature stand covering the study site was assumed to have high LAI values of 3.2-3.5 found at the study site and reported in the literature (Koivusalo et al., 2008, Rautiainen et al., 2012, Vincke and Thiry, 2008, Wang et al., 2004).

Commented [PA11]: Rev#2 spec com; paragraph reformulated to better explain the different scenarios

2.1.2 ~~Determination of l~~Lichen water retention ~~in soil evaporation~~

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An organic lichen layer covers much of the sandy soil at the Rokua study site (Kumpula et al., 2000), so this lichen layer was ~~included~~ introduced in soil evaporation (SE) calculations. ~~Lichen vegetation has the potential to affect SE by influencing the evaporation resistance of soil and by intercepting rainfall before it enters the mineral soil surface (Kelliher et al., 1998).~~ Although lichens do not transpire water, their structural properties allow water storage in the lichen matrix and capillary water uptake from the soil (Blum, 1973, Larson, 1979). ~~The lichen layer also increases soil surface roughness and thereby retards surface runoff (Rodríguez-Caballero et al., 2012).~~

Commented [PA13]: included in the discussion

In this study, water interception storage by the lichen layer was estimated from lichen samples. In total, six samples (species *Cladonia stellaris* and *C. rangiferina*) were taken in May 2011 from two locations 500 m apart, close to borehole MEA506 (see Fig. 1). These samples were collected by pressing plastic cylinders (diameter 10.6 cm) through the lichen layer and extracting intact cores, after which mineral soil was carefully removed from the base of the sample. Thus the final sample consisted of a lichen layer on top and a layer of organic litter and decomposed lichen at the bottom, and was sealed in a plastic bag for transportation. To obtain estimates of water retention capacity, the samples were first wetted until saturation with a sprinkler, left overnight at +4 °C to allow gravitational drainage and weighed to determine 'field capacity'. The samples were then allowed to dry at room temperature and weighed daily until stable final weight ('dry weight') was reached. The water retention capacity (w_r) of the sample was calculated as:

$$w_r = \frac{m_{fc} - m_{dry}}{\rho_w} \cdot \frac{1}{\pi \cdot r^2} \qquad (2)$$

where m_{fc} is the field capacity weight [M], m_{dry} is the final dry weight [M] at room temperature, ρ_w [M L^{-3}] is the density of water, and r [L] is the radius of the sampling cylinder.

The mean water retention capacity of the lichen samples was found to be 9.85 mm (standard deviation (SD) 2.71 mm) and approximations for these values were used in model parameterization (Table 2). ~~In the simulations, the lichen layer was represented as an organic soil layer with similar Brooks and Corey parameterization as for mineral soil.~~ To acknowledge the lack of information about ~~Brooks and Corey~~(B&C) parameter estimates for lichen, the parameters were included in the simulations Monte Carlo runs (see section 2.2.1) with ranges which in our opinion produced reasonable shape of the pressure-saturation curve allowing easy drainage of the lichen.

2.1.3 ~~Geological data from soil samples~~Soil hydraulic properties

~~Particle size distribution~~Soil texture was determined by sieving (ISO 3310-1 standard sieve, US sieve numbers 5, 10, 18, 35, 60, 120, and 230) ~~from~~ 26 soil samples taken from five boreholes at various depths (Fig. 1). 14 of the samples were analyzed also for pressure saturation curves. Samples were characterized as fine or medium sand, while soil ~~type~~ texture in the other boreholes (Fig. 1) had previously been characterized as medium, fine or silty sand throughout the model domain by the Finnish Environmental Administration as expert *in-situ* analysis during borehole drilling. Therefore the soil samples from the five boreholes were considered to be representative of the soil type in the area. Pressure saturation data from the samples was then used to define parameter ranges for the Brooks and Corey equation used in the simulations (Table 2). Furthermore, ~~particle size distribution~~texture values were employed to calculate the range of saturated vertical hydraulic conductivity for the samples, using empirical equations by Hazen, Kozeny-Carman, Breyer, Slitcher, and Terzaghi (Odong, 2007). The hydraulic conductivity for a given sample ranged approximately one order of magnitude between the equations. When using the five equations for the 26 samples in total, the calculated values were within $1.99 \cdot 10^{-5}$ — $1.47 \cdot 10^{-3}$ [m s^{-1}] for all but one sample. The obtained range was considered to reasonably represent the hydraulic conductivity variability in the study area and simulations ~~(Table 2)~~.

~~Water table was monitored for model validation purposes (Fig. 1) using pressure-based dataloggers (Solinst Levelogger Gold). A measurement was made at one-hour intervals in five~~

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~~boreholes screened 1-2 m below the water table. The depth of the unsaturated zone at these boreholes varied from 1 to 15 m. The data were used to estimate groundwater recharge with the water table fluctuation method (see section 2.5).~~

Commented [PA17]: Rev#2 spec com; moved to validation and modified

2.1.4 Estimation of unsaturated ~~layer depth~~zone thickness

The thickness of the unsaturated layer at each model cell was estimated by subtracting interpolated water table level from digital elevation model (DEM) topography calculated based on LiDAR data (National Land Survey of Finland, 2012). The water table elevation was estimated with the ordinary Kriging interpolation method from four types of observations (Fig. 4): water table boreholes (n = 19), stages of kettle hole lakes (n = 82), elevation of wetlands located in landscape depressions (n = 36), and land surface elevation at the model domain boundary (n = 229) ~~(Fig. 5)~~.

Commented [PA18]: Rev#2 spec com; number of interpolation points added

Water table borehole observations give the most accurate and reliable estimate of the water table position because they provide direct measurements on the water table. The water table elevation in a given piezometer was estimated here as the average value of the entire measurement history of each piezometer.

Kettle hole lakes in the area are imbedded in the aquifer and thus reflect the level of the regional water table (Ala-aho et al., 2013). The lake stage was extracted as the DEM elevation for a given lake, while for large lakes several interpolation points were scattered around the lake shore to better steer the interpolation locally.

Wetland elevation was used as a proxy for the water table elevation in locations where more certain observations (piezometers, lake levels) were lacking. If a wetland was present in the topographical depression, the water table was considered to lie at the depression bottom, in order to sustain the conditions needed for wetland formation. Wetlands were detected from the base map and the value for water table proxy was assigned from the DEM.

Finally, the land surface elevation was considered to give a reasonable estimate of the water table position in the transition zone between aquifer recharge area (model domain) and groundwater discharge areas covered by peatlands (Fig. 2). The Rokua aquifer is phreatic in the recharge area and Rossi et al. (2012) demonstrated that the peatlands partially confine the aquifer and can create artesian conditions in the discharge area. Even though some local overestimation of the water table may have resulted from the approximation method at the transition zone, it was found to be important to have some points to guide the interpolation at the model domain boundary in order to acknowledge the characteristics of the sloping water table towards the discharge area (Fig. 2). The proxy used for water table was extracted from the DEM to points approximately 250 m apart at the boundary of the model domain.

~~2.1.4~~2.1.5 Climate data ~~to drive simulations~~

Driving climate data for the model were taken from Finnish Meteorological Institute databases for the modeling period 1 Jan 1961-31 Oct 2010. Daily mean temperature [°C] and sum of precipitation [mm] were recorded at Pelso climate station, 6 km south of the study area (Fig. 1). The most representative long-term global radiation data [kJ m^{-2} d^{-1}] for the area were available as interpolated values in a 10 x 10 km grid covering the whole of Finland. The interpolation data point was found to be at approximately the same location as borehole MEA2110 (Fig. 1). Long-term data on wind speed [m s^{-1}] and relative humidity [%] were taken from Oulunsalo and Kajaani airports, located 60 and 40 km from the study site, respectively. The data from the airports were instantaneous observations at three-hour intervals, from which daily mean values were calculated. All the climate variables were recorded at reference height 2 m except for wind speed, which was measured at 10 m height. The wind speed data were therefore recalculated to correspond to 2 m measurement height according Allen et al. (1998) by multiplying daily average wind speed by 0.748. The suitability of long-term climate data for the study site conditions was verified with observations made at a climate station established at the study site in an overlapping time period (Dec 2009-Oct 2010) and the agreement between the measurements was found to be satisfactory.

Data on long-term lake surface water temperature were needed to calculate lake evaporation (see section 2.~~3~~2.3), but were not available directly at the study site. However, surface water temperature was recorded at Lake Oulujärvi by the Finnish environmental administration (2013) 22 km from the study site in the direction of the Kajaani climate station (Fig. 1). The Oulujärvi water temperature was found to be closely correlated (linear correlation coefficient 0.97) with daily lake water temperature recorded at Rokua during summer 2012. Daily lake surface temperature data for Lake Oulujärvi starting from 21 July 1970 were used in lake evaporation modeling. However, the data series had missing values for early spring and some gaps during five years in the observation period. These missing values were estimated with a

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sine function, corresponding to the average annual lake temperature cycle, and a daily time series was established for subsequent calculations.

~~It was essential to include snow accumulation in the simulations in order to represent the major spring recharge event of snowmelt. The snow accumulation routines in CoupModel were used (Jansson and Karlberg, 2004) and s~~Snowmelt was calculated with a degree-day approach model in Jansson and Karlberg (2004). Snow routines were calibrated separately using bi-weekly snow water equivalent (SWE) data from Vaala snowline measurements (Finnish environment administration, 2011) for the period 1960-2010 (Fig. 1). This separately calibrated snow model was used for all subsequent simulations.

