Modelling snowpack dynamics and surface energy budget in boreal and subarctic peatlands and forests

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Abstract. The snowpack has a major influence on the land surface energy budget. Accurate simulation of the snowpack energy and radiation budget is challenging due to e.g. effects of vegetation and topography, and limitations in the theoretical understanding of turbulent transfer in the stable boundary layer. Studies that evaluate snow, hydrology and land surface models ~~(LSMs)~~ against detailed observations of all surface energy balance components at high latitudes are scarce. In this study, we compared different configurations of SURFEX land surface model against surface energy flux, snow depth and soil temperature observations from four eddy covariance stations in Finland. The sites cover two different climate and snow conditions, representing the southern and northern subarctic zones, and the contrasting forest and peatland ecosystems typical for the boreal landscape. We tested ~~the sensitivity~~ different turbulent flux ~~of surface energy fluxes to different process~~ parameterizations implemented in the Crocus snowpack model. In addition, we examined common alternative approaches to conceptualize soil and vegetation, and assessed their performance in simulating surface energy fluxes, snow conditions and soil thermal regime. Our results show that a stability correction function that increases the turbulent exchange under stable atmospheric conditions is imperative to simulate sensible heat fluxes over the peatland snowpacks, and that realistic peat soil texture (soil organic content) parameterization greatly improves the soil temperature simulations. For accurate simulations of surface energy fluxes and snow/soil conditions in forests, an explicit vegetation representation is necessary. Moreover, we demonstrate the high sensitivity of surface fluxes to a ~~previously rather~~ poorly documented parameter involved in snow cover fraction computation. Although we focused on models within the SURFEX platform, the results have broader implications for choosing suitable turbulent flux parameterization and model structures depending on the potential use cases for high-latitude land surface modelling.

1 Introduction

The boreal zone, characterized by a mosaic of seasonally snow-covered peatlands, forests and lakes, is the largest land biome in the world. Snow conditions in the boreal zone are rapidly changing due to climate warming, which is found to be the strongest during the cold seasons in the Arctic (Serreze et al., 2009; Screen and Simmonds, 2010; Boisvert and Stroeve, 2015; Rantanen et al., 2022). Evidently snow has an important role for water resources and human activities in the cold regions, but it is also known that the snowpack characteristics affect animal movement (Tyler, 2010; Pedersen et al., 2021) and plant distribution (Rasmus et al., 2011; Kreyling et al., 2012; Rissanen et al., 2021). Recent studies show that especially cold-dwelling species have been shifting towards higher latitudes and altitudes in search for more suitable habitats (Tayleur et al., 2016; Couet et al., 2022). Therefore, the rapid warming of the Arctic, and its consequences on the quantity and properties of snow may define the destiny of many species and human activities in the boreal region. To predict future snow conditions, environmental change, and the consequences for water resources, ecosystems and people, predictive and process-based models possess great potential (Clark et al., 2015; Boone et al., 2017). Land surface models (LSMs) have been used for decades in numerical weather prediction ~~(NWP)~~ and in global circulation models ~~(GCMs)~~ (Douville et al., 1995; Niu et al., 2011; Lawrence et al., 2019), and have more recently become common tools for interdisciplinary impact studies (Blyth et al., 2021).

Snowpack has a major impact on the wintertime energy budget due to its influence on the land surface albedo ~~(LSA)~~ (hereafter albedo) and the surface heat fluxes (Cohen and Rind, 1991; Eugster et al., 2000). The heat diffusion within the snowpack is determined by the surface heat fluxes, internal properties of the snowpack and soil thermal regime. Correctly representing the snowpack is thus essential for simulating energy and mass exchange between the snow surface and the atmosphere, as well as below the snowpack (e.g. surface temperatures and soil freezing/thawing dynamics, Koivusalo and Heikinheimo, 1999; Slater et al., 2001). The snowpack energy budget is partitioned into downward~~s~~ and upward~~s~~ shortwave ~~(SWD, SWU)~~ and longwave ~~(LWD, LWU)~~ radiation, turbulent fluxes of sensible ~~(H)~~ and latent heat ~~(LE)~~, snowpack-ground heat flux ~~(G)~~ and phase-changes in the snow. The snowpack energy balance and energy partitioning among the flux components vary strongly across diurnal and seasonal timescales, and between different ecosystems (Clark et al., 2011; Stiegler et al., 2016; Stigter et al., 2021). It is essential that LSMs are able to correctly reproduce this variability.

On the vast boreal and arctic peatlands with shallow vegetation, the snow cover can exclusively determine the wintertime ~~LSA~~ albedo (Aurela et al., 2015). With minimal solar radiation during winter months on these open snow fields, turbulent fluxes make an important component in the energy budget of the snowpack, as they compensate the radiative cooling processes and further contribute to snow melt (Lackner et al., 2021; Conway et al., 2018). Simulation of turbulent fluxes under stable atmospheric conditions is known as one of the major sources of uncertainty in snow models (Lafaysse et al., 2017; Menard et al., 2021). In LSMs the turbulent fluxes are commonly computed with bulk aerodynamic approaches, where ~~H~~ sensible and ~~LE~~ latent heat fluxes are proportional to the turbulent exchange coefficient according to the Monin-Obukhov similarity theory ~~(MOST)~~. These approaches typically use atmospheric stability correction functions based either on the bulk Richardson number (Martin and Lejeune, 1998; Lafaysse et al., 2017; Clark et al., 2015) or the Obukhov length scale (Jordan et al., 1999). It is established that ~~MOST~~ the Monin-Obukhov similarity theory does not well represent low-wind and stable atmospheric conditions above aerodynamically smooth surfaces such as snow (Conway et al., 2018). In such conditions, the simulated surface temperatures have been found to be unrealistically low, as the turbulent boundary layer tends to decouple from the snow surface (Derbyshire, 1999; Andreas et al., 2010). To circumvent this effect, stability correction functions have been modified to permit turbulent fluxes above critical stability thresholds (Lafaysse et al., 2017), by manipulating the wind speed (Martin and Lejeune, 1998; Andreas et al., 2010), or including a windless turbulent exchange coefficient (Jordan et al., 1999). Evaluations of these modifications often rely on validation with observed surface temperatures and snow depths (e.g. for the detailed snowpack model Crocus; Martin and Lejeune, 1998; Lafaysse et al., 2017) while comparisons against turbulent energy flux data remain scarce (Lapo et al., 2019; Conway et al., 2018).

The energy budgets of forest canopies and below-canopy snowpack are different to those on open peatlands, as turbulent exchange is attenuated by the canopy, and the snowpack energy budget and snow melt are mostly driven by the radiation balance (Rutter et al., 2009; Essery et al., 2009; Varhola et al., 2010). However, due to heterogeneous canopy structures and canopy processes (radiation transmittance, snow interception and unloading) together with low solar angles, the albedo dynamics of ~~LSA in~~ seasonally snow-covered boreal forests ~~is~~ are complex (Malle et al., 2021). The absorption of the shortwave radiation can be highly heterogeneous ~~in forest stands~~, having direct implications on canopy temperatures (Webster et al., 2017), and on the resulting longwave radiative fluxes between canopy, snowpack and the atmosphere (Mazzotti et al., 2020b). Forest snow modelling has been identified as a priority in advancing cold region climate and hydrological models (Rutter et al., 2009; Krinner et al., 2018; Lundquist et al., 2021). Various models that have been proposed to represent the large-scale impact of forest on the snowpack energy budget (Niu et al., 2011; Lawrence et al., 2019; Boone et al., 2017) are still prone to large errors, due to the complexity and unresolved spatial scales of the underlying physical processes (Loranty et al., 2014; Thackeray et al., 2019). The forest snow model evaluations against concurrent snowpack and surface energy balance data are also surprisingly scarce. For instance, the explicit forest scheme of SURFEX LSM, MEB (Boone et al., 2017) has so far been evaluated only against data from three neighbour sites in Saskatchewan, Canada (Napoly et al., 2020). This considerably limits knowledge of the model skill to represent snow-forest interactions in regional or global applications.

The texture and thermal properties of the underlying soil can strongly impact the snowpack-ground heat exchange, snowpack energy fluxes and snowpack dynamics (Decharme et al., 2016). Peatlands have high soil organic content (SOC) and are characterized by ~~a~~ high porosity, shallow water table, ~~a~~ weak hydraulic suction, strong gradient in hydraulic conductivity from high values at the top to low values at the subsurface, low thermal conductivity, and large heat capacity (Decharme et al., 2016; Marttila et al., 2021; Morris et al., 2022; Menberu et al., 2021). These properties result in a wet soil profile resistant to temperature variations, while the drier top peat and moss layer can also provide effective insulation particularly during summertime (Beringer et al., 2001; Park et al., 2018; Chadburn et al., 2015). The importance of the soil texture is still often overlooked even in detailed snow models. For instance, in model comparisons of the ESM-SnowMIP project (Ménard et al., 2019), no SOC information was used to parameterize the participating LSMs to the reference sites. In addition, many spatial snow simulations neglect peat soils or SOC altogether, and their hydrological and thermal characteristics are derived from fractions of sand, silt and clay (Vernay et al., 2022; Brun et al., 2013; Mazzotti et al., 2021; Richter et al., 2021)

The goal of this study is to evaluate the ability of SURFEX LSM (Masson et al., 2013) to describe the surface energy balance and its drivers in boreal and subarctic peatlands and forests. We evaluate the effect of alternative turbulent exchange and snowpack parameterizations, and examine the skills of alternative model configurations to represent the soil-snow-vegetation interactions. The modelling framework includes flexible parameterizations for different processes within Crocus snowpack model (Vionnet et al., 2012), and its coupling to ISBA (Noilhan and Mahfouf, 1996; Decharme et al., 2016) and MEB (Boone et al., 2017; Napoly et al., 2017) models enable assessments of soil-snow-vegetation interactions. We compare the model simulations against observed surface energy fluxes, snow depth and soil temperatures from two forest and two peatland sites in Finland. We focus on the snow cover period, but cover also the snow-free season for a reference. ~~On~~ At the peatland sites, we test the sensitivity of the surface heat fluxes to different turbulence and snow parameterizations, and assess how sensitive soil temperature and snowpack dynamics are to SOC. ~~On~~ At the forest sites, we compare the simulations of ISBA composite soil-vegetation and MEB big-leaf forest scheme to assess the suitability of different forest-snow model structures.

2 Materials and methods

2.1 Study sites

We consider coniferous forest and peatland ecosystems in southern and northern Finland. Both areas are located in the boreal biome and have seasonal snow cover (Fig. 1, Table 1). Site photos can be found in the Supplement (Fig. S8).

2.1.1 Pallas Supersite (N-WET and N-FOR)

The Pallas area represents northern subarctic conditions, and is characterized by pine and spruce forests, wetlands, fells and lakes (Aurela et al., 2015; Lohila et al., 2015; Marttila et al., 2021). In this study, we use data from its two eddy-covariance (EC) flux stations.

Lompolojänkkä (northern peatland, N-WET) is a pristine northern boreal mesotrophic sedge fen where the wetter parts are dominated by sedges (*Carex rostrata* (most abundant), *Carex chordorrhiza*, *Carex magellanica* and *Carex lasiocarpa*) and the drier parts consist of shallow deciduous trees (*Betula nana* and *Salix lapponum*). Moreover, the fen has a fairly low coverage of shrubs, mainly *Andromeda polyfolio* and *Vaccinium oxycoccos*. The vegetation height is shallow ($\sim$0.4 m), with exception of isolated trees/bushes on the drier edges of the peatland.

Kenttärova (northern forest, N-FOR) is a northern boreal spruce forest, located on a hill-top plateau with mineral soil approximately 60 meters above Lompolojänkkä wetland. The forest is dominated by Norway spruce (*Picea abies*) with some deciduous trees, mainly birch (*Betula pubescens*) but also aspen (*Populus tremula*) and pussy willow (*Salix caprea*). According to the classification by Brunet (2020), Kenttärova is a sparse forest. Both sites and their measurements have been described in detail by Aurela et al. (2015).

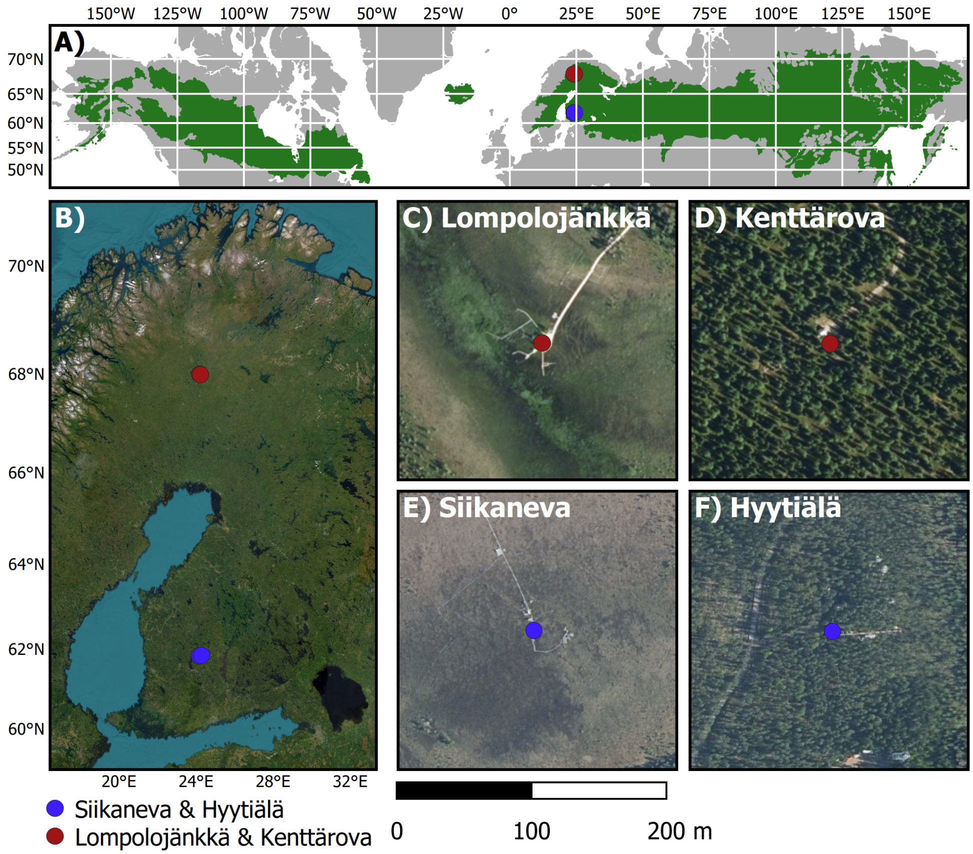

Figure 1. A) Study area locations inside the boreal land biome (green area, Olson et al., 2001), B) study site locations in Finland (Esri, 2023) and C-F) aerial images of each site (NLSF, 2020).

2.1.2 Hyytiälä and Siikaneva (S-WET and S-FOR)

The sites are located in southern subarctic conditions in the Pirkanmaa region in southern Finland, at about 5 km distance from each other. Siikaneva fen (southern wetland, S-WET) is a southern boreal oligotrophic fen dominated by sedges (*Eriophorum vaginatum*, *Carex rostrata* and *Carex limos*), and has an extensive Sphagnum cover (mainly *Sphagnum balticum*, *Sphagnum majus* and *Sphagnum papillosum*). The site has been described in detail in Aurela et al. (2007); Alekseychik et al. (2017), Rinne et al. (2018).

Hyytiälä (southern forest, ~~N-FOR~~ S-FOR) is a managed boreal Scots pine (*Pinus sylvestris*) dominated forest on mineral soil, described in detail by ~~Hari et al. (2013)~~ Launiainen (2010); Launiainen et al. (2022). According to the classification by Brunet (2020), the site is a dense forest.

Table 1. General site information.

Site	Code	Coordinates	Ecosystem	Soil type
Lompolojänkkä	N-WET	67°59.835’ N, 24°12.546’ E	mesotrophic fen	peat
Siikaneva	S-WET	61°49.961’N, 24°11.567’E	oligotrophic fen	peat
Kenttärova	N-FOR	67°59.237’ N, 24°14.579’ E	sparse spruce forest	podzol
Hyytiälä	S-FOR	61°50.471’ N 24°17.439’ E	dense pine forest	podzol

2.2 Models

We use components from the SURFEX LSM (Surface Externalisée, Masson et al., 2013) modelling platform. SURFEX was selected as its modularity and vast range of model structures and incorporated process parameterizations enable its use in diverse applications. Specifically, we used ISBA (Noilhan and Planton, 1989; Noilhan and Mahfouf, 1996) for composite soil-vegetation ~~(on both peatland and forest sites)~~, MEB (Boone et al., 2017; Napoly et al., 2017) for the canopy ~~(on forest sites)~~ and Crocus (Vionnet et al., 2012) and its ensemble/multiphysics version ESCROC (Lafaysse et al., 2017) for the snowpack simulations ~~(all sites)~~. Specifically, ISBA coupled to Crocus is used for both peatland and forest experiments (Fig. 2A,B) whereas MEB coupled to Crocus is only used for the forest experiments (Fig. 2C). In the next subsections, we briefly describe these model components ~~and parameterizations relevant to this study~~. Parameterizations and different configurations of ISBA, MEB and Crocus models are detailed in Sect. 2.3.