2.2 Recharge ~~M~~modeling framework

~~2.2.0~~2.2.1 Water flow simulation in 1-D unsaturated soil profile~~Method to distribute 1-D simulations spatially~~

~~Recharge was estimated by simulating w~~Water flow through an unsaturated one-dimensional (1-D) sandy soil ~~column~~ profile (Fig. 2) ~~with~~ was estimated with the Richards equation using CoupModel (Jansson and Karlberg, 2004). CoupModel ~~(Jansson and Karlberg, 2004)~~ was ~~used in simulations~~selected as the simulation code because of its ability to represent the full soil-plant-atmosphere continuum adequately and to include snow processes in the simulations (Okkonen and Kløve, 2011). The simulated soil profile was vertically discretized into 61 layers with increasing layer thickness deeper in the profile. Layer thickness was 0.1 m for the first 16 layers (until 1.6 m), where the topmost 0.1 m was represented as a lichen layer. Layer thickness was progressively increased by defining 0.2 m thickness for the next 7 layers (between 1.6 and 3 m), 0.5 m for the next 14 layers (between 3 and 10 m), 1 m for the next 7 layers (between 10 and 17 m) and 2 m for last 17 layers ranging from 17 m to the bottom of the profile (51 m).

The time variable boundary condition for water flow at the top of the column was defined by driving climate variables and affected by sub-routines accounting for snow processes with daily time step. The short time step was chosen to fully capture the main recharge input from snowmelt. All water at the top of the domain was assumed to be subjected to infiltration. Deep percolation as gravitational drainage was allowed from soil column base using the unit-gradient

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Commented [PA25]: Ed com, Rev#2 com; description of model discretization is improved and better organized boundary condition (see e.g. Scanlon et al., 2002b). Simulations for the unsaturated 1-D soil profile were made for the period 1970-2010, and before each run 10 years of data (1960-1970) were used to spin up the model.

The simulation of the 1-D soil profile was performed 400 times as Monte Carlo runs to facilitate the propagation of model parameter uncertainty in the final model output. Model was ran each time with different parameter values as specified in Table 2. For each individual simulation homogeneity in the vertical direction in terms of soil hydraulic properties was assumed. The parameters for which values were randomly varied were chosen beforehand by trial and error model runs exploring the sensitivity of parameters with respect to cumulative recharge or evapotranspiration. The parameter ranges were specified from field data when possible; otherwise we resorted to literature estimates or in some cases used ± 50% of the CoupModel default providing a typical parameter for the used equation.

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The sensitivity of the parameters varied in the simulations was tested with Kendall correlation analysis, by testing the correlation between each model parameter and cumulative sums of different evapotranspiration components and soil infiltration for the 400 model runs. Individual simulation with unique parameter values did not produce a groundwater recharge value due to the assembling strategy for recharge; therefore the ET components and soil infiltration were selected as variables for comparison. In addition, correlations were examined as scatter plots to ensure that possible sensitivity not captured by the monotonic correlation coefficient was not overlooked.

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2.2.2 Method to distribute 1-D simulations spatially

Groundwater recharge was estimated for a model domain of 82.3 km^2

(Fig. 1). To distribute the simulations in 1-D soil column spatially, the ~~recharge area~~simulation domain was subdivided into different recharge zones, similarly to e.g. Jyrkämä et al. (2002). ~~As each zone requires a unique simulation, the number of simulation setups rapidly increases, leading to high computational demand and/or laborious manual adjustment of model set-up. In the present study, this was avoided by simulating water flow in a single unsaturated 1-D soil column multiple times with different random parameterizations and distributing the results spatially to model zones. Spatial coupling was done with the ArcGIS software (ESRI, 2011).~~

Zonation in the model was based on two variables: LAI and unsaturated zone ~~depth~~ thickness (~~UZD~~UZT). The calculation of spatially distributed values for LAI and ~~UZD~~ UZT is presented in detail in sections 2.1.1 and 2.1.4. ~~This produced~~Both variables were presented as a grid maps with 20m x 20m cell size with a floating point number assigned to each cell, resulting in a total of 205 708 cells for the model domain. The small model cell size was selected to ensure full exploitation of the forest inventory plots in LAI determination. The spatially distributed data were then divided into 15 classes for LAI and 30 classes for ~~UZD~~ UZT~~(Figs. 2 and 5)~~. The classes are primarily equal intervals, which was convenient in the subsequent data processing, but in addition the frequency distributions of LAI and ~~UZD~~ UZT cell values were used to assign narrower classes for parameter ranges with many values (see histograms in Figs. ~~2~~3 and ~~5~~4). Class interval for LAI was 0.2 units up to a value of 2 (class 1: LAI = 0-0.2, etc.) and 0.3 to the maximum LAI value of 3.5. Class interval for ~~UZD~~ UZT was 1 m to 10 m depth and 2 m to the final depth of 51 m. Finally, the classified LAI and ~~UZD~~ UZT data were combined to a raster map with 20m x 20m cell size, producing 449 different zones with unique combinations of LAI and ~~UZD~~ UZT values. Spatial coupling was done with the ArcGIS software (ESRI, 2011).

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~~Simulations for the unsaturated 1-D soil profile were made for the period 1970-2010, and before each run 10 years of data (1960-1970) were used to spin up the model. The time variable boundary condition for water flow at the top of the column was defined by driving climate variables and affected by sub-routines accounting for snow processes. All water at the top of the domain was assumed to be subjected to infiltration. This model simplification well is justified by the permeable soil type with high infiltration capacity (as noted by Keese et al., 2005). Deep percolation as gravitational drainage was allowed from soil column base using the unit-gradient boundary condition (see e.g. Scanlon et al., 2002b). The column was vertically discretized into 60 layers with increasing layer thickness deeper in the profile: Layer thickness was 0.1 m until 1.6 m (the first layer lichen), 0.2 m between 1.6 and 3 m, 0.5 m between 3 and 10 m, 1 m between 10 and 17 m and 2 m from 17 m to the bottom of the profile (51 m).~~

~~The simulation was performed as 400 Monte Carlo runs to ensure enough model runs would be available for each LAI range. Model was ran each time with different parameter values as specified in Table 2. The parameters for which values were randomly varied were chosen beforehand by trial and error model runs exploring the sensitivity of parameters with respect to cumulative recharge or evapotranspiration. The parameter ranges were specified from field data~~

~~when possible; otherwise we resorted to literature estimates or in some cases used ± 50% of the CoupModel default providing a typical parameter for the used equation.~~

~~Variation in the LAI and UZD parameters were used to allocate the simulations spatially to the study site. To follow the example in Figure 3, a cell with a LAI value of 0.1 was assigned to cell class 1 along with all other cells in the LAI range 0-0.2. In addition to LAI, the model was zoned according to unsaturated zone depth. For each model cell, a value for simulated water flow was extracted from the midpoint of unsaturated soil class corresponding to the cell in question. In the example in Figure 3, a cell with an UZD value of 5.2 m belongs to soil class 6, representing unsaturated depth of 5-6 m. Water flow at 5.5 m depth represents the groundwater recharge time series for the model cell in question. In this way, each of the 400 simulations of the unsaturated soil column provided a water flow time series for each UZD class. When LAI class for the same example cell was considered, there were on average 27 simulation time series (number of total model runs [400] divided by number of LAI cell classes [15]) available for the example cell with UZD 5.2 m and LAI 0.15.~~

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~~After completing the CoupModel simulations for the unsaturated soil column~~Variation in the LAI and UZD parameters were used to allocate the 1-D soil profile simulations spatially to the study site. LAI class in model cell specified a subset of the 400 1-D simulations that were applicable for a given cell. UZT class for each cell (Fig. 2) specified the depth in the simulated 51 m soil profile where the water flux output was extracted. ~~,~~Using this approach each unique recharge zone (a combination of ~~UZD~~ UZT and LAI class) had on average 27 ~~recharge~~ water flow time series (number of total model runs [400] divided by number of LAI cell classes [15]) produced by different random combinations of parameters (Table 2). ~~To~~Equation (3) was used to propagate the variability in the 27 time series into the final areal recharge~~, a recharge value was randomly selected for each time step and each recharge zone from the ensemble of 27 (on average) and multiplied by the number of model cells belonging to the recharge zone in question (Eq. 3). Because the recharge rates were in units of mm/day, the rate was converted to volumetric flux [m^3 d^{-1}] by multiplying it by the cell area (A_c) with appropriate unit transformations. Finally, the volumetric flow rate from all the unique recharge zones was summarized for a given time step and the sum was converted from [m^3 d^{-1}] to [mm d^{-1}] by dividing by the surface area of the total recharge area (A_{tot}).This procedure was carried out for all time steps and then repeated a number of times (here 150 times) to ensure that all of the simulated time series for each recharge zone were represented in the random selection process.~~.

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$$R_{i,j} = \frac{\sum_{l=1}^{449} n(l)*Rs_{i,rand(1:k)}*A_c}{A_{tot}} \tag{3}$$

where $R_{i,j}$ is the final sample of areal recharge [mm day^{-1}], i is the index for simulation time step (= 1:14975), j is the index for sample for a given time step (1:150), l is the index for unique recharge zone, n(l) is the number of cells in a given recharge zone, Rs is the recharge sample [mm/day] for a given recharge zone at time step I, k is the number of time series for a given recharge zone, A_c is the surface area of a model raster cell (=20 m * 20 m = 400 m^2), and A_{tot} is the surface area of the total recharge area.

The resulting R matrix has 150 time series for areal recharge produced by simulations with different parameter realizations. The variability between the time series provides an indication of how much the simulated recharge varies due to different model parameter values. The method allows computationally efficient recharge simulations, because the different recharge zones do not all have to be simulated separately.

The ~~method~~ simulation approach assumes that: (1) over the long-term, the water table remains at a constant level, i.e. the unsaturated ~~depth~~ thickness for each model cells stays the same. Monitoring data from 11 boreholes and seven lakes with more than 5 years of observation history shows level variability of 1 – 1.5 m, with depressions and recoveries of the water table. This variability is within the accuracy of water table estimation by interpolation~~, and therefore we find the assumption of long term equilibrium acceptable for the study site~~. (2) t~~t~~he capillary fringe in the sandy soil is thin enough not to affect the water flow before arriving at the 'imaginary' water table at the center of each soil class. (3) o~~o~~nly vertical flow takes place in the unsaturated soil matrix, a typical assumption in recharge estimation techniques (Dripps and Bradbury, 2010, Jyrkama et al., 2002, Scanlon et al., 2002a)~~(Jyrkama et al., 2002; Scanlon et al., 2002a; Dripps and Bradbury, 2010)(Dripps and Bradbury, 2010, Jyrkama et al., 2002, Scanlon et al., 2002)~~. (4) surface runoff is negligible primarily due to the permeable soil type (as noted by Keese et al., 2005)~~(Keese et al., 2005)~~, and also due to lichen cover inhibiting runoff by increasing surface roughness (Rodríguez-Caballero et al., 2012). The maximum observed daily rainfall for the area has been 57.4 mm. Further assuming that rain for the day fell only during one hour, it would equal to $1.59*10^{-5}$ m s^{-1} input rate of water, which is close to the lower range of saturated hydraulic conductivity at the study site ($1.99*10^{-5}$ m s^{-1}).

Therefore rainstorms at the site very rarely exceed the theoretical infiltration capacity. As a field verification, surface runoff has not been observed during field visits and the area lacks intermittent or ephemeral stream networks. ~~The lichen layer also increases soil surface roughness and thereby retards surface runoff (Rodríguez-Caballero et al., 2012).~~ ~~(5) uncertainties in the estimation of spatially distributed LAI and UZD values justify the use of approximations (i.e. water flow at the UZD class range midpoint and LAI value specified only as a range for each cell) in the cell classification phase.~~

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~~The model set-up used fine temporal and spatial discretization with a daily time step and 20m x 20m cell size, respectively. The short time step was chosen to fully capture the main recharge input from snowmelt and to demonstrate its impact on recharge variability at different water table depths. The small model cell size was selected to ensure full exploitation of the forest inventory plots in LAI determination. Simulation times for the current set-up were approximately 10 hours for 400 simulations of the 50 m soil profile for the period 1961-2010, and 12 hours to redistribute the simulations to the 200 000 model cells for each time step and create 150 realizations of recharge time series. Where the computational capacity or the length of the run times poses a problem, the modeling methodology allows different spatial and temporal dimensions, which would speed up the long simulation times.~~

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~~The sensitivity of the parameters varied in the simulations was tested with Kendall correlation analysis, by testing the correlation between each model parameter and cumulative sums of different evapotranspiration components and soil infiltration for the 400 model runs. Individual simulation with unique parameter values did not produce a groundwater recharge value due to the assembling strategy for recharge; therefore the ET components and soil infiltration were selected as variables for comparison. In addition, correlations were examined as scatter plots to ensure that possible sensitivity not captured by the monotonic correlation coefficient was not overlooked.~~

~~2.3 Estimation of Evapotranspiration~~

2.2.3 Estimation of evapotranspiration

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Four different evaporation processes were considered in this study (Fig. 5); soil evaporation (evaporation from the topmost soil layer, i.e. the lichen matrix), snow evaporation (evaporation from snow surface), transpiration (evaporation through the vascular system of tree canopy) and lake evaporation (evaporation from free water surface) ~~and transpiration (Fig. 4)~~. In areas with

Commented [PA34]: Rev#2 general com; different conceptualization of ET explicitly defined unsaturated soil zones, the first three ~~evaporation~~ components were estimated, along with water flow simulations, using CoupModel. However, as 3.6 % (2.9 km^2) of the surface area of the study site consists of lakes (Fig. 1), lake evaporation from free water surfaces was calculated independently from the CoupModel simulations. ~~Kettle hole lakes in esker aquifers often lack surface water inlets and outlets and are therefore an integral part of the groundwater system (Ala-aho et al., 2013, Winter et al., 1998)(Winter et al., 1998; Ala-aho et al., 2013)(Ala-aho et al., 2013, Winter et al., 1998), so we considered these lakes as contributors to total groundwater recharge. In other words, rainfall per lake surface area is treated equally as addition to the aquifer water storage as groundwater recharge. As a difference, lake water table is subjected to evaporation unlike the groundwater table.~~

~~2.3.1 Transpiration~~

Transpiration from the Scots pine canopy ($L_v E_{tp}$) was calculated using Penman-Monteith (P-M) combination ~~Eq. (4)~~(Appendix 1, Eq. 1)~~:~~.

$$\cancel{L_v E_{tp} = \frac{\Delta R_n + \rho_a \cdot c_p \frac{(e_s - e_a)}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)}} \qquad \cancel{(4)}$$

~~where R_n is net radiation, ρ_a is air density, c_p is the specific heat of air, e_s is the vapor pressure at saturation, e_a is the actual air vapor pressure, r_a is the aerodynamic resistance, Δ is the slope of the saturated vapor pressure-temperature curve, γ is the psychrometer constant, and r_s is surface resistance.~~

~~The aerodynamic resistance (r_a) for transpiration was calculated as:~~

$$\cancel{r_a = \frac{ln\left(\frac{z_{ref} - d}{z_0}\right)}{k^2 \cdot u}} \qquad \cancel{(5)}$$

~~where z_{ref} is the reference height of the measurements, d is the displacement height, z_0 is the roughness length, k is von Karman's constant, and u is wind speed.~~

~~Surface resistance (r_s) was estimated with Eq. 6:~~

$$\cancel{r_s = \frac{1}{max(LAI \cdot g_l; 0.001)}} \qquad \cancel{(6)}$$

~~where g_l is the leaf conductance given by the Lohammar equation (see e.g. Lindroth, 1985).~~

Whenever possible, all the parameters relating to the P~~enman~~-M~~onteith~~ equation were estimated based on data, namely LAI of the canopy. Surface resistance and saturation vapor pressure difference are the main factors controlling conifer forest evapotranspiration, while the aerodynamic resistance is of less importance (Lindroth, 1985, Ohta et al., 2001). In the calculation of aerodynamic resistance with the P-M equation, roughness length is related to LAI and canopy height, according to Shaw and Pereira (1982). Other parameters governing the aerodynamic resistance, except for LAI, were treated as constant. The surface resistance of the pine canopy was estimated with the Lohammar equation (see e.g. Lindroth, 1985), accounting for effects of solar radiation and air moisture deficit in tree canopy gas exchange. Because LAI values have a strong influence in the surface resistance Lohammar equation, the other parameters governing the surface resistance were excluded from the Monte Carlo runs. Distribution of root biomass with respect to depth from the soil surface was presented with an exponential function, because most Scots pine roots are concentrated in the shallow soil zone. A typical root depth value of 1 m was used for the entire canopy (Kalliokoski, 2011, Kelliher et al., 1998, Vincke and Thiry, 2008).

~~2.3.2 Soil evaporation with lichen cover~~

Soil and snow evaporation ~~was~~ were calculated using an empirical approach (Appendix 1, Eq. ~~7~~4) based on the P-M equation, as described in detail in Jansson and Karlberg (2004). ~~In this approach, s~~Soil evaporation ~~(L_vE_{tp})~~ is calculated for the snow-free fraction of the soil surface, and the snow evaporation is solved separately as a part of snow pack water balance.~~:~~

$$\cancel{L_vE_{tp} = \frac{\Delta(R_n - q_h) + \rho_a \cdot c_p \frac{(e_s - e_a)}{r_{as}}}{\Delta + \gamma\left(1 + \frac{r_{ss}}{r_{as}}\right)}} \qquad \cancel{(7)}$$

~~where q_h is the soil surface heat flux, r_{as} is the aerodynamic resistance of soil, and r_{ss} is the surface resistance of soil.~~

~~The aerodynamic resistance of the soil (r_{as}) is calculated as Eq (8):~~

$$\cancel{r_{as} = r_{alai} \cdot LAI + \frac{1}{k^2 \cdot u} \cdot ln\left(\frac{z_{ref} - d}{z_{0M}}\right) \cdot ln\left(\frac{z_{ref} - d}{z_{0H}}\right) \cdot f(R_{ib})} \qquad \cancel{(8)}$$

~~where r_{alai} is an empirical parameter, z_{0M} and z_{0H} are surface roughness lengths for momentum and heat, respectively, and $f(R_{ib})$ is a function governing the influence of atmospheric stability.~~

~~The surface resistance for soil (r_{ss}) is given by:~~

$$\cancel{r_{ss} = \begin{cases} r_\Psi \cdot log(\Psi_s - 1 - \delta_{surf}); & \Psi_s > 100 \\ r_\Psi(1 - \delta_{surf}); & \Psi_s \leq 100 \end{cases}} \qquad \cancel{(9)}$$

~~where r_ψ is an empirical coefficient, Ψ_s is the water tension in the uppermost soil layer, and δ_{surf} is the mass balance at the soil surface (see Jansson and Karlberg, 2004).~~

In areas where the water table is close to the ~~soil~~ ground surface, the water table can provide an additional source of water for evapotranspiration (Smerdon et al., 2008). To take into account the decreased recharge for areas with near surface water tables, the recharge for cells with an unsaturated zone of <1 m (8.3% of the study site, 6.8 km^2) was estimated with a water balance approach. We assumed that for areas with a shallow water table, soil water content was not a limiting factor for transpiration. Therefore an additional water source for transpiration was considered by making the transpiration rate equal to simulated potential transpiration (T) during times when the actual transpiration was simulated (T >0.05 mm). Increasing effect of the water table located at 1 m depth on soil evaporation was tested with simulations and found to be 5-10% higher with than without a water table. Therefore a 7% addition was made to the simulated actual soil evaporation for cells with a shallow water table. Daily recharge (R_{1m}, L T^{-1}) for cells with unsaturated ~~depth~~thickness below 1 m was estimated as:

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$$R_{1m} = I - T_{adj} - ES_{adj} \qquad (\cancel{10}4)$$

where I is infiltration water arriving to lake/soil surface, including both meltwater from the snowpack and precipitation [L T^{-1}] , T_{adj} [mm d^{-1}] is adjusted transpiration, and ES_{adj} [mm d^{-1}] is adjusted soil evaporation.