2.2.1 ISBA

ISBA (Interactions between the Soil, Biosphere and Atmosphere) is the soil and vegetation component of SURFEX (Noilhan and Planton, 1989; Noilhan and Mahfouf, 1996). It simulates the mass and energy fluxes in the soil-vegetation composite, as well as the exchanges between the soil-vegetation and the overlying atmosphere/snowpack (Fig. 2A,B~~,C~~). ISBA is used for the ~~GCM~~ general circulation models by Meteo-France (Mahfouf et al., 1995; Douville et al., 1995; Salas-Mélia et al., 2005; Voldoire et al., 2013, 2019) and for ~~NWP~~ numerical weather prediction in numerous countries (e.g. Hamdi et al., 2014; Bengtsson et al., 2017).

In ISBA, the surface heat flux between the atmosphere and the soil-vegetation composite (G_0, $\mathrm{Wm^{-2}}$) is computed as the residual of the sum of all surface/atmosphere energy fluxes:

$$G_0 = \underline{R_n}SWD(1-LSA)+\epsilon(LWD-\sigma T_s^4)+H+LE \tag{1}$$

where ~~R_n ($\mathrm{Wm^{-2}}$) is the net radiation~~ SWD ($\mathrm{Wm^{-2}}$) and LWD ($\mathrm{Wm^{-2}}$) are the incoming shortwave and longwave radiations, respectively. The land surface albedo is denoted as LSA, ϵ is the surface emissivity, σ is the Stefan-Boltzman constant and T_s (K) is the surface temperature. The sum of the radiation terms is hereafter denoted as R_n ($\mathrm{Wm^{-2}}$).~~, H ($\mathrm{Wm^{-2}}$) is the sensible~~

~~heat flux, and LE ($\mathrm{Wm^{-2}}$) is the latent heat flux. R_n is the sum of the net shortwave radiation and the net longwave radiation:~~

$$R_n = SWD(1-LSA)+\epsilon(LWD-\sigma T_s^4) \tag{2}$$

~~where SWD ($\mathrm{Wm^{-2}}$) and LWD ($\mathrm{Wm^{-2}}$) are the incoming shortwave and longwave radiations, respectively. The LSA is the land surface albedo, and ϵ the surface emissivity, σ the Stefan-Boltzman constant and T_s (K) the surface temperature.~~

The sensible heat flux H ($\mathrm{Wm^{-2}}$) is computed with the bulk aerodynamics approach:

$$H = \rho_a c_\rho C_H V_a (T_s - T_a) \tag{3}$$

where the air density, the specific heat capacity, the wind speed and air temperature are denoted with ρ_a ($\mathrm{kgm^{-3}}$), c_p, ($\mathrm{Jkg^{-1}K^{-1}}$), V_a ($\mathrm{ms^{-1}}$) and T_a (K), respectively. C_H is the turbulent exchange coefficient described later and is one of the parameters that is the focus of this study. When the soil is not covered by snow, the latent heat flux LE ($\mathrm{Wm^{-2}}$) is the sum of evaporation from the bare soil surface, E_g, evaporation of intercepted water on the canopy, E_c, transpiration from the vegetation, E_{tr}, and sublimation from bare soil ice, S_i:

$$LE = L_v(E_g + E_c + E_{tr}) + (L_f + L_v)(S_i) \tag{4}$$

where L_v ($\mathrm{Jkg^{-1}}$) and L_f ($\mathrm{Jkg^{-1}}$) are the latent heat of vaporization and fusion, respectively. Total evapotranspiration (ET) is computed as:

$$ET = E_g + E_c + E_{tr} = (1-veg)\rho_a C_H V_a [h_u q_{sat}(T_s) - q_a] + veg \rho_a C_H V_a h_v \tag{5}$$

where veg is the fraction of vegetation cover, $q_{sat}(T_s)$ ($\mathrm{kgkg^{-1}}$) is the saturated specific humidity at the surface, $q_a(T_s)$ ($\mathrm{kgkg^{-1}}$) is the atmospheric specific humidity, h_u is the dimensionless relative humidity at the ground surface related to the superficial soil moisture content and h_v is the dimensionless Halstead coefficient describing the E_c and E_{tr} partitioning between the leaves covered and not covered by intercepted water (see Noilhan and Mahfouf (1996) for details) ~~The sum of evaporation of canopy intercepted water (E_c) and transpiration (E_{tr}) is:~~

$$E_c + E_{tr} = veg \rho_a C_H V_a h_v [q_{sat}(T_s) - q_a], \tag{6}$$

~~where h_v is the dimensionless Halstead coefficient describing the E_c and E_{tr} partitioning between the leaves covered and not covered by intercepted water (see Noilhan and Mahfouf (1996) for details)~~.

The turbulent exchange coefficient C_H is based on the formulation of Louis (1979):

$$C_H = \left[\frac{k^2}{ln(z_u/z_{0t}) ln(z_a/z_{0t})}\right] f(R_i) \tag{7}$$

where z_u (m) is the reference height of V_a, z_a (m) is the reference height of T_a and humidity, z_{0t} (m) is the roughness height for heat, k (-) is the von Karman constant and $f(R_i)$ (-) describes the decrease of C_H as a function of increasing atmospheric stability, represented through Richardson number (R_i) (Louis, 1979).

Instead of separate treatment of the vegetation canopy and ground, ISBA considers the composite soil-vegetation energy budget (Fig. 2A,B~~,C~~). In the most detailed soil scheme ISBA-Diffusion (ISBA-DIF, Boone et al., 2000; Decharme et al., 2011), used in this study, 1D Fourier law is used to solve the soil heat diffusion, while a mixed-form Richards equation is applied for the 1D soil water movements. Similar as in Napoly et al. (2020), we use the $A-g_s$ stomatal ~~resistance~~ conductance formulation derived from the coupling of photosynthetic CO_2 demand and stomatal function (Calvet et al., 1998).

ISBA uses parameters such as one-sided leaf area index (LAI, $\mathrm{m^2m^{-2}}$), vegetation height, vegetation thermal inertia ~~($\mathrm{Km^2J^{-1}}$)~~, albedos of soil and vegetation, fractions of sand and clay as well as SOC content to characterize the composite soil-vegetation column. These parameters may be defined by the user, or obtained from global or regional databases (e.g. Faroux et al., 2013) and pedotransfer functions (Noilhan and Mahfouf, 1996; Peters-Lidard et al., 1998). In the presence of full snow cover, the surface energy budget is solved by Crocus (Sect. 2.2.3). For partial snow cover, Crocus is used to solve the snow covered fraction while the energy balance of the snow-free fraction is computed by ISBA, and total surface energy fluxes are computed as weighted averages of the snow and snow-free fractions (Sect. 2.3.1).

2.2.2 MEB

MEB (Multi-Energy Balance) is a recent ISBA development to explicitly describe vegetation and soil energy and mass balances. It was developed initially for forests (Boone et al., 2017; Napoly et al., 2017) and found to yield improved snow and soil temperature simulations (Napoly et al., 2020) but has not been evaluated for boreal and subarctic conditions. MEB simulates surface energy budget separately for soil and vegetation canopy ~~(a two-source model)~~. When the ground is snow covered, the energy budget of the snowpack is also explicitly represented ~~(i.e. a three-source model is applied)~~. We used the MEB option, where the forest floor is covered by a litter layer instead of the bare soil surface (Napoly et al., 2020) (Fig. 2~~A~~C). ~~MEB uses a big-leaf approach, meaning that the entire vegetation canopy is lumped into a single effective 'leaf' (Boone et al., 2017)~~. MEB describes vegetation canopy as a single big leaf (Boone et al., 2017). The respective energy balance equations for the canopy, the snowpack and the ground surface/litter layer in MEB are:

$$\begin{cases} C_v \frac{\partial T_v}{\partial t} = R_{nv} - H_v - LE_v + L_f \phi_v \\ C_{g,1} \frac{\partial T_{g,1}}{\partial t} = (1-\rho_{sng})(R_{ng} - H_g - LE_g) + \rho_{sng}(G_{gn} + \tau_{n,Nn} SW_{nn}) - G_{g,1} + L_f \phi_{g,1} \\ C_{n,1} \frac{\partial T_{n,1}}{\partial t} = R_{nn} - H_n - LE_n - \tau_{n,1} SW_{nn} + \epsilon_{n,1} - G_{n,1} + L_f \phi_{n,1} \end{cases} . \tag{8}$$

where C_v, $C_{g,1}$, $C_{n,1}$ ($\mathrm{Jm^{-2}K^{-1}}$) and T_v, $T_{g,1}$, $T_{n,1}$ K are the effective heat capacities and temperatures of the canopy, ground surface/litter layer and snowpack, respectively. In these equations, the subscripts $g,1$ and $n,1$ represent the uppermost layer for the soil and the snowpack, respectively. $G_{g,1}$ and $G_{n,1}$ are respectively the conduction heat flux at the bottom of the uppermost soil or snow layer. G_{gn} is the conduction heat flux at the soil-snow interface. R_{nv}, R_{ng}, R_{nn} ($\mathrm{Wm^{-2}}$) are net radiation, i.e. the sum of net shortwave radiation and net longwave radiation from/to the corresponding layer. The shortwave radiation scheme used in MEB is described ~~in detail~~ in Carrer et al. (2013). Light transmission through the canopy is computed with a so-called sky view factor, which depends on LAI, solar angle and a vegetation dependant -constant (see Eq. 45 in Boone et al., 2017). H and LE flux parameterization as well as the stability correction functions are detailed in Boone et al. (2017). Obviously T_v, $T_{g,1}$, $T_{n,1}$ are also involved in the radiative and turbulent terms, providing a linear system of equations to be solved by an implicit numerical scheme. In this study, MEB is coupled to the ISBA-DIF soil scheme (Sect 2.2.1) and the snowpack model Crocus (Sect. 2.2.3). Energy fluxes between the canopy and the ground surfaces are calculated within MEB, and prescribed as upper boundary conditions in the subsequent Crocus and ISBA-DIF calculations.

2.2.3 Crocus

Crocus is a 1D physically based multilayer snowpack model (Vionnet et al., 2012). It is the most detailed snow scheme in ISBA, and has been used for operational avalanche hazard forecasting in the French mountain ranges for the past three decades (Morin et al., 2020). It aims to mimic the vertical layering of snowpacks with a Lagrangian discretization system, avoiding the aggregation of snow layers with highly different physical properties. A detailed description of Crocus and its integration in SURFEX can be found in Vionnet et al. (2012).

~~In Crocus, the vertical heat diffusion in the snowpack is solved with an implicit backward-difference integration method (Boone and Etchevers, 2000). The snow effective thermal conductivity, k, follows Yen (1981):~~

$$k = k_{ice}(\frac{\rho}{\rho_w})^{1.88} \tag{9}$$

~~where k_{ice} is the thermal conductivity of ice, ρ_w is the density of liquid water and ρ is the density of snow.~~ The snowpack surface ~~net energy flux~~ energy budget is the sum of net radiation, turbulent fluxes and advective fluxes from precipitation. Over the snow, the sensible heat flux is computed similarly as in ISBA (Eq. 3) for soil surface, while the latent heat flux (sublimation/deposition), LE_s, is computed as:

$$LE_s = (L_f + L_v)\rho_a C_H V_a [q_{sat}(T_s - q_a)], \tag{10}$$

where T_s (K) is the snow surface temperature. The bottom of the snowpack and the uppermost soil layer of ISBA are fully coupled with a mass and energy conserving semi-implicit solution. The semi-implicit solution refers to a coupled system in which both components are solved separately with an implicit approach considering that the state of the second system remains constant during the solving of the first system. The heat conduction flux G_{gn} at the snow-soil interface is explicitly computed using the Fourier equation, and depends on the temperature gradient between the bottom snow layer and the uppermost soil layer (Eq. 4 in Decharme et al., 2011). The soil thermal conductivity and heat capacity are described using pedotransfer functions of ISBA (Noilhan and Mahfouf, 1996; Peters-Lidard et al., 1998).

2.3 Model configurations and parametrization

2.3.1 Model configurations

We use three different configurations of ISBA, MEB and Crocus modules (Fig. 2). ~~The first configuration (Fig. 2CA) is the big-leaf approach where the fluxes between the canopy and ground are explicitly computed by MEB, and prescribed in~~

~~the subsequent Crocus snowpack and ISBA-DIF soil modules (later denoted as MEB).~~ The first two ~~other~~ configurations use the composite soil-vegetation conceptualization of ISBA (Fig. 2A,B~~,C~~), and differ only in how the snow cover fraction is represented ~~over the soil-vegetation composite~~. ISBA aggregates the properties of soil and vegetation depending on ~~a so-called~~ vegetation fraction (veg) that covers a given grid-cell.

The first configuration (referred as ISBA-FS) assumes the snowpack to fully cover the soil-vegetation composite regardless of the snow depth (Fig. 2~~C~~A). It is the common approach for snow simulations over shallow vegetation or bare soil (Vernay et al., 2022; Nousu et al., 2019), for large-scale reanalyses (Brun et al., 2013) and some hydrological applications (Lafaysse et al., 2011; Revuelto et al., 2018). In addition, most site-level evaluations of SURFEX snow schemes rely on ISBA-FS configuration (Decharme et al., 2016; Lafaysse et al., 2017).

In the second configuration (referred as ISBA-VS), a part of the soil-vegetation composite is covered by snow while the remaining (non-snow) fraction stays in constant contact with the atmosphere. This proportion is governed by snow depth. The effective snow cover fraction is the weighted average between the snow fraction of vegetation (p_{snv}) and snow fraction of the bare ground (p_{sng}), calculated as (Decharme et al., 2019; Napoly et al., 2020):

~~In the second configuration Then, a dynamic snow fraction determines the part of the soil-vegetation composite that is covered by snow while the remaining (non snow) soil-vegetation fraction stays in constant contact with the atmosphere.. This model version is later denoted as ISBA varying snow cover (ISBA-VS, Fig. 2B). The effective snow cover fraction is defined as the weighted average between the snow fraction of vegetation (p_{snv}) and snow fraction of the bare ground (p_{sng}), calculated as (Decharme et al., 2019; Napoly et al., 2020):~~

$$p_{sn} = veg\, p_{snv} + (1 - veg)\, p_{sng} \tag{11}$$

$$p_{snv} = min(1.0, \frac{HS}{HS + w_{sw} z_0}) \tag{12}$$

$$p_{sng} = min(1.0, \frac{HS}{HS_g}) \tag{13}$$

where HS (m) is the ~~height of snow~~ snow depth, HS_g is the threshold value for ~~height of the snow~~ snow depth (0.01 m by default), and z_0 (m) denotes the surface roughness. The coefficient w_{sw} is supposed to relate to scale-dependent vegetation characteristics and is assigned as 5 by default in SURFEX and in ~~NWP~~ numerical weather prediction configurations (as well as in this study). However, without clear consistency, highly different values of w_{sw} have been used e.g. in climate simulations (w_{sw} = 2 by Decharme et al., 2019) and hydrological applications (w_{sw} = 0.2 by Le Moigne et al., 2020). We present a summary of the application specific treatment of the snow cover fraction and w_{sw} in the Appendix (Table C1). This summary shows that selection of w_{sw} value seems arbitrary and the fractional concept is only loosely linked to any physical relationships between soil, vegetation and snow. Yet, it is necessary for such a composite approach.

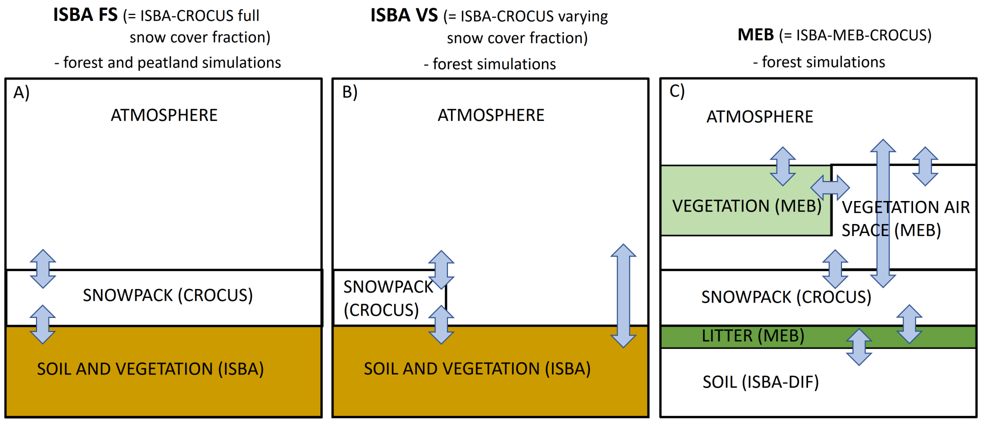

Figure 2. Three different model configurations used in the study: (A) ISBA-FS with full snow cover fraction, (B) ISBA-VS with varying snow cover fraction, and (C) MEB big-leaf approach. Considered energy fluxes between domains are represented with arrows.

~~The third configuration is the common approach for snow simulations over shallow vegetation or bare soil (Vernay et al., 2022; Nousu et al. 2019) and for some large-scale reanalyses (Brun et al., 2013). It assumes that the snowpack is fully covering the soil-vegetation composite, and snow cover fraction is unity regardless of the snow depth (Fig. 2~~C~~A). This version is later denoted as ISBA full snow cover (ISBA-FS). To our knowledge it has not been used in coupled applications with atmospheric models but frequently in hydrological applications (Lafaysse et al., 2011; Revuelto et al., 2018). Most existing site-level evaluations of SURFEX snow schemes also rely on ISBA-FS configuration (Decharme et al., 2016; Lafaysse et al.; 2017).~~

The third configuration (later denoted as MEB) is the big-leaf approach where the fluxes between the canopy and snowpack/ground are explicitly computed by MEB, and prescribed in the subsequent Crocus snowpack and ISBA-DIF soil modules (Fig. 2C) .

2.3.2 ESCROC parameterizations for snow processes and turbulent exchange

We use the multiphysics version of Crocus (ESCROC, Ensemble System Crocus, Lafaysse et al., 2017) to evaluate the impact and associated uncertainties of the different parameterizations of snow processes and turbulent exchange. In ESCROC, the main physical processes and properties of snowpack, as well as the turbulent fluxes, can be represented by several alternative options. These include density of new snow, snow metamorphism, absorption of solar radiation, turbulent fluxes, thermal conductivity, liquid water holding capacity, snow compaction and surface heat capacity (Eqs. 1-17 in Lafaysse et al., 2017). Lafaysse et al. (2017) ~~have shown that consideration of all these combinations is numerically expensive and often unnecessary to depict the overall uncertainty. Indeed,~~ showed that an optimized standard subensemble of 35 members (E_2 subensemble) ~~has been found~~ is sufficient to provide a spread of the appropriate magnitude compared to model errors (Lafaysse et al., 2017). ~~In this work we used~~ We use this subensemble (hereafter ESCROC-E_2) similar to recent studies quantifying the model uncertainty (e.g.

Deschamps-Berger et al., 2022; Tuzet et al., 2020). ~~In our case, the~~ The presented ensemble spread correspond to simulated values between ensemble minimum and maximum.

In Crocus, the default turbulent exchange parameterization (Eq. 7) has been found to underestimate the turbulent fluxes under stable conditions (Martin and Lejeune, 1998). Therefore, different stability dependencies of the C_H have been implemented in ESCROC-E_2. They differ mainly in the R$_i$ thresholds below which C_H is assigned a constant value to enable turbulent heat and mass transport under stable conditions. As shown by Fig. 4 in Lafaysse et al. (2017), these parameterizations are a) classical Louis (1979) formula (later referred as RIL) with threshold at R$_i$ = 0.2, b) RIL with threshold at R$_i$ = 0.1 (RI1), c) RIL with threshold at R$_i$ = 0.026 (RI2), and d) modified formulation with effective roughness length for heat ($10^{-3}\mathrm{m}$), minimum wind speed (0.3 ms^{-1}), and with threshold at R$_i$ = 0.026 (M98) by Martin and Lejeune (1998). ~~Although the~~ The RIL parameterization is widely used in SURFEX applications (e.g. Decharme et al., 2019; Le Moigne et al., 2020), ~~while~~ RI2 ~~parameterization~~ is applied in operational snow modelling in the Alpine area (Vernay et al., 2022), and M98 was recently used in the Canadian Arctic by Lackner et al. (2021). However, evaluations of the different Crocus turbulent flux parameterizations against surface flux data are still lacking. MEB uses a different stability correction term (Boone et al., 2017) and applies only the RIL option for the stable conditions. ~~While The MEB simulations (with ESCROC-E_2) are based on the E_2 subensemble as well, they therefore only use the RIL turbulent exchange parametrization for all members.~~

2.3.3 Site parameters

The parameterization of ISBA and MEB for the study sites is given in Table 2. Summer LAI and vegetation height were obtained from literature, while winter LAI (and monthly LAI cycle) was estimated according to the proportion of deciduous and coniferous vegetation on each site. The LAI of S-FOR refers to conditions before forest thinning in early 2020. The thinning, resulting in ca. 35% reduction in LAI, was neglected in our simulations as major part of the simulation period covers time before the thinning. Vegetation types in ISBA are characterized according to ECOCLIMAP (Champeaux et al., 2005); the forest sites in this study classify as boreal needleleaf evergreen ~~(BONE)~~, while the peatland sites are best represented as boreal grass ~~(BOGR)~~. Additional parameters based on LAI, vegetation height and vegetation type are computed following the standard methods of ISBA (Noilhan and Mahfouf, 1996; Carrer et al., 2013).

Soil texture (sand and clay fractions) for the forest sites are based on *in situ* measurements. The peat soils at S-WET and N-WET were parameterized as fully organic for the uppermost 1 m, in accordance with field measurements (Väliranta and Mathijssen, 2021; Muhic et al., 2023), while the deeper layers were assigned as mineral soil similar to the contiguous forests. Although peat profiles may be deeper, the soils below the damping depth of annual temperature fluctuations (ca 1.1 m for saturated peat soil with porosity ca 90 %) are assumed not to have significant impact on surface energy flux dynamics. The SOC values for mineral soils of N-FOR and S-FOR were taken from Lindroos et al. (2022). The rest of the parameters presented in Table 2 were assigned as estimates. The thermal and water retention parameters are subsequently derived from the pedotransfer functions of ISBA (Noilhan and Mahfouf, 1996; Peters-Lidard et al., 1998).

Table 2. Main model parameters for the study sites. Vegetation types BOGR and BONE correspond to boreal grass and boreal needleleaf evergreen, respectively.

Parameter	N-WET	S-WET	N-FOR	S-FOR	Source
Veg. type	BOGR	BOGR	BONE	BONE	ECOCLIMAP: Champeaux et al. (2005)
Veg. fraction (only with ISBA-VS) (-)	0.95	0.95	0.95	0.95	ECOCLIMAP: Champeaux et al. (2005)
Veg. height (m)	0.4	0.25	13	15	Aurela et al. (2015); Alekseychik et al. (2017) Kolari et al. (2022)
LAI$_{max}$ ($\mathrm{m^2m^{-2}}$)	1.3	0.6	2.1	3.0	Aurela et al. (2015); Alekseychik et al. (2017) Kolari et al. (2022)
LAI$_{min}$ ($\mathrm{m^2m^{-2}}$)	0.3	0.1	1.9	2.4	assigned
Veg. albedo (NIR/VIS) (-)	0.136	0.187	0.145	0.145	assigned
Soil albedo (NIR/VIS) (-)	0.136	0.187	0.145	0.145	assigned
Tair measurement height (m)	2	2	2	2	FMI (2021)
Wind measurement height (m)	13	3	23	16.8	Aurela et al. (2015); Mammarella et al. (2019) Alekseychik et al. (2022a)
Elevation (m)	270	162	347	181	Hari et al. (2013); Alekseychik et al. (2022a); FMI (2021)
Clay (%) (below 1 m at peatlands)	9	7	9	7	measurements
Sand (%) (below 1 m at peatlands)	76	65	76	65	measurements
SOC (0-30cm) ($\mathrm{kgm^{-2}}$)	93,5	93,5	3.0	3.5	Lindroos et al. (2022) Muhic et al. (2023); Väliranta and Mathijssen (2021)
SOC (30-70cm) ($\mathrm{kgm^{-2}}$)	93,5	93,5	1.75	0.75	Lindroos et al. (2022) Muhic et al. (2023); Väliranta and Mathijssen (2021)
SOC (70-100cm) ($\mathrm{kgm^{-2}}$)	93,5	93,5	0	0	Väliranta and Mathijssen (2021); Muhic et al. (2023)
Start of simulation (yyyy-mm)	2013–09	2016–09	2013–09	2008–09	-
End of simulation (yyyy-mm)	2021–07	2021–07	2021–07	2021–07	-

2.4 Data

2.4.1 Model forcing

Meteorological forcing consist of hourly observations of air temperature, wind speed, precipitation rate, air humidity, downward shortwave and longwave radiation ~~T_a, V_a, precipitation rate (P), q_a(T_s), SWD and LWD~~ and atmospheric pressure. The available meteorological observations from the nearest meteorological stations were obtained from Finnish Meteorological Institute (FMI) open database (FMI, 2021) (Station IDs: N-WET 778135, N-FOR 101317, S-FOR 101987). Meteorological observations at the S-WET site come from the SMEAR database (Alekseychik et al., 2022a). At S-WET and S-FOR the shortwave and longwave radiation were obtained from the SMEAR database, while at N-WET and N-FOR data from FMI stations were used. The diffuse to total shortwave radiation ratio, r, was estimated as a function of the cosine of the sun zenith angle, μ.

More specifically, a 3rd degree polynomial fit between r and μ was obtained using the atmospheric model SBDART (Ricchiazzi et al., 1998) to simulate diffuse and total solar radiation in clear sky conditions. The atmospheric profile was set to typical winter conditions, 0.09 for the aerosol optical thickness, 300 DU for the ozone column and 0.854 gcm^{-2} for the water vapor column.

The data gaps in meteorological observations were first filled by the contiguous sites (e.g. N-FOR for N-WET and vice versa) and the remaining gaps by other nearby meteorological stations (IDs: N-WET/N-FOR 101932, S-WET/S-FOR 101520). The missing radiation observations were first filled by the contiguous sites, and the remaining gaps by ERA5 reanalysis data (Hersbach et al., 2020). Only ~~a little~~ less than 10 hours of ERA5 data was used for N-WET, N-FOR and S-WET ~~less than 10 hours~~. However, S-FOR radiation observations contained more gaps, specifically LWD in 2008–2012. ~~and thus, a comparison of site observations and ERA5 estimates is provided in the Appendix (Fig. B1).~~.~~Overall the agreement of ERA5 and observed LWD is good.~~ A good agreement between site observations and ERA5 estimates of LWD is shown in Appendix (Fig. B1). Furthermore, fraction of snow to total precipitation P_{ice}/(P_{ice}+P_{liq}) is assumed to linearly decrease from 1 to 0 at air temperatures between 0°C and 1°C. ~~Furthermore, precipitation rate was split between snow and rain based on T_a~~:

~~*** DELETED EQUATION ***~~

~~where P_{ice} and P_{liq} denote the snowfall and rainfall rates, respectively. Between 0°C and 1°C the fraction of ice/snow changes linearly (a = 1 - b).~~

2.4.2 Model evaluation data

We use surface energy flux observations, ~~height of snow~~ snow depth ~~(HS)~~ and soil temperatures in model evaluation. The availability period of each variable is given in Table B1. On all sites, upward~~s~~ shortwave radiation (SWU) and upward~~s~~ longwave radiation (LWU) were measured using pyranometers and pyrgeometers, while ground heat flux (G) was measured using soil heat flux plates between 5 and 10 cm depths.

The sensible heat (H) and latent heat (LE) were measured by the eddy-covariance (EC) technique. The EC systems consist of USA-1 (METEK) three-axis and Gill HS-50 sonic anemometers as well as closed-path LI-7000 and LI-7200 (Li-cor, Inc.) CO_2/H_2O analysers. The detailed descriptions of the instrumentation, footprint analysis and the procedures for obtaining the turbulent heat fluxes from raw eddy covariance data is detailed in the original data and site publications ~~by Aurela et al., 2015, Mammarella et al., 2016, Mammarella et al. 2019, Aleksyichnik et al., 2022.~~ (N-WET/N-FOR; Aurela et al. (2015), and S-WET/S-FOR; Mammarella et al. (2016, 2019); Alekseychik et al. (2022b). In short, the ~~sensible and latent heat fluxes~~ H and LE were screened for instrument failure and data outliers, and data ~~quality flags were made~~ was quality flagged according to friction velocity (u_*) and flux stationarity (FST) criteria (Foken et al., 2005):

- flag 2: all data (after screening of instrument failures and outliers)
- flag 1: $u_* \geq 0.1\ \mathrm{ms}^{-1}$ and 0.3 $\leq$ FST $\leq$ 1.0
- flag 0: $u_* \geq 0.1\ \mathrm{ms}^{-1}$ and FST $\leq$ 0.3

At S-WET, S-FOR and N-FOR, automated ~~HS~~ snow depth observations are directly used (FMI, 2021). On N-WET the automated ~~HS~~ snow depth measurement is at 0.7 m height, and therefore the exceeding snow depths were taken from biweekly manual measurements. To account for the spatial variability of snow depth in the forests, manual ~~HS~~ snow depth measurements from a snow course in the close proximity of the automated measurements were used (Aalto et al., 2022; Marttila et al., 2021). Each site has different configuration of soil temperature sensors. At N-FOR and N-WET stations, soil temperatures are measured at 5 and 20 cm depths (Aurela et al., 2015). Soil temperatures at S-FOR and S-WET are measured at depths of 0, 5, 10, 30, 50 and 75 cm (Aalto et al., 2022).

2.5 Model experiments

On the peatland sites, we evaluate the skill of ISBA-FS (Sect. 2.3.1) and effect of ESCROC-E_2 parameterizations (Sect. 2.3.2) on surface heat fluxes over snowpack and bare ground. The simulations are further used to assess the differences in ~~HS~~ snow depth simulations ~~and soil temperature~~ between ESCROC-E_2 turbulent exchange options. ~~For a more detailed evaluation of two contrasting turbulent exchange options within ESCROC, we conducted deterministic ISBA-FS simulations with i) site parameters as shown in Table 2. and all default ESCROC-E_2 parameterizations as in Fig 2. in Lafaysse et al. (2017) (processes listed in Sect. 2.3.2, referred as RIL-SOC), and ii) site parameters as shown in Table 2. and all the default ESCROC-E_2 parameterizations except the turbulent exchange option switched to M98 (referred as M98-SOC).~~ For a more detailed evaluation of two contrasting turbulent exchange options, we conducted deterministic ISBA-FS simulations with site parameters and default ESCROC-E_2 snow parameterizations, but different treatments of turbulent exchange; RIL and M98 (see Sect. 2.3.2, referred as RIL-SOC and M98-SOC). Moreover, the influence of soil texture was explored by comparing M98-SOC to simulation where soil was parameterized as mineral, similar to the contiguous forest site (referred as M98-MIN). ~~Moreover, we explore the influence of soil texture on the soil thermal regime and on snowpack dynamics. Hence, an additional deterministic simulation was conducted where the soil was characterized as mineral soil, as had been measured and used for the contiguous forest site, while turbulent exchange was set to M98 (referred as M98-MIN). The M98-MIN simulation was compared to the previously described M98-SOC simulation, where the soil was characterized as fully organic until 1 m depth (Table 2).~~

On the forest sites, we examine the skills of the different alternatives to represent the energy and mass budgets of soil and vegetation (ISBA-VS, ISBA-FS, MEB in Sect. 2.3.1), and their implications on HS, soil temperature and surface energy fluxes. First, we compare ESCROC-E_2 simulations with these three configurations focusing on the ~~HS~~ snow depth and soil temperature. The ISBA-VS simulations are conducted with the default snow cover fraction parameterization (Eq. 11). For a more detailed comparison of the simulated and observed above-canopy surface energy fluxes by ISBA-VS and MEB, we conducted deterministic simulations with the default Crocus parameterizations (as in Fig 2. in Lafaysse et al. (2017)).

Model simulation periods for each site are in Table 2. For each site, the model initial state was obtained by a spin-up simulation from the start date (Table 2) to September 2020. In total of ca. 290 ensemble and deterministic simulations were conducted.

2.6 Model evaluation metrics

Time series ~~plots~~ of daily averages ~~d variables~~ are used to represent the results, whereas mean absolute error (MAE), mean bias error (MBE) and coefficient of determination (R^2) are used in quantitative model-data comparison. To detect possible biases in model simulations, we use scatter plots and quantile-quantile plots of sorted observations against sorted simulations. The sign convention is so that the surface energy fluxes are presented relative to the surface (i.e. negative flux means that surface is losing energy). In time series ~~plots~~, the turbulent flux observations include all EC data (Sect 2.4.2, quality flag $\leq$ 2). We demonstrate the results by using the winter season 2018–2019 as an example time series period thanks to its best coverage of energy flux data (least gaps), and typical ~~representation of the~~ snow conditions on all ~~the~~ sites. For the scatter and quantile-quantile plots, only flux data with quality flag $\leq$ 1 is used, and the results computed as aggregated 6-hour means include ~~the full~~ periods where ~~simulations and~~ observations are available (referred later as evaluation period). We compare snow and snow-free conditions by grouping the results into time windows where models and observations agree of the ground conditions (snow or snow-free).

3 Results

3.1 Observed energy balance at peatland and forest sites

The energy budget at high-latitudes have a strong seasonal variation driven by solar radiation (Fig. 3). In winter (December, January, February), longwave radiation balance to large extent determines R_n, particularly in the northern Finland. Daily average R_n is negative down to -50 Wm^{-2} and lower, which implies considerable radiative cooling. Towards spring the radiation budget is gradually counterbalanced by shortwave radiation. On the peatlands, a large fraction of SWD is reflected during snow cover, and daily R_n turns positive in late melting season (Fig. 3A,C). At the forest sites, the timing of R_n becoming positive is less sensitive to the presence of snow ~~on the ground~~, as a large proportion of ~~the~~ SWD is absorbed by ~~the~~ vegetation. In summer, high solar elevation and the absence of the reflective snow surface cause daily R_n to be up to 200 Wm^{-2}.