~~Simulations with a water table fixed at different depths in the soil profile would have been possible in the CoupModel setup. However, it would have doubled the amount of model runs for each considered water table depth and water table was not explicitly simulated for computational efficiency. Upward fluxes were not excluded from the recharge time series and negative fluxes were considered as "negative recharge" at any depth. Only the simplification is made that water available for upward fluxes comes only form the soil moisture storage, not from the water table.~~

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~~To take into account the decreased recharge for areas with near surface water tables, the recharge for cells with an unsaturated zone of <1 m was estimated with a water balance approach. We assumed that for areas with a shallow water table, soil water content was not a limiting factor for transpiration. Therefore an additional water source for transpiration was considered by making the transpiration rate equal to simulated potential transpiration (T) during~~

~~times when the actual transpiration was simulated (T >0.05 mm). Increasing effect of the water table located at 1 m depth on soil evaporation was tested with simulations and found to be 5-10% higher with than without a water table. Therefore a 7% addition was made to the simulated actual soil evaporation for cells with a shallow water table. Daily recharge (R_{1m}, $L\ T^{-1}$) for cells with unsaturated depth below 1 m was estimated as:~~

~~$R_{1m} = I - T_{adj} - ES_{adj}$ (10)~~

~~where I is infiltration water arriving to lake/soil surface, including both meltwater from the snowpack and precipitation [$L\ T^{-1}$] , T_{adj} [$mm\ d^{-1}$] is adjusted transpiration, and ES_{adj} [$mm\ d^{-1}$]is adjusted soil evaporation.~~

Kettle hole lakes in esker aquifers often lack surface water inlets and outlets and are therefore an integral part of the groundwater system (Ala-aho et al., 2013, Winter et al., 1998), so we considered these lakes as contributors to total groundwater recharge. In other words, rainfall per lake surface area is treated equally as addition to the aquifer water storage as groundwater recharge. As a difference, lake water table is subjected to evaporation unlike the groundwater table. ~~Lake evaporation~~

~~Lake cells were identified according from a base map and the daily lake recharge (R_{lake}, [$L\ T^{-1}$]) per unit area was then calculated with a water balance approach as:~~

~~$R_{lake} = I - E_{lake}$ (11)~~

~~where E_{lake} [$L\ T^{-1}$] is lake evaporation.~~

Lake evaporation (E_{lake}) was estimated with the mass transfer approach (see e.g. Dingman, 2008) according to Eq. (7~~12~~) in Appendix 1.

~~$E_{lake} = K_E \cdot v_a \cdot (e_s - e_a)$ (12)~~

~~where K_E is mass transfer coefficient [$ML^{-1}T^{-2}$], v_a is wind speed [$L\ T^{-1}$], e_s [$ML^{-1}T^{-2}$] is saturated vapor pressure at lake water surface temperature, and e_a [$ML^{-1}T^{-2}$] is air vapor pressure. The mass transfer coefficient (K_E) represents the efficiency of vertical water transport from the evaporating surface and it can be treated as a function of lake size:~~

~~$K_E = 1.69 \times 10^{-5} \cdot A_L^{-0.05}$ (13)~~

~~where A_L is lake surface area [km^2]~~

~~The groundwater recharge study area has lakes of variable size, from less than 1 ha to 25 ha (Fig. 1). Lake size variability was included in the total recharge calculation by randomly selecting a~~ ~~K_E~~ ~~value (from the range 1-25 ha) in Eq. (13) when calculating lake evaporation, and thereby groundwater recharge in model cells with lakes (see section 2.2).~~ The mass transfer method was selected because of its simplicity, daily output resolution, low data requirement, and physically-based approach. However various calculation methods could easily be used in the modelling framework, depending on the data availability (see e.g. Rosenberry et al., 2007). If lake percentage in the area of interest is high, more sophisticated methods may be required to better represent the system. ~~However, f~~For the Rokua site the bias introduced by a simplistic approach was considered minor.

~~2.4 Estimation of unsaturated layer depth~~

~~The depth of the unsaturated layer at each model cell was estimated by subtracting interpolated water table level from digital elevation model (DEM) topography calculated based on LiDAR data (National Land Survey of Finland, 2012). The water table elevation was estimated with the ordinary Kriging interpolation method from four types of observations: water table boreholes, stages of kettle hole lakes, elevation of wetlands located in landscape depressions, and land surface elevation at the model domain (Fig. 5).~~

~~Water table borehole observations give the most accurate and reliable estimate of the water table position because they provide direct measurements on the water table. The water table elevation in a given piezometer was estimated here as the average value of the entire measurement history of each piezometer.~~

~~Kettle hole lakes in the area are imbedded in the aquifer and thus reflect the level of the regional water table (Ala-aho et al., 2013). The lake stage was extracted as the DEM elevation for a given lake, while for large lakes several interpolation points were scattered around the lake shore to better steer the interpolation locally.~~

~~Wetland elevation was used as a proxy for the water table elevation in locations where more certain observations (piezometers, lake levels) were lacking. If a wetland was present in the topographical depression, the water table was considered to lie at the depression bottom, in order to sustain the conditions needed for wetland formation. Wetlands were detected from the base map and the value for water table proxy was assigned from the DEM.~~

~~Finally, the land surface elevation was considered to give a reasonable estimate of the water table position in the transition zone between recharge and discharge areas. The Rokua aquifer is phreatic in the recharge area and Rossi et al. (2012) demonstrated that the peatlands partially confine the aquifer and can create artesian conditions in the discharge area. Even though some local overestimation of the water table may have resulted from the approximation method at the transition zone, it was found to be important to have some points to guide the interpolation at the model domain boundary in order to acknowledge the characteristics of the sloping water table towards the discharge area. The proxy used for water table was extracted from the DEM to points approximately 250 m apart at the boundary of the model domain.~~

~~2.5~~2.3 Model validation

Model performance was tested by comparing the simulated recharge values with two independent recharge estimates in local and regional scale; the water table fluctuation (WTF) method and base flow estimation, respectively. The WTF method is routinely used to estimate groundwater recharge because of its simplicity and ease of use. It assumes that any rise in water level in an unconfined aquifer is caused by recharge arriving at the water table. For a detailed description of the method and its limitations, see e.g. Healy and Cook (2002). The recharge amount (R, L T^{-1}) is calculated based on the water level prior to and after the recharge event and the specific yield of the soil:

$$R = S_y \frac{\Delta h}{\Delta t} \qquad (~~14~~5)$$

where S_y is the specific yield, h is the water table height [L], and t is the time of water table rise [T].

The WTF method requires groundwater level data with adequate resolution for both time and water level, to identify periods of rising and falling water table. Water table was monitored ~~for model validation purposes (Fig. 1)~~ using pressure-based dataloggers (Solinst Levelogger Gold)~~. A measurement was made at one hour intervals in five boreholes screened 1-2 m below the water table. The depth of the unsaturated zone at these boreholes varied from 1 to 15 m. The data were used to estimate groundwater recharge with the water table fluctuation method (see section 2.5).~~ ~~Such data, with hourly interval water level~~ recording~~ings~~ ~~were available for the study site~~with hourly interval ~~-~~from six water table wells with average unsaturated zone ~~depths~~ thicknesses of 1.2, 1.6, 5.0, 8.0, 9.3, and 14.7 m (Fig. 1). Wells where the water table was <2 m from the ground surface responded to major precipitation events. In the deeper wells, only

Commented [PA37]: Rev#2 spec om; moved from section 2.1.3 and modified the recharge from snowmelt was seen as water table rise. Estimates of the soil specific yield are required for the calculations (Eq. ~~14~~5), but no soil samples were available from the wells used in water table monitoring. Drilling records for these wells reported fine and medium sand, which was consistent with the particle size distribution for other wells in the area. Therefore an estimated value of 0.20-0.25 for the specific yield of all wells was used, according to typical values for fine and medium sand (Johnson, 1967).

The recharge estimated with the WTF method was compared with the simulated recharge during the recorded water level rise in the well. For each well, the cumulative sum of simulated water flow was extracted from soil profile depth corresponding to well water table depth. As an example, the simulated recharge in well ROK1 (unsaturated ~~depth~~ thickness on average 14.7 m) was extracted from soil class 12, corresponding to recharge for unsaturated thickness of 14-16 m. All 400 model runs were used, providing 400 estimates for recharge for each time period of recorded water level rise.

A regional estimate of groundwater recharge was estimated as baseflow of streams originating at the groundwater discharge area. Because the Rokua esker aquifer acts as a regional water divide, stream flow was monitored around the esker, in total of 18 locations (Fig. 1). The flows were measured total of 8 times between 6 July 2009 and 3 August 2010 (see Rossi et al., 2014). The lowest total outflow during 9-10 February 2010 was recorded after three months of snow cover period, when water contribution to streams from surface runoff was minimal. The minimum outflow was considered as baseflow from the aquifer reflecting long term groundwater recharge in the area.

~~However, some groundwater discharges to larger regional lakes and rivers traveling underneath the measured small streams (Rossi et al., 2014), and thereby the baseflow to the small streams was expected to be lower than the total recharge.~~

Commented [PA38]: removed to avoid contradiction, and explained in the discussion

3 Results

3.1 Model validation with the WTF and baseflow methods

Model validation showed that the modeling approach could reasonably reproduce (1) the main groundwater recharge events when compared to the WTF method (Fig. 6) and (2) the regional level of recharge compared to stream baseflow. ~~The model showed reasonable performance and consistency against independent recharge estimates obtained with both WTF (Fig. 6) and baseflow methods (Fig. 6 and Table 3, respectively).~~ The WTF method agreed well with the simulated values, with overlapping estimates between the methods for all but two ~~boreholes~~recharge events. Also the median value of simulations was close to WTF method, with some bias to higher estimates from the simulations. The discrepancy can be due to very different assumptions behind the methods and uncertainty in local parameterization; in the WTF method for the specific yield (S_y) and for simulations mainly the hydraulic conductivity which dictates the simulated timing of recharge. Uncertainty in the S_y estimate is acknowledged by showing S_y a range rather than a single value (Fig. 6), but still S_y is not truly known for the location of observation boreholes. Simplifying assumptions and subjective interpretation of both timing and height of water table rise create additional inaccuracies in the WTF estimate.