The R_n is balanced mostly by H and LE, and to a lesser extent snowpack/ground heat flux (Fig. 3). The residual line represents the amount of energy that would be required to close the observed energy budget (Fig. 3). It includes changes in internal energy of the snowpack and vegetation, but also reflects the common energy balance closure problem in EC-measurements (Mauder et al., 2020) (see Sect. 4.4). The energy balance closure in snow-free conditions was typical for EC-measurements, ranging from 0.81 to 0.99 (Mauder et al., 2020). The lack of snowpack heat flux and/or temperature profile measurements did not enable assessing the closure during snow cover periods. In winter, LE and G are small and the radiative cooling is counterbalanced mostly by downward H, corresponding to warming of the snowpack and/or vegetation, and cooling of the ambient air. The R_n during winter falls lower (more negative) on the northern sites, and thus also downward H becomes stronger (daily average up to 50 Wm^{-2}). In summer, both H and LE are negative (upward~~s~~) heating the atmosphere, while downward G drives the warming of soil profile. At all sites, LE increases along the growing season and peaks approximately in July. In autumn, the turbulent fluxes decrease as response to reduced R_n.

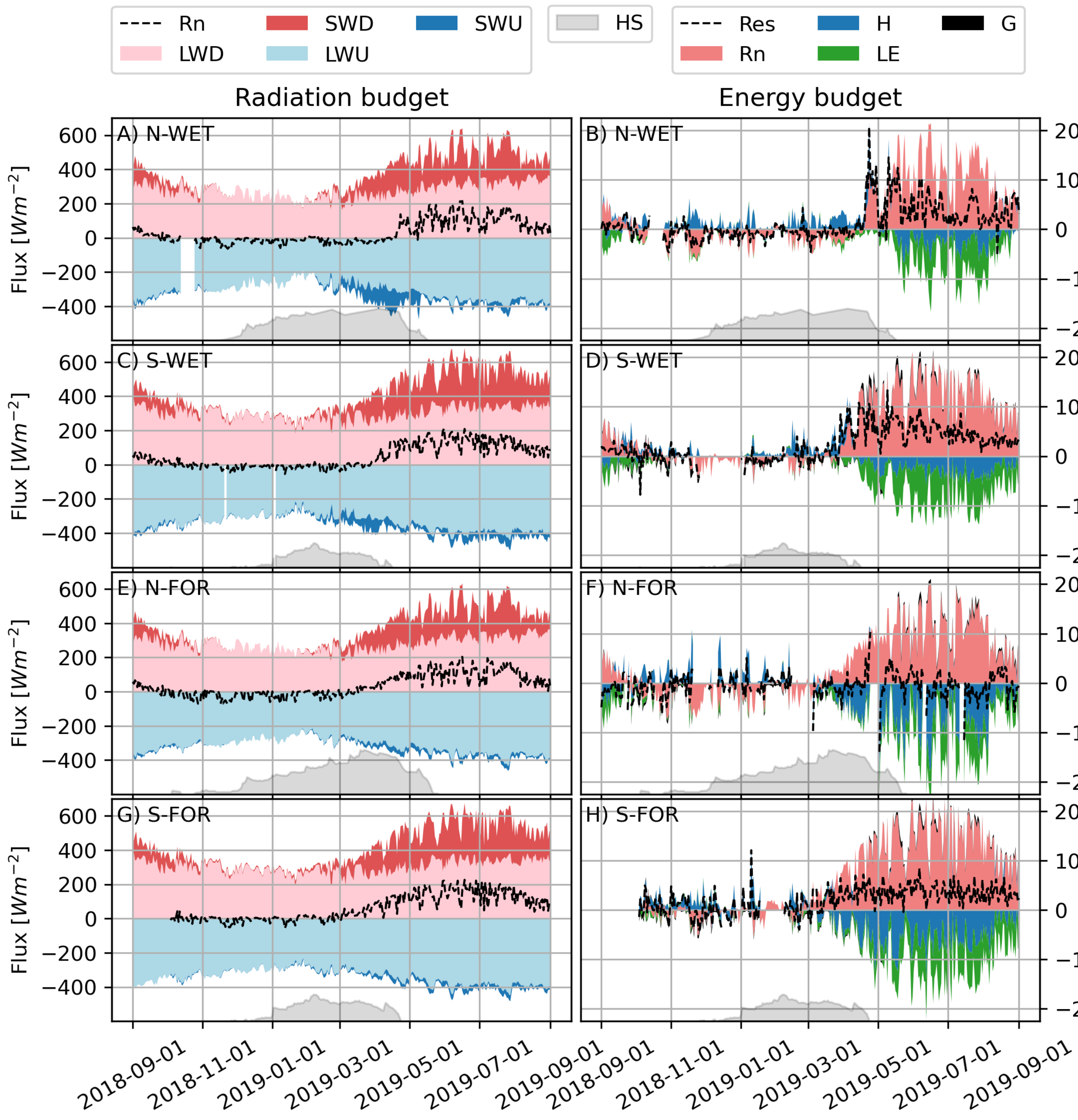

Figure 3. Observed daily averaged radiation budget (left) and surface energy budget (right) of hydrological year of 2018–2019. Colored stacks represent the observed fluxes relative to the surface as shown in legends (i.e. incoming fluxes are positive and outgoing fluxes negative). Dashed line in energy budget plot corresponds to the residual after the sum of each energy component whereas the dashed line in the radiation plot shows the net radiation (R_n). Note different scale in left and right columns. Ground heat flux (G) is missing on N-WET. The observed evolution of the ~~height of snow~~ snow depth (HS) is shown in gray polygon (not in scale).

3.2 Peatland simulations

The sensitivity of surface heat fluxes and ~~height of snow~~ snow depth to different ESCROC-E_2 model parameterizations is shown in Fig. 4. The spread corresponds to the difference between the minimum and maximum of the ensemble. Notably, H has relatively high spread, especially on N-WET, and the observed H often lies near the limit or even outside the simulated range at both S-WET and N-WET. Modelled wintertime LE is low and, as for H, the observed values are near the limit or outside the simulated range, especially in spring. LWU has strong day to day variation well captured by the model, and the spread is rather small relative to the total flux.

3.2.1 Impact of alternative turbulence (C_H) parameterizations

To assess the sensitivity of ~~HS~~ snow depth simulations to alternative turbulence parameterizations, and to alternative snow process options, we examined simulations where the ESCROC-E_2 members are grouped according to their turbulent flux option (Fig. 5). During snow accumulation periods, the spread is small and the groups are consistently overlapping on both sites, indicating that the differences in snow accumulation and maximum snow depth are driven mostly by the uncertainty of snow process descriptions. The spread increases during and after snow melt events, indicating higher importance of turbulent fluxes on snow melt dynamics. While it is difficult to identify a group that fits observed snow depths best, the winter melt event in 2018–2019 on ~~N-FOR~~ N-WET is only captured by the M98 and RI2 parameterizations.

These findings are consistent with the comparison of simulated H and LWU by the two deterministic runs (RIL-SOC and M98-SOC) against observations (Fig. 6). With the RIL-SOC parameterization, the magnitude of H is largely underestimated, while this bias is to most extent corrected by using M98-SOC. Improved simulation of H and surface temperature also entail improved LWU (Fig. 6), but the modelled H fluxes still only moderately correlate (R^2) with observations. In terms of LE, the simulations are not improved by the M98-SOC (see Fig. S1), possibly due to low magnitude and high relative uncertainty of wintertime LE over snow. ~~but as noted earlier, the overall magnitude of LE flux is small. However, regardless of the major improvement, the modeled fluxes still only moderately correlate (R²) with observations.~~

3.2.2 Radiative fluxes

We compare the simulated and observed ~~LSA~~ albedo, SWU, LWU and surface temperatures with snow-free and snow conditions in Fig. 7. These experiments correspond to the deterministic M98-SOC simulation.

The modelled SWU generally matches the observations well, but the scatter increases with increasing SWD, indicating uncertainties in simulated ~~LSA~~ albedo when shortwave forcing is high over the snowpack in spring. These cause a slight underestimation of simulated spring ~~LSA~~ albedo, also visible in the time series especially on N-WET (Fig. 7). Moreover, simulated ~~LSA~~ albedo tends to be overestimated during shallow snow depth both in spring and autumn. This is because the ISBA-FS approach assumes snow to completely cover the ground regardless of the snow depth, while in reality the fractional snow cover can lower the ~~LSA~~ albedo. In contrast in May 2019, the underestimation of ~~LSA~~ albedo in N-WET is due to an

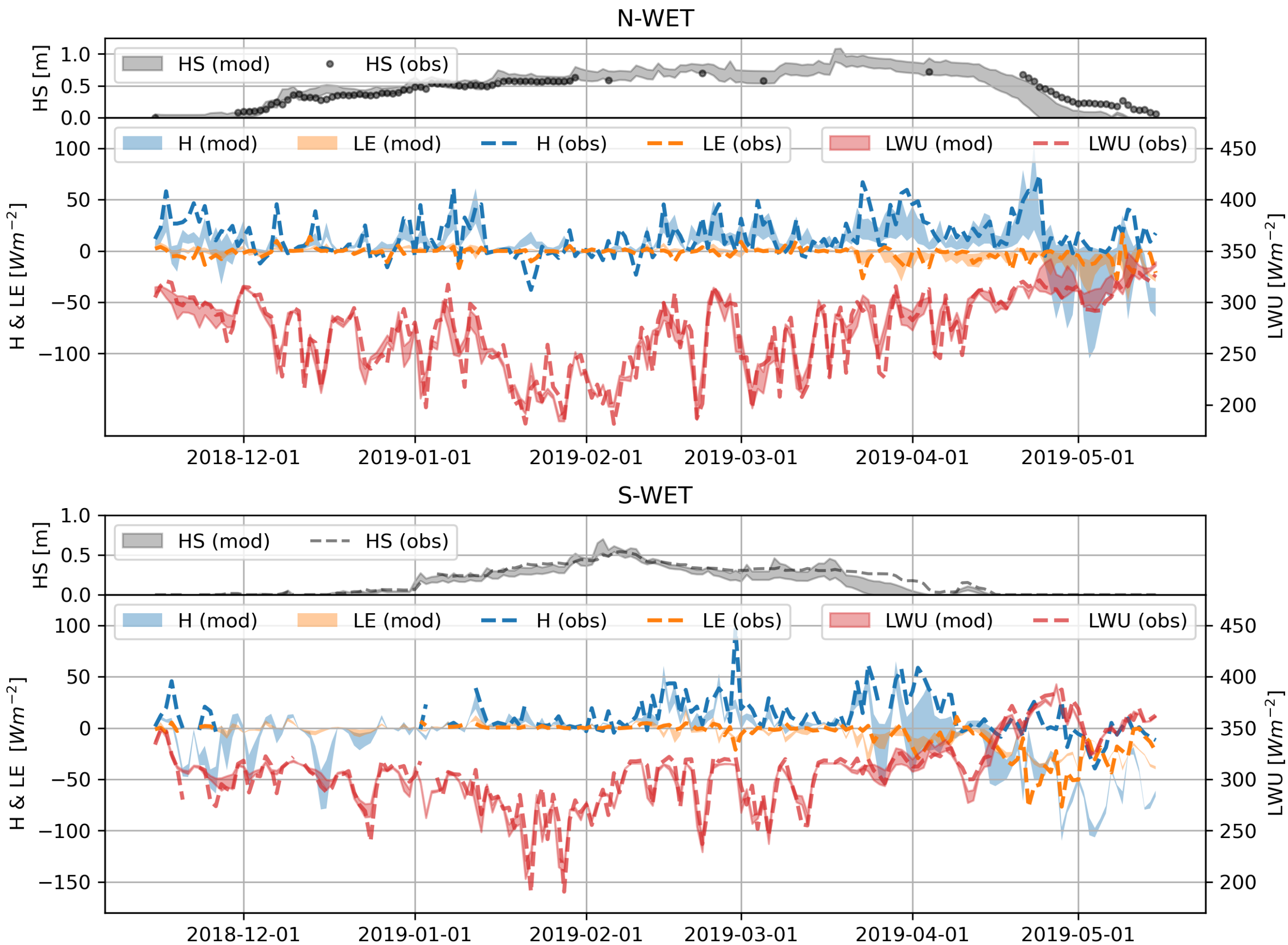

Figure 4. Time series of daily averaged surface heat flux spread simulated by ISBA-FS with ESCROC-E_2 35 ensemble members against corresponding observed values during 2018–2019 snow season. H, LE and LWU correspond to sensible heat, latent heat and upward longwave radiation fluxes, respectively. The observed and simulated evolution of ~~height of snow~~ snow depth (HS) are shown in gray.

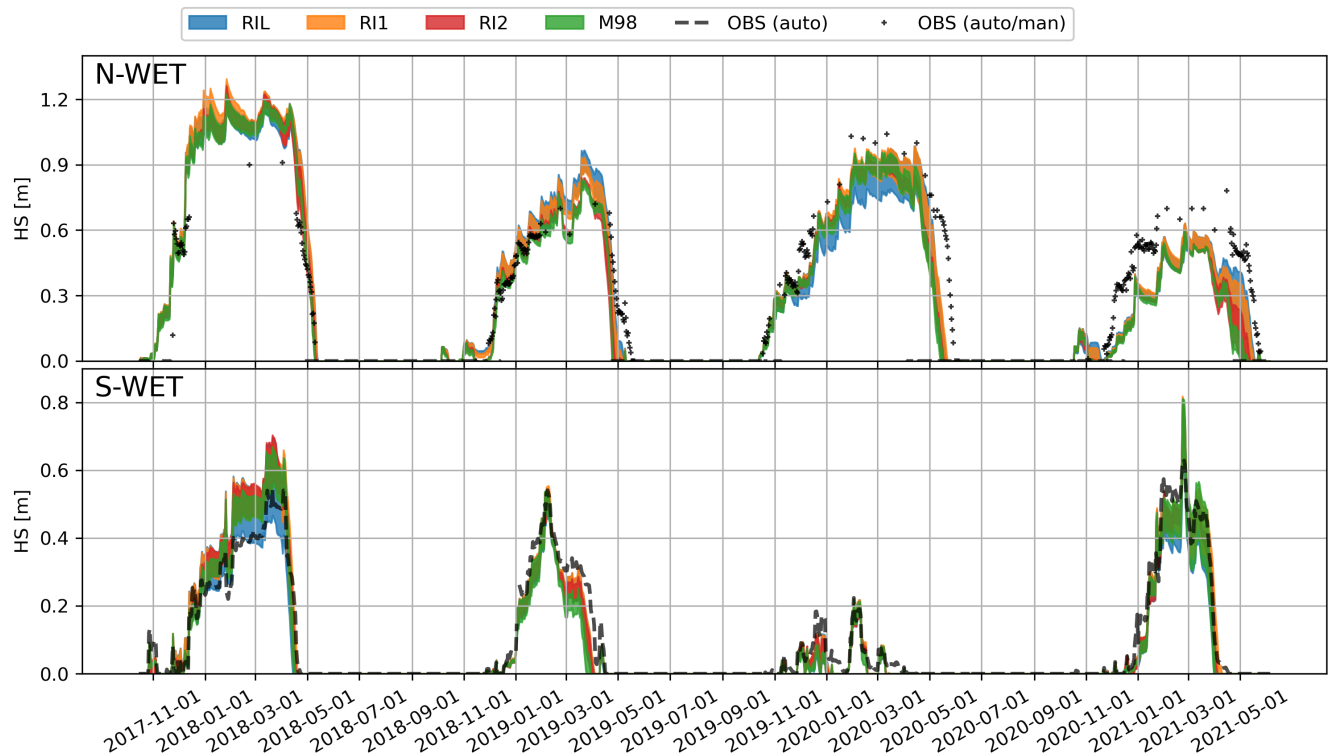

Figure 5. Time series of snow depths simulated by ISBA-FS ESCROC-E_2. The 35 ensemble members are grouped by their turbulent flux parameterization, and the spread of each group is presented in colored ranges. Observed snow depths are presented in black dots and dashed lines.

incorrect timing of snow melt~~ing~~ (too early snow disappearance in the simulation). The mean absolute errors in simulating SWU are small and of similar magnitude (from $\sim$4 to 9 $\mathrm{Wm^{-2}}$) both for snow and snow-free conditions.

Warmer surface temperatures during snow-free season result in higher LWU compared to winter and spring (Fig. 7). The surface temperatures and LWU are generally well simulated across sites and ground conditions at least with the presented time intervals. During snow cover, the upper tail of the radiation distribution is slightly higher than simulated; however the mean biases are generally very low. There are no other visible biases in LWU simulations and the other metrics are also very good, consistent with Fig. 4. The mean absolute errors in simulating LWU are similar for snow and bare ground, about ($\sim$3 to 8 $\mathrm{Wm^{-2}}$).

3.2.3 Soil thermal regime

The effect of soil parameterization on simulated soil temperature dynamics at S-WET is shown in Fig. 8. Due to shallow water table, the soil profile remains nearly saturated throughout the year. As the porosity and field capacity in the M98-SOC parameterization are much higher than in the M98-MIN, the former has also significantly higher heat capacity and

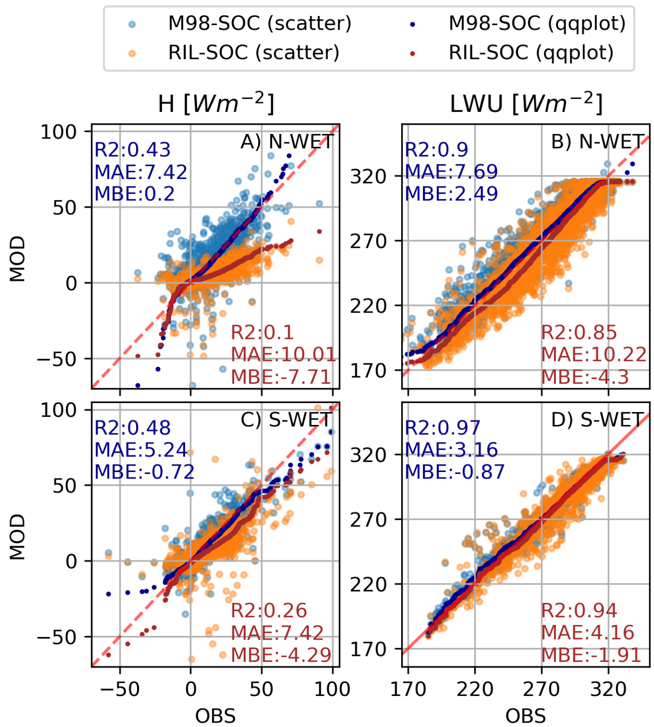

Figure 6. Scatterplots and quantile-quantile plots of sensible heat flux (H) ~~H~~ and ~~LE~~ upward longwave radiation (LWU) ~~LWU~~ during snow cover for the evaluation period with the RIL-SOC and M98-SOC turbulence parameterizations.

smaller thermal diffusivity. This means soil temperature variations are attenuated in M98-SOC compared to M98-MIN, and this attenuation becomes increasingly important in deeper soil layers (Fig. 8). The results show that including a realistic soil profile (SOC) greatly improves the peatland soil temperature simulations at depths 50–70 cm, but only slightly close to the surface (0–10 cm) (see Fig. S2 for comparisons of more soil depths). On both sites, the simulated surface soil temperature variations in summer are greater than observed. This is presumably because ISBA does not include the insulating moss/litter layer on top of the peat soil, as well as due to water table dynamics, potentially affected by lateral flows not accounted for. Due to the weak influence on the surface soil temperatures, the soil parameterization (M98-SOC vs. M98-MIN) does not significantly affect the simulated snow depth (Fig. S2).

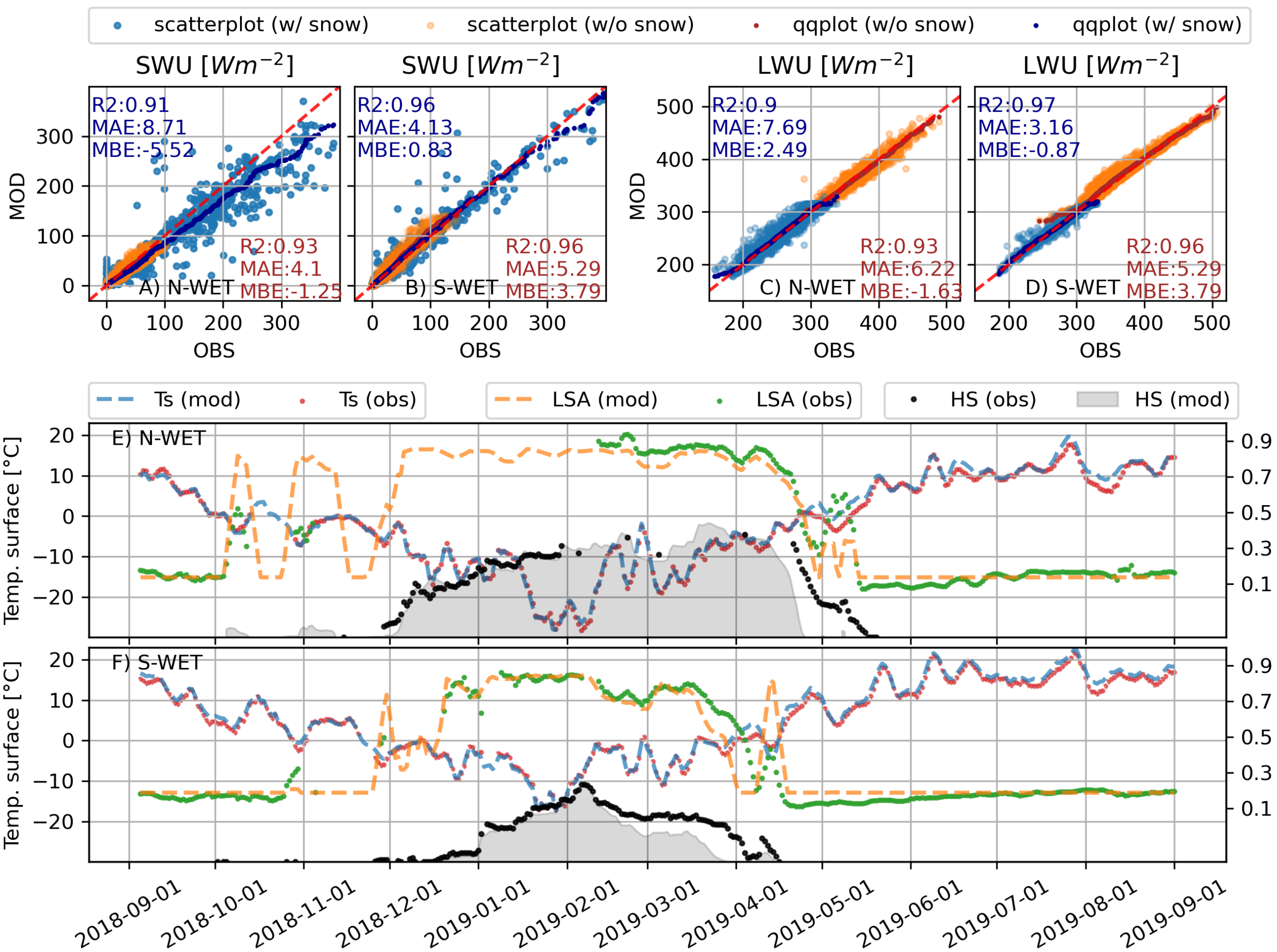

Figure 7. Scatter plots and quantile-quantile plots of modelled against observed upward shortwave radiation (SWU) (A,B) and upwards longwave radiation (LWU) (C,D) on peatland sites with snow cover (w/ snow) and without snow cover (w/o snow) as well as the time evolution of 5-day rolling means of albedos (LSA) and surface temperature (Ts) as simulated and observed from September 2018 to September 2019 (E,F). The evolution of the ~~height of snow~~ snow depth (HS) is not in scale.

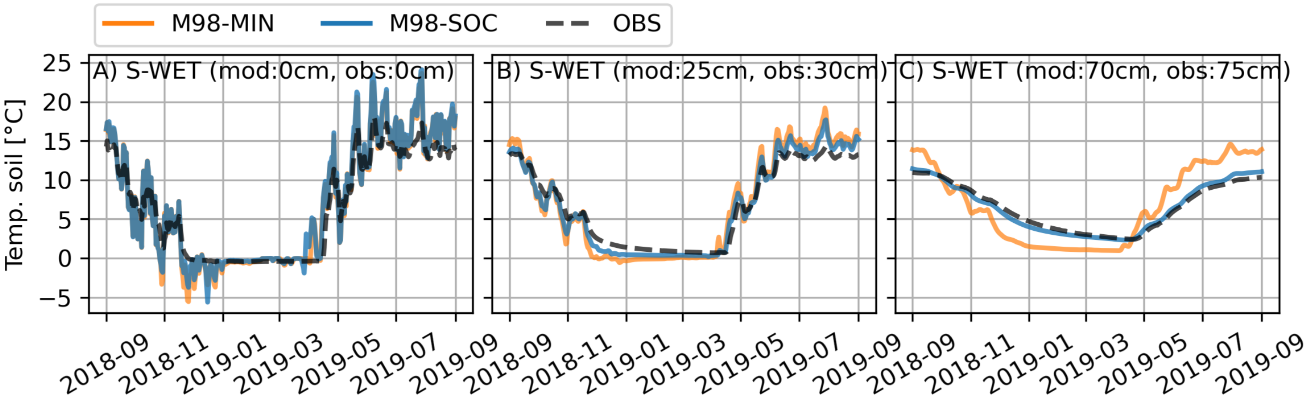

Figure 8. Effect of soil parameterization on ~~simulated and observed~~ soil temperatures (A-C) during a ~~one~~ hydrological year at the S-WET site. M98-MIN refers to mineral soil and M98-SOC to peat soil. The observed soil temperatures are compared to the closest model layer; the depths of measurements and simulations are presented in each panel.

3.3 Forest simulations

3.3.1 Impact of vegetation representation on snow depth

The three different vegetation representations (Sect. 2.3.1) have highly contrasted effect on the forest energy budget, snowpack, and soil temperature simulations. In general, the snowpack simulations for the forest sites are poorer than for the peatland sites; however the observed snow depths also vary considerably within the forests (see OBS in Fig. 9 and Sect. 4.4).

The simulated snow depth with the ISBA-VS (composite soil-vegetation and varying snow cover fraction, Fig. 2B) does not agree with the observations; the model version heavily overestimates accumulation on N-FOR in 2021 and predicts extremely rapid, strong and too early melt events (Fig. 9). Replacing the default snow cover fraction parameter ($w_{sw} = 5$) with $w_{sw} = 0.2$ (used for hydrological modelling in Le Moigne et al., 2020) yields slightly better ~~HS~~ snow depth dynamics for N-FOR, but the results remain unsatisfactory (Fig. C1 in the Appendix). The different sensitivity of w_{sw} parameter for S-FOR and N-FOR simulations is explored via soil temperature simulations in Fig. C2; With the default snow cover fraction parameter, particular warm events on N-FOR heat up the soil causing the snowpack to melt, while simulation with $w_{sw} = 0.2$ manages to retain freezing soil temperatures.

MEB (explicit canopy, ground and snowpack energy balance) simulates the snow accumulation periods at N-FOR very well but peak snow is reached too early and maximum snow depths are underestimated. This is due to combined impact of overestimated compaction and too early start and progression of the snow melt. The role of both processes was evident from comparison of modelled and observed snow water equivalent (see Fig. S6).

ISBA-FS performs better during the snow accumulation period, with simulated snow depths very close to observations. However, the ablation of snow is too rapid, and the final melt out dates are close to those simulated by MEB. On the S-FOR

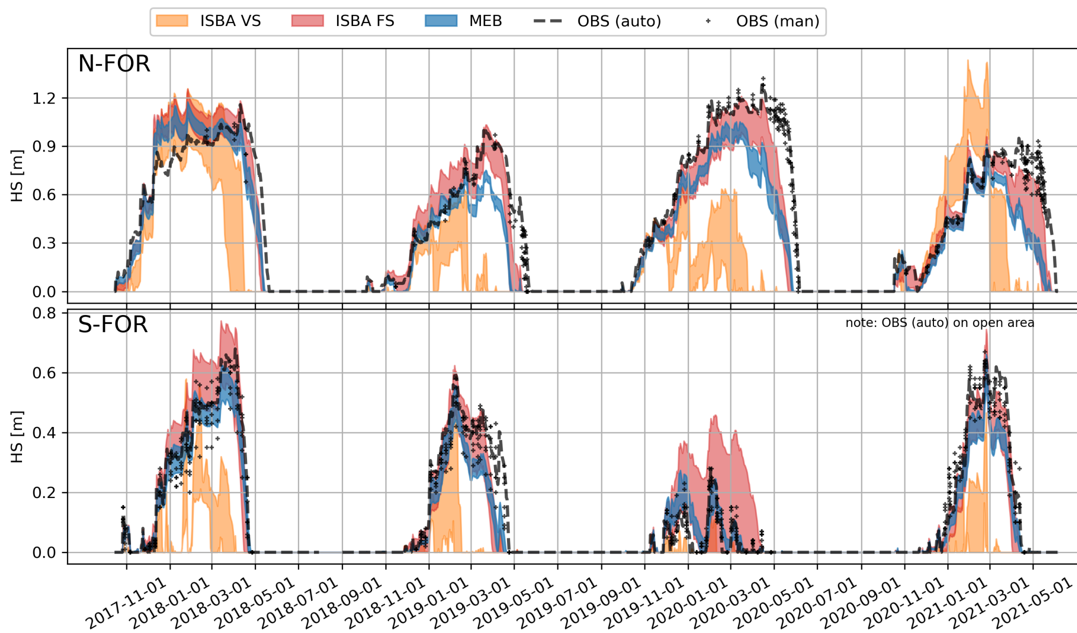

Figure 9. Effect of alternative configurations of ISBA and MEB on the ~~height of snow~~ snow depth (HS). The envelopes visualize the corresponding ensemble spreads between minimum and maximum values.

site, MEB captures both snow accumulation (including peak snow depths), melt dynamics and final melt out dates rather well. ISBA-FS predictions are generally close to MEB. As MEB only considers one option for turbulent exchange (RIL), the spread of the ensemble is smaller than for the ISBA configurations (Fig. 9). The uncertainties of other snow processes accounted for in ESCROC-E_2 are not sufficient to explain the discrepancies between simulated and observed snow depths, suggesting that uncertainties in the canopy process representations prevail in these simulations.

3.3.2 Impact of vegetation representation on soil temperature

Similar to the snow depth, soil temperature predictions by ISBA-VS are erroneous, with drastically underestimated temperatures and unrealistic dynamics (Fig. 10) (see more soil depths in Fig. S3). While MEB and ISBA-FS provided very similar snow depth, the soil temperatures simulated by MEB agree better with the observations although there is a cold bias in autumn and a warm bias in summer (Fig. 10). On N-FOR the warm bias in winter by MEB may be important for determining the soil frost regime. Interestingly, ISBA-FS seems to capture the winter soil temperatures better on N-FOR, but this may be due to the larger cold bias in autumn likely caused by the lack of explicit litter and canopy layers. All model versions tend to overestimate day-to-day temperature variability.

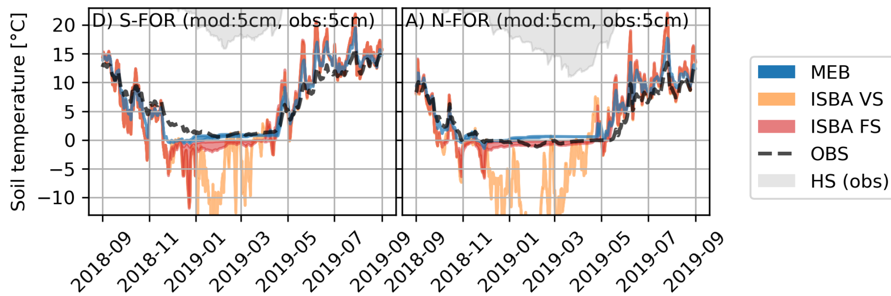

Figure 10. Effect of alternative configurations of ISBA and MEB on soil temperatures. The envelopes visualize the corresponding ensemble spreads. The observed evolution of the ~~height of snow~~ snow depth (HS) is not in scale. The observed soil temperatures are compared to the closest model layer; the depths of measurements and simulations are presented in each panel.

3.3.3 Impact of vegetation representation on surface energy fluxes

Figure 11 compares the deterministic simulations (Sect. 2.5) by MEB and ISBA-VS against observed above-canopy energy fluxes at N-FOR. The snow cover periods are defined according to agreement between MEB simulations and observations, and thus, the ISBA-VS simulations are often snow-free as seen in Fig. 9.

MEB is superior to ISBA-VS in simulating ~~all~~ energy fluxes. SWU simulations with snow cover are clearly improved by MEB, but the spread remains relatively large and ~~LSA~~ albedo is underestimated when incoming radiation is small and overestimated when incoming shortwave radiation is higher. The time evolution of ~~LSA~~ albedo on N-FOR and S-FOR is presented in Sect 3.3.4. The LWU is very well simulated by both model configurations. Turbulent fluxes are clearly better simulated by MEB, but the performance metrics of turbulent fluxes are worse than for radiative fluxes. ISBA-VS uses vegetation fraction parameter to scale the partitioning of latent heat flux between vegetation and soil (Eq. 5 & 6). However, because same roughness length and thus turbulent exchange coefficient (C_H) is used for both soil and vegetation, the soil evaporation and snow sublimation are likely overestimated and result in clearly wrong partitioning between H and LE (Fig. 11C,D and G,H). In the case of N-FOR, especially the summer energy fluxes were majorly improved by simply assigning the vegetation fraction to unity (i.e. full vegetation coverage and no soil evaporation, see Fig. S7).