Independent ~~The order of magnitude for rR~~regional estimate of recharge, stream baseflow, ~~corresponded well to simulated recharge, level of match depending on the examined simulation period (Table 3). The measured baseflow~~ was 70 500 m^3 s^{-1}, or 312.7 mm a^{-1} when related to the recharge area. The order of magnitude agreed with long term simulated average of ~~simulations ,~~ 362.8 mm a^{-1}. Typical error in individual stream flow measurements is within 3-6 % of the measured value (Sauer and Meyer, 1992) ~~(XXX)~~, which brings minor uncertainty in the baseflow value. ~~When comparing to the simulated long term average recharge and recharge for previous year, the measured baseflow was lower than the simulated recharge. Then again, when extracting the recharge data for the exact stream discharge measurements dates 9-10 February 2010, stream baseflow exceeded the simulated recharge.~~ ~~Model validation showed that the modeling approach could reasonably reproduce (1) the main groundwater recharge events when compared to the WTF method and (2) the regional level of recharge compared to stream baseflow. The WTF estimates for recharge agreed with the simulations, with a slight tendency for higher estimates by the simulations. The discrepancy can be due to different assumptions behind the methods and uncertainty in local parameterization; in the WTF method for the specific yield and for simulations mainly the hydraulic conductivity which dictates the~~

Commented [PA39]: section modified and some parts of the discussion are moved here to avoid repetition

Commented [PA40]: rev#2 specific comment; uncertainty in the WTF method highlighted

Commented [PA41]: Rev#com; uncertainty in baseflow mentioned

~~timing of recharge. However, there were overlapping estimates for almost every recharge event which shows consistency between the methods.~~ The smaller value for stream baseflow ~~was lower than the~~compared to simulated long term average recharge~~, which was expected because of the site~~ can be explained with conceptual understanding of site hydrogeology (Ala-aho et al., 2015, Ala-aho et al., 2013, Rossi et al., 2012, Rossi et al., 2014)~~(Ala-aho et al., 2013, Rossi et al., 2012, Rossi et al., 2014)~~. Part of the recharged groundwater does not discharge to the small streams whose baseflow was measured, but flows underneath the stream catchments and seeps out to regional surface bodies (Lake Oulujärvi and River Oulujoki) further away from the recharge area (Fig. 2~~3~~). Fully integrated surface-subsurface hydrological modeling study of the same site presented in Ala-aho et al. (2015) simulated an outflow of 79 mm a^{-1} to regional surface water bodies~~Simulation results presented in Ala-aho et al. All of the outflow from the aquifer was likely not captured by the baseflow measurements as some of the water discharges to larger streams and lakes outside of the stream catchments (Rossi et al., 2014). When simulated recharge was extracted specifically for the baseflow measurements dates, the lower values for simulated recharge were also anticipated. The recharge displayed strong seasonal variability (see Fig. 7), but the discharge to streams is in general more stable because of the stabilizing effect of the groundwater storage. In conclusion, the order of magnitudes in the regional baseflow estimate and the simulation results were consistent. Despite the very different assumptions on which the modelling and field based methods were based, all provided similar estimates for groundwater recharge at the study site.~~.

Commented [PA42]: Rev#2 specific com; references added and the conceptual model supported with a reference to fig. 3.

~~3.2~~

~~3.3~~ Temporal variations in groundwater recharge

3.2 ~~Recharge and evapotr~~

~~3.4 anspiration time series~~

~~The dynamics of water flow time series responded to snowmelt and rain storm events rapidly at 1.5 m depth, but because of permeable sandy soils a clear signal of annual snowmelt was evident throughout the depth of the aquifer (Fig. 7). The data showed a delay in response to wet seasons when moving down in the soil profile, as expected. For example, snowmelt in the beginning of May 2008, gave the highest flow rate at 11 m in 19 May 2008, at 23 m in 29 June~~

~~2008 and at 49 m in 5 April 2009. Temporal variability is pronounced higher in the soil profile showing larger variability between maximimum and minimum flow.~~

~~Average land surface ET components remained relatively constant between years, but the simulated ET displayed a wide spread between simulations (Fig. 8). Estimated annual evaporation evapotranspiration from the land surface (mean 237.6 mm) was somewhat lower than previous regional estimates of total ET (300 mm; (Mustonen, 1986)). Lake evaporation rates were generally higher than evapotranspiration from the land surface (420.0 mm). due to the different method for estimating lake evaporation. The variation in simulated lake evaporation was considerably lower than that in ET, as a different approach was used to account for uncertainty in the simulations. Transpiration showed greater variation between simulations than soil evaporation and total land surface ET. On average, transpiration also comprised a slightly larger share of total evaporation than soil evaporation. Simulated snow evaporation was a small, yet not insignificant, component in the total ET from land surface.~~

When recharge simulation time series were summarized to annual values (1 Oct-30 Sept), recharge rates co-varied with annual infiltration with linear correlation coefficient of 0.89 (Fig. ~~9~~7) ~~.~~ as expected based on previous work in humid climate and sandy soils (Keese et al., 2005, Lemmelä, 1990). Both annual recharge and infiltration displayed an increasing trend. The plot also showed the level of uncertainty in annual recharge values introduced by differences in model parameterization (~~see Table 2~~black area). The difference between minimum and maximum value for simulated annual recharge was on average 23.0 mm. Thus the ~~maximum~~ variablity in recharge estimates was 6.3 % of mean annual recharge 362.8 mm.

~~Annual recharge was strongly correlated with annual sum of precipitation (linear correlation coefficient 0.89) as expected based on previous work in humid climate and sandy soils (Keese et al., 2005, Lemmelä, 1990).~~ According to the simulations, the *effective rainfall*, i.e. the percentage of corrected rainfall resulting in groundwater recharge annually, was on average 59.3%. This is in agreement with previous studies on unconfined esker aquifers at northerly latitudes, in which the proportion of annual precipitation percolating to recharge is reported to be 50-70% (71% by Zaitsoff (1984), 54% by Lemmelä and Tattari (1986) and 56% by Lemmelä (1990)). The percentage of effective rainfall varied considerably, by almost 30 %-units, between different hydrological years, from 44.8% in some years up to 73.1% in others.

Commented [PA43]: rev#1 com#1, ed com; removed (with related fig. 7) to improve focus and reduce length reduced

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Commented [PA45]: chapter moved downwards to emphasize the central results

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~~**Tests on whether the interannual variation in effective rainfall percentage could be explained by sum of annual precipitation or maximum snow water equivalent showed no correlation between either of these variables and effective recharge coefficient for a given year.**~~

3.3 Influence of LAI on spatial variation of groundwater recharge

The spatial distribution of groundwater recharge was mostly due to variations in LAI ~~originating from forestry data~~, but also influenced by~~,~~ distance to water table, and distribution of lakes (Fig. ~~10~~8). Higher evaporation rates from lakes led to lower recharge in lakes (see red spots in Fig. 8~~10~~). Similarly, ~~large~~high LAI led to high ET and resulted in low recharge in plots with high LAI. Other areas of low recharge, although not as obvious at the larger spatial scales shown in Fig. ~~10~~8, were cells with a shallow water table (section 2.~~3~~2.2). The effect of high ET at locations with a shallow water table can best be seen in south-east parts of the aquifer.

Kendall correlation analysis of simulation parameters and annual average model outputs identified LAI as the most important parameter controlling evapotranspiration and infiltration (Table 3~~4~~). Parameters related to soil hydraulics and evaporation showed some sensitivity to simulation results, while the parameters for lichen vegetation were only slightly sensitive or insensitive to simulation output variables. The LAI parameter governed the level of evaporation for different ET components (Fig. ~~11~~9). Evaporation from soil (and snow) compensated for mean annual ET for LAI values up to around 1.0, after which total ET increased as a function of LAI.

The scenarios for low (0 … 0.2) and high (3.2 … 3.5) LAI changed the groundwater recharge rates compared to the current LAI distribution (in Fig. ~~9~~7). In the high LAI scenario the annual recharge was on average 101.7 mm lower than in the low LAI scenario. These results suggest that management of the Scots pine canopy has a significant control on the total recharge rates in unconfined esker aquifers.

Average land surface ET components remained relatively constant between years, but the simulated ET displayed a wide spread between simulations (Fig. ~~8~~10). Estimated annual evapotranspiration (mean 237.6 mm) was somewhat lower than previous regional estimates of total ET (300 mm; (Mustonen, 1986)). Lake evaporation rates were generally higher than evapotranspiration from the land surface (420.0 mm). The variation in simulated lake

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Commented [PA48]: Rev#2 spec com; removed, remark not needed evaporation was considerably lower than that in ET, as a different approach was used to account for uncertainty in the simulations. Transpiration showed greater variation between simulations than soil evaporation and total land surface ET. On average, transpiration also comprised a slightly larger share of total evaporation than soil evaporation. Simulated snow evaporation was a small, yet not insignificant, component in the total ET~~ from land surface~~.