3.3.4 Evolution of ~~LSA~~ albedo

Figure 12 illustrates the time evolution of modelled and observed ~~LSA~~ albedo and the shortwave components in 2018–2019 on both forest sites. Compared to the measurements, the modelled early and mid winter ~~LSA is~~ albedos are underestimated while the spring ~~LSA~~ albedos ~~is~~ are slightly overestimated, consistent with results in Sect 3.3.3. The likely reason for winter ~~LSA~~ albedo underestimation is because the models do not represent changes in ~~LSA~~ albedo due to intercepted snow. The overestimation in spring is presumably due to representing effective ~~LSA~~ albedo of snow and forest canopy with only bulk

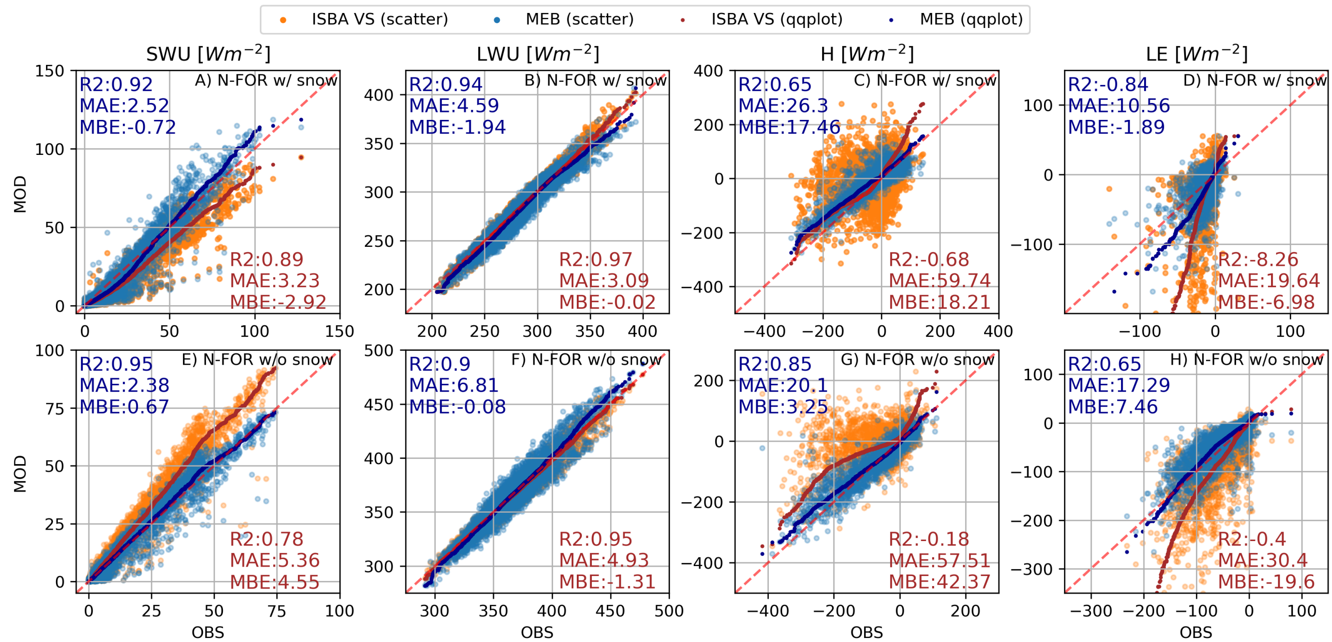

Figure 11. Simulated against observed upward~~s~~ short and longwave radiation (SWU and LWU, columns 1 and 2) and turbulent fluxes (H and LE, columns 3 and 4) on N-FOR site for the full evaluation period. Ground conditions are presented as i) with snow cover (w/ snow, row 1) and ii) without snow cover (w/o snow, row 2).

canopy parameters, as well as effect of spring needle and litter fall decreasing snow albedo. Moreover, the simulated ~~LSA~~ albedo is dominated by the vegetation albedo parameter, and thus, it is not highly sensitive to snowpack albedo dynamics.

3.4 Summary: Surface energy budget on peatland and forest sites

Finally, to sum up the whole surface energy budget, we compare how the simulated R_n and turbulent fluxes (H+LE) match the observations at the four sites. These deterministic simulations are conducted with simulation setups that provided the best fit to data: the deterministic simulation as M98-SOC for the peatland sites, and deterministic MEB simulation for the forest sites (Fig. 13).

Despite the challenges in simulating snow depth evolution at the forest sites, the energy budget simulations are generally better than on the peatlands. Due to the challenges to accurately simulate ~~LSA~~ albedo and surface temperatures on the open sites, the simulated R_n is considerably worse on peatland sites than on forests (see Fig. 7). Especially the high R_n periods, representing the spring conditions, are biased on peatland sites, while the negative R_n (i.e. the winter conditions) are simulated rather well. The challenges in describing forest wintertime ~~LSA~~ albedo and thus SWU (as in Fig. 12 and Fig. 11) do not significantly bias the R_n simulations, as in wintertime the shortwave radiation balance has small role compared to the longwave radiation balance. The results propose that canopy temperature, which particularly in dense forests (e.g. S-FOR) has central role

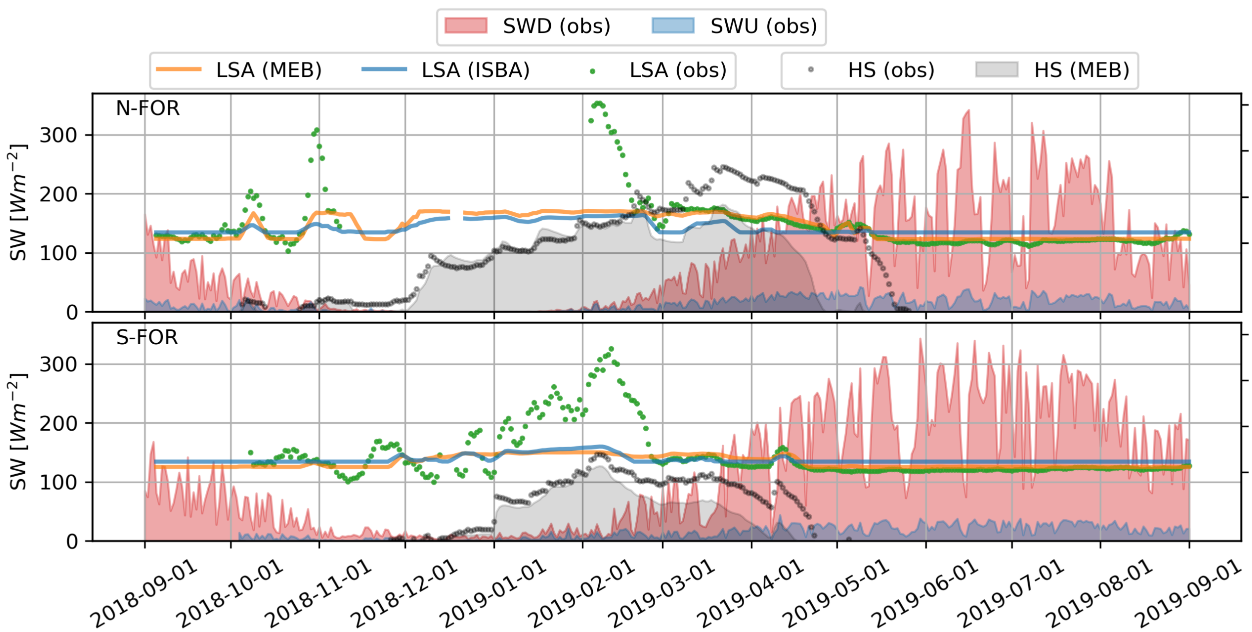

Figure 12. Simulated and observed ~~land surface~~ albedo (LSA) and downward and upward shortwave radiation (SWD and SWU) in 2018–2019. LSA ~~simulations and observations are~~ is presented as 5-day rolling means. The observed evolution of the ~~height of snow~~ snow depth (HS) is not in scale.

for upward longwave radiation, must be adequately simulated by MEB. When it comes to the turbulent fluxes, the simulations capture the main seasonal patterns. However, there are still high uncertainties (scatter) both on peatland and forest sites. The relative uncertainties in simulated and observed energy fluxes are significantly greater in winter than in summer. Performance of the simulated summer energy fluxes is very good (Fig. S4).

4 Discussion

4.1 Insights on energy flux partitioning in boreal environments

~~We used a novel dataset including all surface energy balance components from two peatland and two forest sites. Observations showed that in winter the latent heat flux was minimal at all sites and the negative net radiation was almost completely counterbalanced by the sensible heat flux (see Fig 13 and Fig. 4). The G had only small contribution to the winter energy budget on the studied peatlands, whereas it has been reported to have a rather important role for the open sites in Canadian Arctic (Lackner et al., 2021), Siberia (Langer et al., 2011) and Svalbard (Langer et al., 2011). This is due to the high heat capacity of the peatland, and its large water storage which progressively freezes from the top keeping the temperatures in the soil-snow interface nearly constant at minimum 0°C. Although the winter average daily~~ ~~R_n~~ ~~and H were similar to those~~

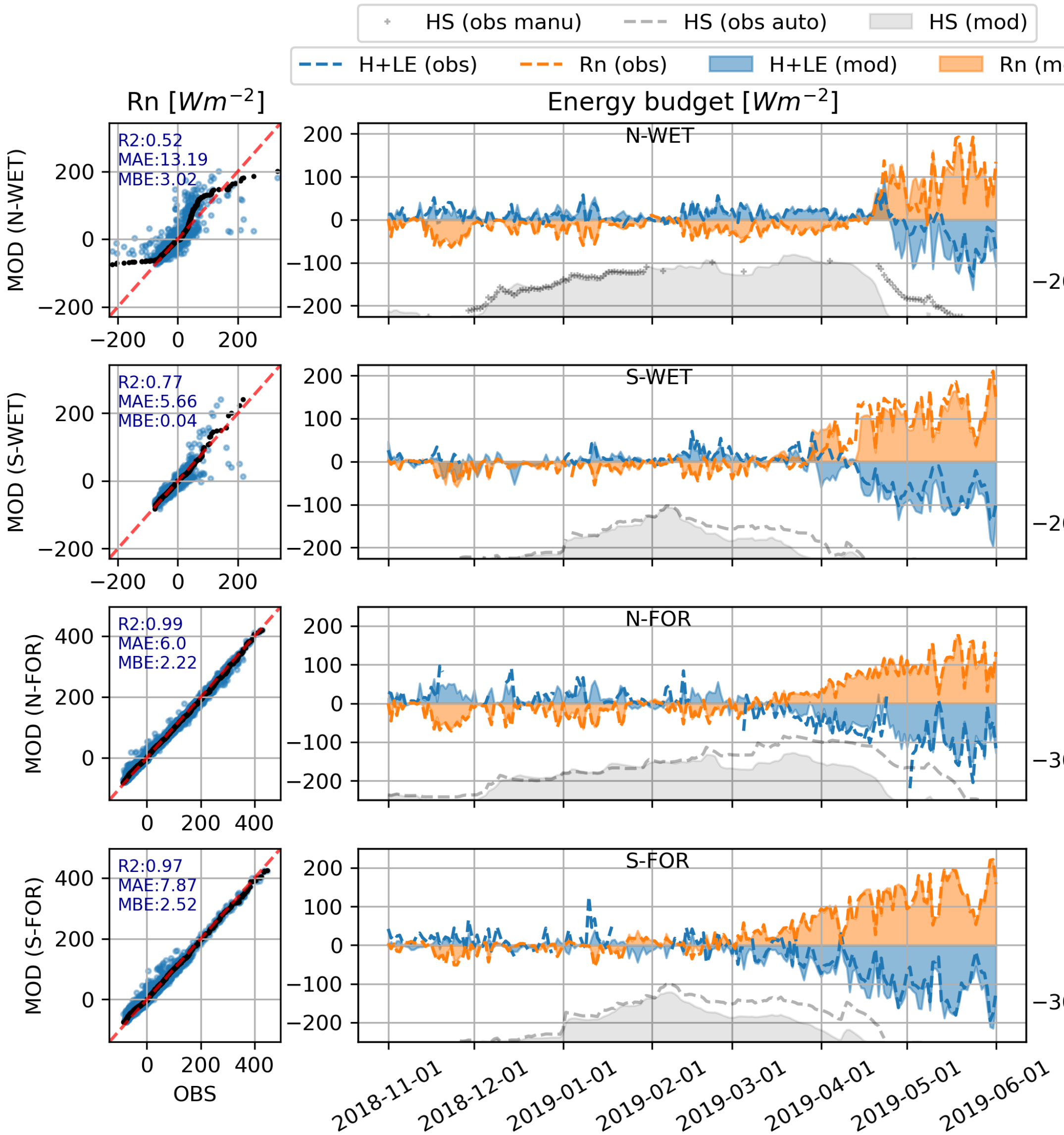

Figure 13. Simulated (MOD) against observed (OBS) daily surface energy budget during winter 2018-2019. The left column shows net radiation (R_n) and right column presents the sum of turbulent fluxes (H+LE). The scatter plots represent full simulation periods when snow cover was present. The observed evolution of the ~~height of snow~~ snow depth (HS) is not in scale.

~~observed in the Canadian Arctic by Lackner et al. (2021), the extremes were considerably larger on the sites studied here. This is most likely due to the more southern location of the study site in Lackner et al. (2021) (56°N), as sites at higher latitudes (Langer et al. (2021) in Siberia 72°N, and Westermann et al. (2009) in Svalbard 78°N) have reported R_n and H extremes closer to those observed in this study.~~

The energy budget observed with this novel dataset over boreal and subarctic peatlands showed that despite the prevailing stable atmospheric conditions (Table 3), the radiative cooling was mostly counterbalanced by sensible heat flux (H) (Fig. 3). During the snow season, the dominating regimes were strongly stable at N-WET (70.2 %) and weakly stable at S-WET (54.6 %). Despite the stronger stability at N-WET, we observed higher H fluxes compared to S-WET, and considerably higher H than Lackner et al. (2021) at the Canadian site dominated by weakly stable conditions. Thus, we presume that greater radiative cooling leads to stronger near-surface air temperature gradient and larger downward sensible heat flux.

The small role of the ground heat flux (G) at the studied peatlands is due to the high heat capacity of the peat soil, and its large water storage which progressively freezes from the top keeping the temperatures in the soil-snow interface nearly constant at minimum 0°C (Fig. 8). Other studies in tundra environments of the Canadian Arctic, Siberia and Svalbard have reported larger contributions of G to the wintertime energy budget (Lackner et al., 2021; Langer et al., 2011; Westermann et al., 2009).

~~In cold regions, periods with stable atmospheric conditions in winter constitute an important part of the annual cycle. Inspired by Lackner et al. (2021), we used the bulk Richardson number to classify atmospheric stability during our study period to unstable ($Ri_b < 0$), weakly stable ($0 \leq Ri_b \geq 0.25$) and strongly stable regimes ($Ri_b > 0.25$) (Table 3). In weakly stable boundary layers, wind shear is sufficient to maintain constant turbulence, while in strongly stable boundary layers, turbulence is intermittent (Steeneveld et al., 2014; Sun et al., 2012). During snow covered season on N-WET, the strongly stable turbulence regime was prevailing (70.2 %) while weakly stable conditions were more rare (16.5 %). On S-WET, the weakly stable conditions were more common (54.6 %). Regardless of the prevailing strong stability regime on N-WET, we observed higher H fluxes on N-WET compared to S-WET, and considerably higher H than Lackner et al. (2021) at the Canadian site dominated by weakly stable conditions. We presume this to be due to greater radiative cooling on our sites, that is counterbalanced by large sensible heat flux even under strongly stable conditions.~~

~~We observed shorter melting period in the peatlands compared to adjacent forest sites (see Fig. S5). This is in line with Lundquist et al., 2013, who established that longer snow retention in forest occurs in colder climates, where the effect of shading (delaying melt) outweighs the impact of longwave radiation enhancement (accelerating melt).~~

We observed shorter snow melt period in the peatlands compared to adjacent forest sites (see Fig. S5), in line with Lundquist et al. (2013), who proposed that increased canopy shading (delaying melt) outweighs the impact of longwave radiation enhancement (accelerating melt). Our datasets support this (Fig. S5); however, the forest sites tended to also accumulate more snow than the peatland sites (wind erosion is presumably higher on peatlands), which may have contributed to longer snow duration in the forest.

~~Below-canopy measurements of surface fluxes from snow-covered forest floor would be required to investigate the actual contribution of individual energy fluxes to snow melt, but only a few efforts have been made to acquire such datasets in boreal~~

Table 3. Occurrence of different turbulence regimes at S-WET and N-WET. The regimes are defined based on the bulk Richardson number (Ri). Unstable conditions as Ri < 0, weakly stable conditions as $0 \leq \text{Ri} \leq 0.25$, and strongly stable conditions as Ri > 0.25.

		Turbulence regimes		
Site	Surface	Unstable [%]	Weakly stable [%]	Stable [%]
N-WET	all	35.1	15.1	49.8
N-WET	snow	13.3	16.5	70.2
N-WET	ground	63.2	15.0	21.8
S-WET	all	59.7	33.7	6.6
S-WET	snow	26.5	54.6	18.8
S-WET	ground	78.9	20.5	0.6

~~forest environments (Mazzotti et al., 2020; Reid et al., 2014), while evaluation of above-canopy fluxes is more common (Napoly et al., 2020; Essery 2009; Rutter et al., 2009).~~

4.2 Implications for simulating snow and energy balance at peatland sites

~~Our results provide insights and recommendations on modelling turbulent fluxes over snow. With the ESCROC-E_2 multiphysics framework, we were able to assess the uncertainties in simulated turbulent fluxes without neglecting the possible contribution from snowpack process descriptions. Our evaluation with multiple years of EC and radiation data of all energy balance components from two subarctic climates allowed deeper analysis of the model performance.~~

~~Our simulations showed large differences in surface heat fluxes between turbulent flux parameterizations, especially on N-WET (Fig. 4), while the fluxes were not impacted as much by alternative snow process parameterizations. The results indicate that modeling turbulent fluxes over snow (i.e. mostly in stable conditions) has major uncertainties, in line with Menard et al. (2021); Conway et al. (2018); Lapo et al. (2019). These uncertainties are larger than in unstable (summer) conditions (Fig. S4), and significantly greater than uncertainty of the radiation balance components. Further, the ESCROC simulations showed that the turbulent exchange parameterizations have noticeable impact on snow melt simulations. These results are in line with simulations at Col de Porte, France and ESM-SnowMIP sites (Menard et al., 2021). In contrast, Lackner et al. (2021) found only small differences between the Crocus turbulence parameterizations in their study in the Canadian Arctic, most likely due to less stable conditions.~~

Our results highlighted the uncertainties in modelling turbulent fluxes over snowpack, and identified the turbulent exchange parameterizations (M98-SOC and RI2-SOC) that improve the simulated surface energy fluxes and snowpack dynamics at high-latitude winter conditions (Fig. 5, 6). In stable (winter) conditions, the uncertainties in turbulent fluxes are in line with Menard et al. (2021); Conway et al. (2018); Lapo et al. (2019), and larger than in unstable (summer) conditions (Fig. S4). Moreover, the turbulent fluxes of sensible and latent heat have greater uncertainty than the radiation balance components (Fig. 13). The ESCROC-E_2 simulations showed that the turbulent exchange parameterizations have also impact on snow melt simulations, in line with simulations at Col de Porte, France and ESM-SnowMIP sites (Menard et al., 2021). In contrast, Lackner et al. (2021) found only small differences between the Crocus turbulence parameterizations in their study in the Canadian Arctic, most likely due to smaller sensible heat fluxes and less frequent strongly stable conditions.

Improved surface temperature simulations by the M98-SOC (absolute biases decreased by 0.3 °C at S-WET and 0.4 °C at N-WET), provide support to Martin and Lejeune (1998) and Gouttevin et al. (2023), who adjusted the turbulent fluxes under stable conditions to reproduce surface and air temperature observations. The default ISBA turbulent flux parameterization (RIL), although widely used e.g. in numerical weather prediction and general circulation models (Mahfouf et al., 1995; Salas-Mélia et al., 2005; Voldoire et al., 2013, 2019), provided the poorest fit with the observed surface heat fluxes, and produced a cold bias in snow surface temperature (-0.4°C at S-WET and -1.1°C at N-WET). This finding is consistent with ESM-SnowMIP (Menard et al., 2021) results, where the default configuration of Crocus had one of the lowest skill for surface temperature simulations (-2°C mean cold bias) among the compared snow models. Even with the M98-SOC simulation we found rather low skill of turbulent flux simulations. To summarize, our findings highlight the limitations of the Monin-Obukhov similarity theory to simulate turbulent fluxes under stable atmospheric conditions, and emphasize the need for further model developments with observations in various environments.

~~On peatlands, the M98 (and RI2) option was superior to RIL option. The improved surface temperature simulations at both sites (absolute biases decreased by 0.3 °C at S-WET and 0.4 °C at N-WET) provide support to Martin and Lejeune (1998) and Gouttevin et al. (2023), who adjusted the turbulent flux simulations under stable conditions to reproduce surface and air temperature observations. The default turbulent flux parameterization (RIL), although widely used e.g. in NWPs and GCMs (Mahfouf et al., 1995; Salas-Melia et al., 2005; Voldoire et al., 2013; Voldoire et al., 2019) , provided the poorest fit with the observed surface heat fluxes, and produced a cold bias in snow surface temperature between -0.4°C (S-WET) and -1.1°C (N-WET). The cold bias produced by RIL is consistent with ESM-SnowMIP (Menard et al., 2021) results, where the default configuration of Crocus had one of the lowest skill for surface temperature simulations (-2°C mean cold bias) among the compared snow models. However, even with the M98 option, we found rather low skill of turbulent flux simulations. Also Lapo et al. (2019) obtained the best simulations by permitting turbulent exchange under stable conditions (with critical stability threshold) when comparing different stability schemes at a site in Colorado. Overall, our findings highlight the limitations of MOST theory, to simulate turbulent fluxes under stable atmospheric conditions, and emphasize the need for further model development and evaluation against observations in various environment.~~

The soil temperature simulations confirmed that it is necessary to realistically describe the organic peat soil hydraulic and thermal properties (Menberu et al., 2021; Morris et al., 2022; Mustamo et al., 2019) to accurately simulate soil thermal regime and consequent freezing/thawing processes in peatlands (Fig. 8) (Dankers et al., 2011; Lawrence and Slater, 2008; Nicolsky et al., 2007). This is line with Decharme et al. (2016) who implemented SOC parameterization in ISBA, and showed improved soil temperature simulations across northern Eurasia. Implementation of water table dynamics and lateral flow could further improve the soil temperature simulations on boreal and subarctic peatlands. The thermal state and ice/liquid water content have also major cascading effects on runoff generation during snow melt (Ala-Aho et al., 2021). Moreover, the interactions between low vegetation and snow are likely improved by using explicit vegetation (MEB in SURFEX). However, as MEB has never been applied on snow-covered low vegetation, additional development and evaluation would have been required that were beyond the scope of this study.

~~ISBA coupled to Crocus is occasionally used for climate and permafrost studies in the Arctic (Gascon et al., 2014; Sauter et al., 2015; Graham et al., 2017; Royer et al., 2021), but evaluations of soil temperature profile simulations of this model system in northern peatlands have not been previously made. Decharme et al. (2016) implemented parameterization of SOC in ISBA, and showed that the performance of ISBA coupled to the ES snow scheme improved significantly the soil temperature simulations across northern Eurasia. Our site-level study with Crocus confirms that adequate representation of peat soils hydraulic and thermal properties (Menberu et al., 2021; Morris et al., 2022; Mustamo et al., 2019) is necessary for accurate simulation of soil thermal regime and consequent freezing/thawing processes (Dankers et al., 2021; Lawrence et al., 2008; Nicolsky et al., 2007). Implementation of water table dynamics and lateral flow could further improve the soil temperature simulations on boreal peatlands. The thermal state and ice/liquid water content have also major cascading effects on runoff generation during snow melt (Ala-aho et al., 2021). Moreover, the interactions between low vegetation and snow would be likely improved by using explicit vegetation (MEB in SURFEX). However, as MEB has never been applied on snow-covered low vegetation, additional developments and evaluation would have been required that were beyond the scope of this study.~~

4.3 Implications for simulations at forest sites

4.3.1 ISBA-VS

~~We showed that turbulent fluxes simulated by ISBA-VS are poorly correlated with the observed ones, consistent with Napoly et al. (2020). We found ISBA-VS to drastically overestimate the LE, likely because of too high simulated soil evaporation, due to its conceptualization of vegetation and snow cover fraction. This is presumably because ISBA-VS uses same turbulent exchange coefficient (C_H) both for computing vegetation evapotranspiration and soil evaporation. At N-FOR, using ISBA-VS with vegetation fraction set to 1 (i.e. omitting soil evaporation) resulted in significantly improved turbulent flux simulations. In winter, the errors might be also linked to an overestimation of the diurnal amplitude of the ground heat flux from the surface fraction not covered by snow. Indeed, the simulated snow cover fractions at our forest sites (Eq. 11-13) never exceeded 0.20, meaning that major part of the soil-vegetation composite always remained in direct contact with the atmosphere without the insulating effect of the snow cover. In terms of LSA, Napoly et al. (2020) found ISBA-VS LSA to depend on forest density: the wintertime LSA of dense forest was overestimated due to overestimation in grid-cell snow covered fraction. Our simulations, in contrast, underestimated the LSA of a sparse forest (N-FOR), which implies a too low snow cover fraction.~~

~~As demonstrated by Napoly et al. (2020), the snow cover fraction approach of ISBA (Fig. 2B) is essentially a compromise that attempts to retain the insulating impact of the snowpack over the soil while still simulating turbulent exchange from the vegetation. We found this compromise to be largely biased towards correctly simulated surface energy fluxes at the expense of poor soil temperature simulations, as a major part of the composite was always directly coupled to the atmosphere. The energy exchange between the atmosphere and the soil-vegetation composite directly impacts the snowpack, and leads to strongly biased snow depth simulations, consistent with Napoly et al. (2020). Overall, ISBA-VS with correct tuning (e.g. setting veg.~~

~~fraction to unity on N-FOR), may be an imperfect but sufficient compromise for forest simulations in applications that foremost require an efficient way to represent grid-cell averaged surface energy fluxes and are not specifically focused on soil or snow cover state. However, the high sensitivity of such empirically based parameters (e.g. veg. fraction) highlights the limitations of ISBA-VS to provide lower boundary conditions of boreal forests for NWP and GCM applications. Also considering the very low skill obtained in snow depth and soil temperatures for this configuration, its use in hydrological applications or surface offline reanalyses (Le Moigne et al., 2020) is highly questionable. Nevertheless, local-scale evaluation might not directly translate to large-scale spatial simulations, as further discussed in Sect. 4.4.~~

We found the ISBA-VS to be largely biased towards correctly simulated surface energy fluxes at the expense of poor soil temperature and snow depth simulations, as a major part of the composite was always directly coupled to the atmosphere (Fig. 9, 10). ISBA-VS with correct tuning (e.g. setting veg. fraction to unity on N-FOR), may be an imperfect but sufficient compromise to simulate snow-free forests in applications that first and foremost require grid-cell averaged surface energy fluxes (i.e. numerical weather prediction). However, we agree with Napoly et al. (2020) that the snow cover fraction approach of ISBA (Fig. 2B) is essentially a compromise that attempts to retain the insulating impact of the snowpack over the soil while still simulating turbulent exchange from the vegetation. The energy exchange between the atmosphere and the soil-vegetation composite directly impacts the snowpack, and led to strongly biased snow depth simulations, similar to Napoly et al. (2020). This fractional approach and high sensitivity of empirically based parameters (e.g. veg. fraction) highlights the uncertainties of ISBA-VS to provide accurate year-around lower boundary conditions of boreal forests for numerical weather prediction and general circulation model applications. Also considering the very low skill obtained in snow depth and soil temperatures for this configuration, its use in hydrological applications or surface offline reanalyses (Le Moigne et al., 2020) is highly questionable. Nevertheless, local-scale evaluation might not directly translate to large-scale spatial simulations, as further discussed in Sect. 4.4.

4.3.2 ISBA-FS

We found that snow and soil simulations in forests were strongly improved when the snow cover fraction was set to unity (ISBA-FS)~~. This adjustment allows~~, allowing the snowpack to fully insulate the soil, similarly to the ~~open~~ simulations at the peatland sites (Fig. 9). These results suggest that if the focus is on snowpack dynamics and soil temperature simulations, ignoring snow-vegetation interactions is a better compromise than using varying snow cover fraction as it is currently implemented in SURFEX (e.g. with only 1 soil column). Consequently, ISBA-FS should be preferred to ISBA-VS in surface reanalyses as in Brun et al. (2013); Vernay et al. (2022). However, ISBA-FS reaches its conceptual limits when forest energy balance, snow, and soil state variables are all of interest. For instance, neglecting snow interception and subsequent canopy snow losses may cause large errors in simulated snow water equivalent in dense forests, and unrealistic contribution of the canopy evapotranspiration may be expected if reanalyses are further used for hydrological modelling. Obviously, highly biased surface energy fluxes would also be expected for any coupling with an atmospheric model.

4.3.3 MEB

~~MEB was developed to solve the aforementioned challenges and reconcile the needs of diverse applications (Boone et al., 2017). It has been previously evaluated on French forest sites, and benchmarked for numerous FLUXNET sites (Napoly et al., 2017). The evaluation of snow-forest interactions has, however, been limited to only three sites in Canada by Napoly et al. (2020) and the ES snow scheme. Our study complements these with two new sites (different vegetation characteristics and climates), and explores MEB performance when it is coupled to the detailed snowpack model Crocus.~~

~~Our results show significant improvements in simulated turbulent fluxes and LSA compared to ISBA-VS. However, we could identify two clear systematic biases in upwards shortwave radiation simulations: LSA was underestimated in winter and overestimated in spring (Fig. 12). The winter LSA was most likely underestimated because intercepted snow increased the LSA, a process assumed negligible in MEB (Napoly et al., 2020). This assumption is based on Pomeroy and Dion (1996), who argued that snow has no significant impact on the canopy albedo or on~~ $\cancel{R_n}$~~. Recently, the increase of LSA by intercepted snow has been shown (Webster and Jonas, 2018), and simple descriptions can already be found in some forest snow models (Mazzotti et al., 2020). Although our results propose the intercepted snow has a clear impact on the LSA, its impact on~~ $\cancel{R_n}$ ~~was weak. Also the wintertime LE simulations performed poorly, suggesting challenges in simulating interception-sublimation processes. The spring LSA bias is in line with Malle et al. (2021), who found LSA at sparse boreal forests to be overestimated by the LSM CLM5. This could be due to simplistic canopy parameterization of MEB. For instance, different tree species with similar LAI and height have considerably different geometries and canopies tend to be heterogeneous. In this case, a bulk 'big-leaf' canopy representation may fail to capture complex effect of canopy shading, particularly at low solar elevation angles typical of high latitudes (Malle et al., 2021).~~

MEB appears as a better compromise than ISBA-VS and ISBA-FS for modelling forest energy exchanges as the snow depth and soil temperature simulations were highly improved compared to ISBA-VS while surface-atmosphere energy fluxes are obviously much more realistic than with ISBA-FS. Significant improvements in energy flux simulations were also obtained compared to ISBA-VS (Fig. 11).

Nevertheless, two systematic biases affecting upward shortwave radiation were identified: albedo was underestimated in winter and overestimated in spring (Fig. 12). The winter albedo underestimation was most likely because intercepted snow increased the observed albedo, a process that is not accounted for in MEB (Napoly et al., 2020). This assumption is based on Pomeroy and Dion (1996), while the increase of forest albedo by intercepted snow has been more recently shown (Webster and Jonas, 2018), and simple descriptions can already be found in some forest snow models (Mazzotti et al., 2020a). Although our results propose that the intercepted snow has a clear impact on the albedo, its impact on R_n was small. The spring albedo bias is in line with Malle et al. (2021), who found albedo at sparse boreal forests to be overestimated by the LSM CLM5. ~~This could be due to the simplistic~~ Big-leaf canopy parameterizations ~~of MEB that~~ may fail to fully capture ~~effect of~~ canopy shading, particularly at low solar elevation angles typical of high latitudes (Malle et al., 2021).

In contrast to Napoly et al. (2020), we found MEB to systematically simulate too early snow melt, especially on N-FOR (Fig. 9). These errors are partly explained by inaccuracies in canopy radiative transfer (albedo biases), but they also suggest errors in simulated below-canopy surface heat fluxes. The differences in snow simulations were rather small between MEB and ISBA-FS especially at N-FOR, suggesting that sparse canopies did not majorly alter simulated snow accumulation and ablation, at least when compared to snow depth observations between the trees. Consistently, Meriö et al. (2023) have shown decreased snow depths at the immediate vicinity of tree trunks, but high snow depth between trees at N-FOR.

~~The snow depth simulations by MEB were highly improved compared to ISBA-VS but slightly worse compared to the ISBA-FS, especially at the sparse forest (N-FOR). This suggests that sparse canopies did not majorly alter simulated snow accumulation and ablation, at least when considering snow depths between the trees. Meriö et al. (2023) demonstrated this at N-FOR with high-resolution UAV snow depth mapping, showing decreased depths at the immediate vicinity of tree trunks, but high snow depth between trees. Although Napoly et al. (2020) found rather good agreement between observed melt out dates and those simulated by MEB, we found MEB to systematically simulate too early snow melt, especially on N-FOR (Fig. 9). These errors are partly explained by inaccuracies in canopy radiative transfer (LSA biases), but they also suggest errors in simulated below-canopy surface heat fluxes; evaluating them would have required complementary observations from below the canopy. Finally, soil temperatures were better simulated with MEB than with ISBA-VS or ISBA-FS, especially at the dense forest (S-FOR). In summary, an explicit representation of vegetation and ground is necessary to simulate accurately both snowpack characteristics and soil temperature, as well as the surface energy fluxes in boreal forests.~~

4.4 Limitations and outlook

Our eddy-covariance fluxes are among the longest datasets ever used for the evaluation of turbulent flux simulations over snow. The EC-data, however, contains both random and systematic uncertainties (e.g. Aubinet et al., 2012). The absolute values of winter H and LE are small and their relative uncertainty is high; compared to summertime measurements the wintertime energy balance closure ratio is typically poorer particularly at the northern ecosystems (Reba et al., 2009; Molotch et al., 2009; Launiainen, 2010). As our analysis used numerous site years from multiple sites, and we used established quality criteria for filtering the EC-fluxes, we expect that uncertainties in flux data do not significantly affect the study results. Moreover, the conclusions regarding the validity of each model version were not affected by selected flux quality flag (Sect. 2.4.2). Intrinsic uncertainties in meteorological forcing are known to exist, especially in northern conditions (instrument freezing, snow blocking, undercatch etc., Stuefer et al., 2020), and data gaps further add up possible sources of errors. Uncertainties in model forcing can affect model-data comparisons, especially during the gap-filled periods (Raleigh et al., 2015).