~~3.5 Spatial distribution of groundwater recharge~~

~~The spatial distribution of groundwater recharge was mostly due to variations in LAI originating from forestry data, distance to water table, and distribution of lakes (Fig. 10). Higher evaporation rates from lakes led to lower recharge in lakes (see red spots in Fig. 10). Similarly, large LAI led to high ET and resulted in low recharge in plots with high LAI. Other areas of low recharge, although not as obvious at the larger spatial scales shown in Fig. 10, were cells with a shallow water table (section 2.3.2). The effect of high ET at locations with a shallow water table can best be seen in south-east parts of the aquifer.The scenarios for low (0 … 0.2) and high (3.2 … 3.5) LAI would change the groundwater recharge rates compared to the current LAI distribution (Fig. 9). In the high LAI scenario the annual recharge was on average 101.7 mm lower than in the low LAI situation. These results suggest that management of the Scots pine canopy has a significant control on the total recharge rates in unconfined esker aquifers.~~

~~3.6 Influence of simulation parameters on groundwater recharge~~

~~Kendall correlation analysis of simulation parameters and annual average model outputs identified LAI as the most important parameter controlling evapotranspiration and infiltration (Table 4). Parameters related to soil hydraulics and evaporation showed some sensitivity to simulation results, while the parameters for lichen vegetation were only slightly sensitive or insensitive to simulation output variables.~~

~~The LAI parameter governed the level of evaporation for different ET components (Fig. 11). Evaporation from soil (and snow) compensated for mean annual ET for LAI values up to around 1.0, after which total ET increased as a function of LAI.~~

~~The scenarios for low (0 … 0.2) and high (3.2 … 3.5) LAI would change the groundwater recharge rates compared to the current LAI distribution (Fig. 9). In the high LAI scenario the annual recharge was on average 101.7 mm lower than in the low LAI situation. These results suggest that management of the Scots pine canopy has a significant control on the total recharge rates in unconfined esker aquifers.~~

4 Discussion

~~The modeling approach developed here used forestry inventory data to simulate spatial and temporal variations in recharge. The Richards equation-based 1-D simulations were spatially distributed using Monte Carlo runs for an unsaturated soil column. Within the Monte Carlo process, residence time in the unsaturated zone was accounted for, while uncertainty in selected model parameters was propagated to the final recharge time series.~~ The method used here to estimate LAI from forestry inventories introduces a new approach for incorporating large spatial coverage of detailed conifer canopy data into groundwater recharge estimations. LAI values reported for conifer forests in Nordic conditions similar to the study site are in the range 1-3, depending on canopy density and other attributes (Koivusalo et al., 2008, Rautiainen et al., 2012, Vincke and Thiry, 2008, Wang et al., 2004). The LAI values obtained for the study site (mean 1.25) were at the lower end of this range. Furthermore, the data showed a bimodal distribution, with many model cells with low LAI (< 0.4) lowering the mean LAI. The low LAI values were expected because of active logging and clearcutting activities in the study area. Although the equations to estimate LAI are empirical in nature and based on simplified assumptions, the method can outline spatial differences in canopy structure. Wider use of this method in Finland is practically possible, as active forestry operations in Finland have yielded an extensive database on canopy coverage, which could be used in groundwater management. However, the LAI estimation method could be further validated with field measurements or Lidar techniques (Chasmer et al., 2012, Riaño et al., 2004).

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Commented [PA50]: discussion on LAI is moved up front to improve the focus of the paper

Plant cover, represented as LAI, proved to be the most important model parameter important parameter controlling total ET, and thereby the amount of groundwater recharge (Table ~~4~~3, Fig. 9~~11~~). The LAI parameter was included in the equations controlling both transpiration and soil evaporation, and therefore the sensitivity of the parameter is not surprising. While soil evaporation partly compensated for the lower transpiration with low LAI values, the total annual ET values progressively increased as a function of LAI (Fig. ~~11~~9). Interestingly, the simulations suggested that ET remains constant at constant level in the LAI range 0-1, potentially due to the sparse canopy changing the aerodynamic resistance and partitioning of radiation limiting soil evaporation, while still not contributing much to transpiration in total ET. This implies that the maximum groundwater recharge for boreal Scots pine remains rather constant up to a threshold LAI value of around 1. This knowledge can be used when co-managing forest and groundwater resources in order to optimize both.

Importance of LAI has been reported in earlier studies estimating groundwater recharge (Dripps, 2012, Keese et al., 2005, Sophocleous, 2000), but here the vegetation was represented with more spatially detailed patterns and a field data-based approach for LAI. According to previous studies, average ET from boreal conifer forests is around 2 mm d^{-1} during the growing season (Kelliher et al., 1998) , which is similar to our average value of 1.6 mm d^{-1} for the period 1 May-31 Oct. Some earlier studies have claimed that the influence of LAI on total ET rates from boreal conifer canopies is minor (Kelliher et al., 1993, Ohta et al., 2001, Vesala et al., 2005), but our simulation results indicate that higher LAI values lead to higher total ET values. The simulations showed that variable intensity of forestry, from low canopy coverage (LAI = 0-0.2) to dense coverage (LAI = 3.2-3.5) resulted in a difference of over 100 mm in annual recharge (Fig. 7~~9~~). It can be argued that the scenarios are unrealistic, because high LAI values, covering the whole study site, may not be achieved even with a complete absence of forestry operations. Nevertheless, the result demonstrates a substantial impact of forestry operations on esker aquifer groundwater resources.

~~Model validation showed that the modeling approach could reasonably reproduce (1) the main groundwater recharge events when compared to the WTF method and (2) the regional level of recharge compared to stream baseflow. The WTF estimates for recharge agreed with the simulations, with a slight tendency for higher estimates by the simulations. The discrepancy can be due to different assumptions behind the methods and uncertainty in local parameterization; in the WTF method for the specific yield and for simulations mainly the hydraulic conductivity which dictates the timing of recharge. However, there were overlapping estimates for almost every recharge event which shows consistency between the methods. The stream baseflow was lower than the long term average recharge, which was expected because~~

~~of the site hydrogeology. All of the outflow from the aquifer was likely not captured by the baseflow measurements as some of the water discharges to larger streams and lakes outside of the stream catchments (Rossi et al., 2014). When simulated recharge was extracted specifically for the baseflow measurements dates, the lower values for simulated recharge were also anticipated. The recharge displayed strong seasonal variability (see Fig. 7), but the discharge to streams is in general more stable because of the stabilizing effect of the groundwater storage. In conclusion, the order of magnitudes in the regional baseflow estimate and the simulation results were consistent. Despite the very different assumptions on which the modelling and field based methods were based, all provided similar estimates for groundwater recharge at the study site.~~

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~~There were different water flow rates at different depths (Fig. 7), demonstrating the role of the unsaturated zone in recharge. The high fluctuation in water flow at 1.5 m revealed the recharge dynamics in aquifers with a shallow water table. Such aquifers would be highly sensitive to annual fluctuations in recharge and respond rapidly to dry periods. On the other hand, rainy years would most likely replenish the aquifer water stores very quickly. Deeper in the soil profile, the response to wet and dry seasons was more modest, but still exhibited a clear seasonal signal. The water flow appeared to have dry and wet cycles of 5-10 years. Considering this, aquifers with unsaturated zones measuring tens of meters are likely to respond only to wet and dry cycles in climate patterns, rather than the weather in individual years. The temporal availability of the groundwater resource is most likely different for aquifers with different unsaturated zone geometry, as suggested by e.g. Hunt et al. (2008) and Smerdon et al. (2008).~~

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~~According to the simulations, the percentage of precipitation forming groundwater recharge varied considerably between years, as also reported in previous studies on transient recharge (Assefa and Woodbury, 2013, Dripps and Bradbury, 2010). Even though annual recharge was correlated with annual precipitation, and therefore years with high precipitation resulted in higher absolute recharge (Fig. 9), the percentage of effective rainfall did not increase as a function of annual sum of precipitation. This is somewhat surprising, because the rather constant evaporation potential between years (Fig. 8) and high soil hydraulic conductivity could be expected to result in a higher percentage of rainfall reaching the water table in rainy years. Some studies (Dripps and Bradbury, 2010, Okkonen and Kløve, 2010) have suggested that when the main annual water input arrives as snowmelt during the low evaporation season, it is likely to result in higher percentage recharge than in a year with little snow storage and~~

precipitation distributed evenly throughout summer and autumn, which may contribute to the variability in the recharge coefficient. However, when the maximum annual SWE value was used as a proxy for annual snowfall, there was no evidence of snow amount explaining the interannual variability in the recharge coefficient. Other factors contributing to recharge coefficient variability may be related to soil moisture conditions prior to snowfall, or the intensity of summer precipitation events (Smerdon et al., 2008, Stähli et al., 1999)(Stähli et al., 1999; Smerdon et al., 2008)(Smerdon et al., 2008, Stähli et al., 1999). Furthermore, the variability can to some extent be an effect of annual summation for the period 1 Oct-30 Sept, usually considered the hydrological year in the Nordic climate. Therefore the rainy autumn season is cut in 'half', and because recharge event comes with some delay from precipitation, the rainfall considered for a given year may not be reflected in the recharge for the year.

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The above-mentioned reasons make the concept of effective rainfall, which is currently routinely used to estimate groundwater recharge for groundwater management in e.g. Finland (Britschgi et al., 2009), susceptible to over- or under-estimation of actual annual recharge. This applies especially for aquifers with a thick unsaturated zone, where rainy years produce higher average recharge with some delay and for a longer duration (see Fig. 7). Therefore, if allocated water abstraction permits e.g. 50% effective rainfall coefficient to be assumed for each year, it potentially allows overuse of the resource during dry seasons. While aquifer storage can buffer occasional over-extraction, the lowering of the water table may diminish groundwater discharge to surface water bodies, depending on the geometry of the aquifer (Zhou, 2009).

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The method used here to estimate LAI from forestry inventories introduces a new approach for incorporating large spatial coverage of detailed conifer canopy data into groundwater recharge estimations. LAI values reported for conifer forests in Nordic conditions similar to the study site are in the range 1-3, depending on canopy density and other attributes (Koivusalo et al., 2008, Rautiainen et al., 2012, Vincke and Thiry, 2008, Wang et al., 2004)(Wang et al., 2004; Koivusalo et al., 2008; Vincke and Thiry, 2008; Rautiainen et al., 2012)(Koivusalo et al., 2008, Rautiainen et al., 2012, Vincke and Thiry, 2008, Wang et al., 2004). The LAI values obtained for the study site (mean 1.25) were at the lower end of this range. Furthermore, the data showed a bimodal distribution, with many model cells with low LAI (< 0.4) lowering the mean LAI. The low LAI values were not considered to be an error in data or calculations, but were in fact expected because of active logging and clearcutting activities in the study area. Although the equations to estimate LAI are empirical in nature and based on simplified assumptions, the

~~method can outline spatial differences in canopy structure. However, the LAI estimation method could be further validated with field measurements or Lidar techniques (Chasmer et al., 2012, Riaño et al., 2004)(Riaño et al., 2004; Chasmer et al., 2012)(Chasmer et al., 2012, Riaño et al., 2004).~~

~~Plant cover, represented as LAI, proved to be the most important model parameter determining the total recharge amount. This has been reported in earlier studies estimating groundwater recharge (Dripps, 2012, Keese et al., 2005, Sophocleous, 2000)(Sophocleous, 2000; Keese et al., 2005; Dripps, 2012)(Dripps, 2012, Keese et al., 2005, Sophocleous, 2000), but here the vegetation was represented with more spatially detailed patterns and a field data-based approach for LAI. According to previous studies, average ET from boreal conifer forests is around 2 mm~~ ~~d^{-1} during the growing season (Kelliher et al., 1998) , which is similar to our average value of 1.6 mm d^{-1} for the period 1 May-31 Oct. Some earlier studies have claimed that the influence of LAI on total ET rates from boreal conifer canopies is minor (Kelliher et al., 1993, Ohta et al., 2001, Vesala et al., 2005)(Kelliher et al., 1993; Ohta et al., 2001; Vesala et al., 2005)(Kelliher et al., 1993, Ohta et al., 2001, Vesala et al., 2005), but our simulation results indicate that higher LAI values lead to higher total ET values. While soil evaporation partly compensated for the lower transpiration with low LAI values, the total annual ET values progressively increased as a function of LAI (Fig. 11). Interestingly, the simulations suggested that ET remains constant at constant level in the LAI range 0-1, potentially due to the sparse canopy changing the aerodynamic resistance and partitioning of radiation limiting soil evaporation, while still not contributing much to transpiration in total ET. This suggests that the maximum groundwater recharge for boreal Scots pine remains rather constant up to a threshold LAI value of around 1. This knowledge can be used when co-managing forest and groundwater resources in order to optimize both.~~

~~The method allowed different land use scenarios in forestry management to be tested. The simulations showed that variable intensity of forestry, from low canopy coverage (LAI = 0-0.2) to dense coverage (LAI = 3.2-3.5) resulted in a difference of over 100 mm in annual recharge (Fig. 9). It can be argued that the scenarios are unrealistic, because high LAI values, covering the whole study site, may not be achieved even with a complete absence of forestry operations. Nevertheless, the result demonstrates a substantial impact of forestry operations on esker aquifer groundwater resources. Wider use of this method in Finland is practically possible, as~~

~~active forestry operations in Finland have yielded an extensive database on canopy coverage, which could be used in groundwater management.~~

The lichen layer covering the soil surface was explicitly accounted for in the simulation set-up, which to our knowledge is a novel modification. Kelliher et al. (1998) concluded that precipitation intercepted by lichen was an important source of understorey evaporation, especially directly after rain events. In addition, Bello and Arama (1989) reported that lichen could intercept light rain showers completely and that only intense rain events caused drainage from lichen canopy to mineral soil. While the lichen layer might have an increasing effect on soil evaporation through 'interception storage', Fitzjarrald and Moore (1992) suggest that a lichen cover may in fact have an insulating influence on heat and vapor exchange between soil and atmosphere, therefore impeding evaporation from the mineral soil. In the present study, the lichen layer appeared to have minor influence on total evaporation, soil evaporation and infiltration, as these variables showed only little sensitivity to lichen ~~Brooks and Corey~~B&C parameters (Table ~~4~~3). However, the approach to represent lichen with B&C model needs to better examined, as water retention capacity of lichen layer was treated equal to porosity, which is not strictly coherent with the Brooks and Corey (B&C) model. Nevertheless, the used approach successfully produced an additional interception storage of water in the correct range (generally 3-7 mm depending on random parameterization, data not shown). ~~However,~~~~Therefore~~ ~~m~~The performed~~ore intensive~~ laboratory measurement of lichen water retention ~~and~~should be supplemented with detailed analysis of lichen pressure-saturation ~~conduction properties is required~~curve and hydraulic conductivity to clarify the role of lichen in soil evaporation, and thereby groundwater recharge.

Commented [PA55]: Rev#2 spec com, violation of B&C model assumptions in lichen layer is discussed and further work in this area called for.

Stochastic variation of selected model parameters illustrated the uncertainties relating to numerical recharge estimation using the Richards equation in one dimension. The capability and robustness of the Richards equation to reproduce soil water content and water fluxes have been demonstrated extensively in various studies (Assefa and Woodbury, 2013, Scanlon et al., 2002b, Stähli et al., 1999, Wierenga et al., 1991). ~~However~~Therefore~~,~~ we considered that model calibration and validation with point observations of variables such as soil volumetric water content or soil temperature would not provide novel insights into water flow in unsaturated soils. Instead, we incorporated the parameter uncertainty ranges, usually used in model calibration, to the final recharge simulation output. An important outcome was that the uncertainty in the model output caused by different model parameterizations was small in comparison with the interannual variation in recharge. The error caused by uncertainty in the model assumptions or driving climate data was not addressed in this study. ~~We presume that for the given case study, the uncertainty and suitability of the driving climate data would introduce more uncertainty into the model output than model parameterization.~~

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~~While it can be argued that all relevant parameters were not included and parameter ranges could be more carefully determined, t~~T~~he parameter set used was able to provide information on parameter sensitivity. LAI was the most important parameter controlling total ET, and thereby the amount of groundwater recharge (Table 4, Fig. 11). The LAI parameter was included in the equations controlling both transpiration and soil evaporation, and therefore the sensitivity of the parameter is not surprising. However, LAI is a measurable parameter in the otherwise semi-empirical equations used to simulate evaporation, and physically based parameters are preferable to empirical fitting parameters in deterministic simulation approaches. Thus the ability of the approach to reduce a large part of model variability by allocating the LAI parameter spatially is a substantial advantage in reducing the model uncertainty.~~

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The sensitivity analysis focused on total cumulative values of ~~flux variables~~fluxes and did not address the temporal variations in the variables. ~~Therefore the soil hydraulic parameters showed only minor sensitivity, perhaps misleadingly.~~ Soil hydraulic parameters mainly influenced the timing of recharge through residence time in the soil, not so much the total amount. Therefore the soil hydraulic parameters showed only minor sensitivity, perhaps misleadingly. It should be noted that vertical heterogeneity in the soil profile hydraulic parameters can reduce the total recharge rates (Keese et al., 2005). However, vertical heterogeneities were ignored in this study not only to simplify the model, but also because the drilling logs showed only little variation in the area. Work of Wierenga et al. (1991) supports the simplification by showing that excluding moderate vertical heterogeneities does not significantly affect the performance of water flow simulations with the Richards equation. ~~Spatial differences in hydraulic parameters could be more accurately implemented in the modeling approach by creating a third zonation based on soil type, in addition to LAI and UZD. This would require the parameter ranges for hydraulic conductivity and Brooks and Corey parameters to be expanded to cover the properties of different soil types. Even then, the model is applicable only in situations where the soil type is permeable enough to allow rapid infiltration, so that surface runoff can be assumed to be of minor importance.~~

Commented [PA58]: Rev#2 spec com; assumption of vertical homogeneity justified

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Simulations acknowledged shallow water table contribution to evapotranspiration in an indirect, conceptual approach. Including a water table fixed at different depths in the soil profile would have been possible in the CoupModel setup. Influence of water table fixed at 2 m depth was tested and found to increase ET 3.5% for LAI values of 3, but for LAI values of 0.5 and 1.5 the increase in ET was only trivial. We expect only minor increase in ET with deeper water table configuration (with the given soil type), and therefore argue that excluding water table results in only minimal overestimation of total recharge at the study site. It should be noted that upward water fluxes were not excluded from the water flow time series and negative fluxes were considered as "negative recharge" at any depth. The simplification is made that water available for upward fluxes comes only form the soil moisture storage, not from the water table.

Commented [PA60]: Rev#1com; justification for excluding water table from the simulations

According to the simulations, the percentage of precipitation forming groundwater recharge varied considerably between years, as also reported in previous studies on transient recharge (Assefa and Woodbury, 2013, Dripps and Bradbury, 2010). Even though annual recharge was correlated with annual precipitation, and therefore years with high precipitation resulted in higher absolute recharge (Fig. ~~9~~7), the percentage of effective rainfall did not increase as a function of annual sum of precipitation. This is somewhat surprising, because the rather constant evaporation potential between years (Fig. ~~8~~10) and high soil hydraulic conductivity could be expected to result in a higher percentage of rainfall reaching the water table in rainy years. Some studies (Dripps and Bradbury, 2010, Okkonen and Kløve, 2010) have suggested that when the main annual water input arrives as snowmelt during the low evaporation season, it is likely to result in higher percentage recharge than in a year with little snow storage and precipitation distributed evenly throughout summer and autumn, which may contribute to the variability in the ~~recharge~~effective rainfall coefficient. However, when the maximum annual SWE value was used as a proxy for annual ~~snowfall~~snow storage, there was no evidence of snow amount explaining the interannual variability in the recharge coefficient. Other factors contributing to recharge coefficient variability may be related to soil moisture conditions prior to snowfall, or the intensity of summer precipitation events (Smerdon et al., 2008, Stähli et al., 1999).

The above-mentioned reasons make the concept of effective rainfall, which is currently routinely used to estimate groundwater recharge for groundwater management in e.g. Finland (Britschgi et al., 2009), susceptible to over- or under-estimation of actual annual recharge. This applies especially for aquifers with a thick unsaturated zone, where rainy years produce higher average recharge with some delay and for a longer duration (Zhou, 2009)~~ (see Fig. 7)~~. ~~While aquifer storage can buffer occasional over-extraction, the lowering of the water table may diminish groundwater discharge to surface water bodies, depending on the geometry of the aquifer (Zhou, 2009).~~

Commented [PA61]: removed to improve focus of discussion

5 Conclusions

A physically-based approach to simulate groundwater recharge for sandy unconfined aquifers in cold climates was developed. The method accounts for the influence of vegetation, unsaturated zone ~~depth~~thickness, presence of lakes, and uncertainty in simulation parameters in the recharge estimate. It is capable of producing spatially and temporally distributed groundwater recharge values with uncertainty margins, which are generally lacking in recharge estimates, despite understanding of uncertainty related to recharge estimates being potentially crucial for groundwater resource management. However, the parameter uncertainty defined for the study area was of minor significance compared with interannual variations in the recharge rates introduced by climate variations. ~~The uncertainty caused by model parameterization was decreased by allocating the LAI parameter spatially in the model area.~~

The simulations showed that Scots pine canopy, parameterized as leaf area index (LAI), was important in controlling the total amount of groundwater recharge. Forestry inventory databases were used to estimate and spatially allocate the LAI and the results showed that such inventories could be better utilized in groundwater resource management. Forest cuttings were demonstrated to increase groundwater recharge significantly. A sensitivity analysis on the parameters used showed that ~~understorey~~ soil evaporation could compensate for low LAI-related transpiration up to a LAI value of approximately 1, which may be important in finding the optimal level for forest management in groundwater resource areas. The concept of effective rainfall gave inconsistent estimates of recharge in annual timescales, showing the importance of using physically-based recharge estimation methods for sustainable groundwater recharge management.

Author contribution

P. Ala-aho and P.M Rossi collected and analyzed the field data. P. Ala-aho designed the simulation set-up, performed the simulations and interpreted the results. P. Ala-aho prepared the manuscript with contributions from all co-authors.

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Tables

Table 1. Characteristics of the study site annual climate.

VARIABLE	MEAN	STD
Precipitation [mm]	591	91
Air Temperature [°C]	-0.7	1.1
Reference ET [mm]	426	26

Table 2. Randomly varied parameters, related equations and parameter ranges included in the model runs. For full description of parameters and equations, see Jansson and Karlberg (2004).

Parameter	Part of the model affected	Range	Units	Source
LAI (leaf area index)	Transpiration	0 … 3.5	-	Data, see section 2.1.1
h (canopy height)	Transpiration	5 … 15	m	Data
r_{alai} (increase in aerodynamic resistance with LAI)	Soil evaporation	25 … 75	-	±50% , estimate
r_{Ψ} (soil surface resistance control)	Soil evaporation	100…300	-	±50% approximately to cover the surface resistance reported 150-1000 (Kelliher et al., 1998)
λ_L (pore size distribution index)	Soil evaporation, lichen	0.4 … 1	-	Estimate, to cover an easily drainable range of pressure-saturation curves
Ψ_L (air entry)	Soil evaporation, lichen	1.5 … 20	-	Estimate, to cover a easily drainable range of pressure-saturation curves

θ_L (porosity)	Soil evaporation, lichen	7.5…12.5	%	Data, lichen mean water retention ±SD from samples
$k_{mat,L}$ (matrix saturated hydraulic conductivity)	Soil evaporation, lichen	5 10^4 … 5 10^7	mm d^{-1}	Estimate, high K values assumed
t_{WD} (coefficient in the soil temperature response function)	Water uptake	10 … 20	-	±50%, estimate
Ψ_c (critical pressure head for water uptake reduction)	Water uptake	200…600	-	±50%, estimate
$k_{mat,S}$ (matrix saturated hydraulic conductivity)	Soil profile	1.707 10^3…127.2 10^3	mm d^{-1}	Data from soil sample particle size analysis
k_{minuc} (minimum unsaturated hydraulic conductivity	Soil profile	1 10^{-4} …1 10^{-1}	mm d^{-1}	Estimate k_{mat} * 1E-5
λ_s (pore size distribution index)	Soil profile	0.4 … 1	-	Range to cover measured pressure-saturation curves
Ψ_s (air entry)	Soil profile	20 … 40	-	Range to cover measured pressure-saturation curves
θ_s (porosity)	Soil profile	0.25…0.36	%	Range from soil samples
θ_r (residual water content)	Soil profile	0.01…0.05	%	Range to cover measured pressure-saturation curves

~~**Table 3.** Stream baseflow estimates compared to simulated recharge outputs calculated for different timeperiods~~

~~Baseflow for 9-10 February 2010 [mm a^{-1}]~~	~~Long term average recharge [mm a^{-1}]~~	~~Recharge for preceding year 2009 [mm a^{-1}]~~	~~Simulated recharge for 9-10 February 2010~~
~~312.7~~	~~362.8~~	~~421.8 (min)~~ ~~439.5 (max)~~	~~110.0 (min)~~ ~~135.8 (max)~~

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Table 43. Kendall correlation coefficient for simulation parameters and average annual sum of simulation output variables. ET = evapotranspiration, E = evaporation, for other symbols see Table 2.

Parameter	Total ET	Transpiration	Soil E	Snow E	Infiltration
LAI	0.59*	0.84*	-0.73*	-0.37*	0.18*
h	0.59*	0.84*	-0.73*	-0.37*	0.18*
r_{Ψ}	-0.11*	-0.03	-0.03	-0.61*	0.58*
r_{alai}	-0.13*	-0.02	-0.11*	0.03	-0.05
λ_L	-0.09*	-0.01	-0.11*	0.01	-0.03
Ψ_L	0.01	-0.04	0.11*	-0.04	0.06
θ_L	0.06	0.03	0.01	-0.00	0.09*
$k_{mat,L}$	-0.01	0.02	-0.04	-0.00	-0.00
$k_{mat,S}$	-0.10*	-0.04	-0.07*	0.02	0.01
k_{minuc}	-0.10*	-0.04	-0.07*	0.02	0.01
t_{WD}	-0.05	-0.02	-0.03	-0.05	0.03

Ψ_c	0.18*	0.12*	-0.02	-0.04	0.05
λ_s	0.13*	0.06	0.06	-0.00	-0.23*
Ψ_s	-0.11*	-0.05	-0.04	-0.05	0.04
θ_s	0.02	-0.01	0.03	0.10*	-0.18*
θ_r	0.07*	0.05	-0.01	0.01	0.16*

*Significant correlation, $p<0.05$

Figure captions

Figure 1. Recharge area of the Rokua esker aquifer. Boreholes in the area were used for model validation and soil type characterization. Baseflow was measured from streams originating outside the groundwater recharge area. Profile of cross-section A-B is presented in Fig. 2.

Figure 2. Cross-section A-B (Fig. 1) to demonstrate the geometry of the unsaturated zone and the aquifer (vertical axes exaggerated). A simulated soil profile is shown to give an example on how 1-D simulations are represented in the model domain, UZT represents the unsaturated zone thickness parameter.

Figure 3. Spatial distribution of leaf area index (LAI) and a 20m x 20m cell-based histogram of LAI values. In areas where forestry inventory data were lacking, a weighted average value of 1.25 was used in simulations.

Figure 4. Estimated ~~depth~~ thickness of the unsaturated zone in the model area and interpolation points for estimation of water table elevation.

~~**Figure 3.** Example of selection of water flow simulation data for a random cell in the model domain for which LAI = 0.1 and UZD = 5.2 m.~~

Figure 5. Flow chart of different evaporation processes considered in the study. Total evapotranspiration comprises of soil evaporation from the topmost soil layer, i.e. the lichen matrix, snow evaporation from snow surface, transpiration through the vascular system of tree canopy and lake evaporation from free water surface ~~and transpiration (Fig. 4)~~.

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Figure 6. Assemblage of simulated recharge for individual recharge events, shown as boxplots where circles represent the median, bold lines 25-75th percentiles of the simulations, thin lines the remaining upper and lower 25th percentiles and crosses are outliers. The location of the boxplots on the x-axis is the WTF estimate for a given recharge event using a specific yield value of 0.225. The dashed lines indicate the uncertainty in the WTF estimates caused by the selection of specific yield. The two estimates would agree perfectly (given the uncertainty in S_y) if all simulations shown as boxplots fell between the dashed lines.

~~**Figure 7.** Average water flow outputs at different soil profile depths with LAI range [1.2 … 1.4]. Y-axis is limited to 5 mm d^{-1} to highlight the flow dynamics in the deeper layers, even though peak signal at 1.5 m reaches a value of several cm annually.~~

Figure 7. Annual recharge time series from simulations where the black area covers the minimum and maximum values for different recharge samples. The annual recharge pattern closely followed trends in infiltration. Effects of different land use management practices over time on annual recharge rates are shown as high and low leaf area index (LAI) scenarios.

Figure 8. Spatial distribution of mean annual recharge, which was influenced mainly by the Scots pine canopy (LAI), the presence of lakes and, to some extent, areas with a shallow water table.

Figure 9. Example of scatter plots with the mean annual ET components are plotted as a function of the variable leaf area index (LAI), showing clear dependence of all ET components on LAI.

Figure 10. Values of different evapotranspiration (ET) components (mean and standard deviation) simulated for the study period.