~~We forced the model simulations with meteorological data from the study sites, and the data gaps were filled with observations from nearby stations and ERA5 reanalysis product. Intrinsic uncertainties in meteorological observations are known to exist, especially in northern conditions.The data gaps further add up possible sources of errors. Uncertainties in model forcing can affect model-data comparisons, especially during the gap-filled periods. Our EC-based fluxes are among the longest datasets ever used for the evaluation of turbulent flux simulations over snow. The EC-data, however, contains both random and systematic uncertainties. The absolute values of winter H and LE are small in northern conditions, and their relative uncertainty is high; compared to summertime measurements the wintertime energy balance closure ratio is typically poorer. As our analysis uses numerous site years from multiple sites, and we used established quality criteria for filtering the EC-fluxes, we expect that uncertainties in flux data do not significantly affect the study results. Moreover, the conclusions regarding the validity of each model version were not affected by selected quality flag.~~

~~Some p~~ Potentially important snow processes on subarctic sites are still absent in Crocus ~~. These include,~~ including wind-induced ~~erosion and accumulation due to~~ snow transport and internal water vapor transfer due to large temperature gradient in the snowpack. Wind~~-induced snow transport~~ can move ~~mass laterally~~ and change the properties of snow (Pomeroy and Essery, 1999; Meriö et al., 2023; Liston and Sturm, 2002), and is especially noticeable on open peatlands. In Crocus, wind modifies the properties of falling snow (Vionnet et al., 2012) but without any lateral transport or modifications of the mass. Although we achieved satisfactory model performance even without accounting for this process, Meriö et al. (2023) showed notable wind transport in transition zones between open peatland and forest at the N-WET site~~, that may alter the total snow mass and the properties of the surface snow layer~~. Although the spatial scale of wind transport prevents an explicit simulation of this process in large-scale LSMs, improved parameterizations of the wind impact ~~of near-surface~~ on snow properties should be considered in the future. Omission of internal water vapor transfer by diffusion and/or convection has been suspected to be responsible for errors in simulated snow properties (density, microstructure) in Arctic snowpack (Barrere et al., 2017; Domine et al., 2018) and consequently in thermal conductivity and soil thermal regime. ~~Nevertheless, a~~ A realistic implementation of water vapor transfer within the snowpack is lacking in most state-of-the-art LSMs. Complementary observations and model developments ~~/evaluations~~ are required to understand if the simulated snow properties are affected by this kind of errors ~~in our study cases~~. Furthermore, the spring and autumn conditions on the peatlands are particularly difficult to correctly simulate; in addition to the snow cover, also e.g. ponding of liquid water and refreezing of the ponds are not uncommon (Noor et al., 2022) and can alter the ~~LSA~~ albedo. ~~These processes are included neither in ISBA nor Crocus.~~

In forests, the spatial heterogeneity of snow cover can be high, as demonstrated by numerous studies (Marttila et al., 2021; Mazzotti et al., 2020b; Noor et al., 2022) and confirmed by our data (Fig. 9). The small-scale forest structure has an important role in the evolution of the snow cover, and may affect the representativeness of point measurements (Bouchard et al., 2022). Consequently, the comparison of point observations and models intended for forest stand and larger scales (such as the big-leaf approach of MEB), can ~~be flawed~~ suffer from this scale miss-match (Essery et al., 2009; Rutter et al., 2009). More realistic below and above canopy heat flux simulation could be achieved by more sophisticated canopy representations, including multiple layers and species (Bonan et al., 2021; McGowan et al., 2017; Launiainen et al., 2015; Gouttevin et al., 2015). For site-level or limited area modelling, high resolution models that explicitly resolve tree-scale canopy structure are a promising alternative to traditional LSMs (Broxton et al., 2015; Mazzotti et al., 2020b).

~~However, only a few attempts have been made to measure the spatiotemporal variability of below canopy energy fluxes, representing the forest floor and understory. In particular, below canopy measurements of turbulent energy exchange are scarce and have to date not been routinely used in snow modeling.~~

~~The forests considered in this study were rather homogeneous and our EC-data represents the footprint average fluxes. Some attempts to capture the spatiotemporal variability of below-canopy energy fluxes, representing the forest floor and understory, have recently been made with distributed measurements or moving platforms, yet these datasets are short-term. In particular, below-canopy measurements of turbulent energy exchange are scarce and have to date not been routinely used in snow modeling. Simultaneous above- and below-canopy measurements may have great potential for snow model evaluations at forest sites. In the absence of energy flux measurements below the canopy, observations of soil temperature and snow conditions~~

~~allowed an indirect assessment of below-canopy energy budget, and highlighted necessary improvements. In the future, more realistic below-canopy and above-canopy heat flux simulation could be achieved by more sophisticated canopy representations, including multiple layers and species. For site-level or limited area modelling, high resolution models that explicitly resolve tree-scale canopy structure are a promising alternative to traditional LSMs.~~

The generality of our findings should be tested by additional snow model and LSM evaluation studies, extended to more contrasting climates and wider range of ~~different~~ ecosystem types. For this purpose, ~~reference~~ model evaluation datasets should be complemented with more boreal and Arctic sites and observations of all components of surface energy balance~~s~~, particularly turbulent fluxes. ~~Such a dataset would facilitate similar experiments with other models.~~

5 Conclusions

We used eddy-covariance based energy flux data, radiation balance and snow depth and soil temperature measurements in two boreal and subarctic peatlands and forests to evaluate turbulent exchange parameterizations and alternative approaches to represent the soil and vegetation continuum in ~~LSMs~~ land surface models. ~~While our model experiments relied on~~ We used the SURFEX platform but our findings are largely transferable to other model systems. Our evaluation with the ensemble snowpack ~~model~~ parameterizations (ESCROC-E_2) ensures that uncertainties in snow processes (not evaluated in this study) do not affect the robustness of our main conclusions.

Peatland simulations showed that using a stability correction function that increases the turbulent exchange under stable atmospheric conditions is imperative to simulate the snowpack energy budget. ~~Although~~ This adjustment led to major improvements under stable conditions during snow cover, but the model performance still remained lower than ~~under~~ in snow-free conditions. Furthermore, correct hydraulic and thermal parameterization of organic peat soils was found necessary to reproduce the observed soil thermal regime in peatlands ~~which implies that inclusion of SOC is a prerequisite for the application of ISBA to peatland environments~~. The findings have direct implications for modelling snow dynamics, peatland hydrology as well as permafrost dynamics.

~~Forest simulations showed that the surface energy budgets~~ The surface energy budgets of forest sites were well simulated by the explicit big-leaf approach (MEB), while the composite soil-vegetation approach (ISBA-VS) performance was satisfactory only after an adjustment of a sensitive vegetation fraction parameter. In particular, shortwave and longwave radiation balances were simulated well by both approaches, whereas the turbulent fluxes had significantly higher uncertainty. Only the explicit vegetation model (MEB) was able to simultaneously simulate realistic surface energy budget and snow/soil conditions ~~while the~~. The composite approaches only succeeded in either simulating the correct surface energy budget (ISBA-VS) or snow/soil conditions (ISBA-FS). Furthermore, we demonstrated that the composite approaches rely on a previously poorly documented parameterization of the snow cover fraction with high sensitivity on model outputs despite a limited physical interpretation.

With well-selected model configuration and parameterization, SURFEX model platform can realistically simulate surface energy fluxes and snow and soil conditions in the subarctic and boreal peatlands and forests. The common version of ISBA (ISBA-VS) can provide rather realistic lower boundary conditions for numerical weather prediction ~~(NWP)~~ and global circulation models ~~(GCMs)~~, ~~in~~ at the expense of non-realistic predictions of forest snow and soil conditions necessary for hydrological applications. We expect that the future inclusion of MEB in operational systems will reconcile these applications. Our results can be used to inform the choice of model configuration for studies of subarctic and boreal regions ecology, hydrology and biogeochemistry under the ongoing environmental change.

Author contributions.

JPN, ML, SL, PA and HM designed the research. JPN led the study and was in charge of the experiments. JPN, GM and ML performed the model evaluations with scientific contributions of SL, HM and PA. JPN was responsible for writing the article, with significant contributions from GM, ML and SL and input from all the authors. BC and MF supported the ensemble simulations and other technical aspects. MA, AL, PK, PA and HM provided surface energy data and ancillary data.

Competing interests.

The authors declare that they have no conflict of interest.

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Code availability.

The SURFEX is an open source project (http://www.umr-cnrm.fr/surfex) but it requires registration. The full procedure and instructions are available at https://opensource.umr-cnrm.fr/projects/snowtools_git/wiki/Procedure_for_new_users. The SURFEX version used in this work is available in git (tagged as *boreal_ecosystems*). The ESCROC version developed for external SURFEX users, is available at https://github.com/bertrandcz/CrocO_toolbox.

Data availability.

Meteorological data, evaluation data and SURFEX specific namelist and forcing files are available as an electronic supplement at https://doi.org/10.5281/zenodo.8252267 under the terms and conditions of the Creative Commons Attribution 4.0 International license

Appendix A: Tables of abbreviations, acronyms and mathematical symbols

Table A1. Table of acronyms and abbreviations.

Acronyms and abbreviations	Definition
LSM	Land surface model
~~NWP~~	~~Numerical weather prediction~~
~~GCM~~	~~Global circulation model~~
LSA	Land surface albedo
LAI	Leaf area index
LWD	Downward~~s~~ longwave radiation
LWU	Upward~~s~~ longwave radiation
SWD	Downward~~s~~ shortwave radiation
SWU	Upward~~s~~ shortwave radiation
H	Sensible heat
LE	Latent heat
R_n	Net radiation
G	Snowpack-ground heat flux
~~MOST~~	~~Monin-Obukhov similarity theory~~
SOC	Soil organic content
EC	Eddy-covariance
HS	~~Height of snow~~ Snow depth
RIL	Classical Louis (1979) formula for the turbulent exchange coefficient
RI1	RIL with threshold at R_i = 0.1
RI2	RIL with threshold at R_i = 0.026
M98	Martin and Lejeune (1998) formula for C_H
BOGR	Boreal grass
BONE	Boreal needleleaf evergreen
FMI	Finnish Meteorological Institute
SMEAR	Station for Measuring Forest Ecosystem–Atmosphere Relations
FST	Flux stationarity

Table A2. Table of mathematical symbols (Part I).

Symbol	Definition
ϵ	surface emissivity
σ	Stefan-Boltzman constant
T_s	surface temperature
ρ_a	air density
ρ_w	liquid water density
ρ	snow density
c_p	specific heat capacity
V_a	wind speed
T_a	air temperature
C_H	turbulent exchange coefficient
ET	total evapotranspiration
E_g	evaporation from the bare soil surface
E_c	evaporation of intercepted water on the canopy
E_{tr}	transpiration from the vegetation
S_i	sublimation from bare soil ice
L_v	latent heat of vaporization
L_f	latent heat of fusion
veg	fraction of vegetation cover
$q_{sat}(T_s)$	saturated specific humidity at the surface
$q_a(T_s)$	atmospheric specific humidity
h_u	dimensionless relative humidity at the ground surface related to the superficial soil moisture content
h_v	dimensionless Halstead coefficient describing the Ec and Etr partitioning between the leaves covered and not covered by interc
z_u	reference height of the wind speed
z_a	reference height of the air temperature and humidity
z_{0t}	roughness height for heat
k	von Karman constant
R_i	bulk Richardson number

Table A3. Table of mathematical symbols (Part II).

Symbol	Definition
C_v	effective heat capacity of the canopy
$C_{g,1}$	effective heat capacity of the ground surface/litter layer (uppermost layer)
$C_{n,1}$	effective heat capacity of the snowpack (base layer)
T_v	temperature of the canopy
$T_{g,1}$	temperature of the ground surface/litter layer (uppermost layer)
$T_{n,1}$	temperature of the snowpack (base layer)
R_{nv}	net radiation of the canopy
R_{ng}	net radiation of the ground surface/litter layer
R_{nn}	net radiation of the snowpack
k	snow effective thermal conductivity
LE_s	latent heat flux of the snowpack
G_n	heat conduction at the snow-soil interface
p_{sn}	effective snow cover fraction
p_{snv}	effective snow cover fraction of the vegetation
p_{sng}	effective snow cover fraction of the soil
HS_g	threshold value for height of the snow
w_{sw}	coefficient relating to vegetation characteristics
E_2	ESCROC optimized standard subensemble
P	precipitation rate
P_{ice}	snowfall rate
P_{liq}	rainfall rate
r	diffuse to total shortwave radiation ratio
μ	cosine of the sun zenith angle
u_*	friction velocity

Appendix B: Model evaluation data availability and radiation forcing evaluation

The availability periods of the model evaluation data are presented in B1.

Table B1. Model evaluation data availability for each site.

Variable	N-WET	S-WET	N-FOR	S-FOR
~~Height of snow~~ Snow depth	2017-11–2021-05	2016-09–2021-07	2013-09–2020-09	2008-09–2021-07
Soil temperature	2013-09–2019-12	2017-06–2021-07	2016-09–2020-09	2008-09–2021-07
Upward LW flux	2017-07–2021-06	2016-09–2021-07	2013-09–2021-07	2008-09–2021-07
Upward SW flux	2017-07–2021-06	2016-09–2021-07	2013-09–2021-07	2008-09–2021-07
Sensible heat flux	2013-09–2021-06	2016-09–2020-12	2013-09–2021-07	2008-09–2021-07
Latent heat flux	2013-09–2021-06	2016-09–2020-12	2013-09–2021-07	2008-09–2021-07
Ground heat flux	2013-09–2017-07	2016-09–2021-07	2013-09–2021-02	2008-09–2021-07

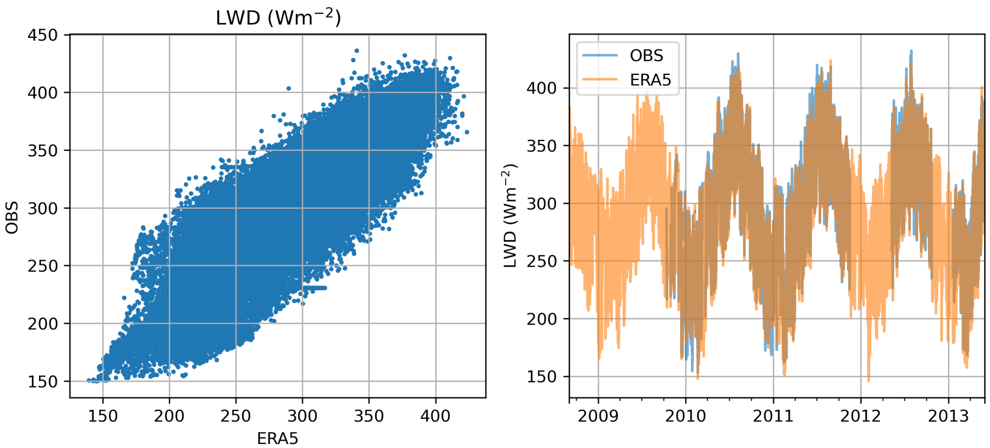

Figure B1. Comparison of longwave radiation forcing data between ERA5 and site observations (OBS) on S-FOR.

Appendix C: Effect of snow cover fraction parameter w_{sw}

Table C1. Summary of snow cover fraction and values of w_{sw} used in different SURFEX/ISBA applications. The parameter w_{sw} is rarely documented, and hence, these application specific values were obtained through communications with the authors.

Application	Name	Domain	Resolution	Snow fraction	w_{sw}	Reference
Numerical weather prediction	AROME, ARPEGE	Europe (many)	1.3 - 10 km	varying	5	Bengtsson et al.
						Courtier et al. (1
Global climate modelling	CNRM-CM6	Global	100 km	varying	2	Decharme et al.
Regional climate modelling	CNRM-AROME	European Alps	2.5 km	varying	1	Caillaud et al. (2
Regional climate modelling	CNRM-ALADIN	Europe, North Africa	12 km	varying	2	Nabat et al. (202
Hydrological modelling	SIM2	France	8 km	varying	0.2	Le Moigne et al.
Regional reanalysis	CERRA-Land	Europe	5.5 km	varying	0.1	Verrelle et al. (20
Snow cover reanalysis	S2M	French Alps	massif-scale	full (1)	-	Vernay et al. (20
Snow cover reanalysis	ERA-Interim-Crocus	Northern Eurasia	80 km	full (1)	-	Brun et al. (2013
Avalanche hazard forecasting	S2M	French Alps	massif-scale	full (1)	-	Morin et al. (202

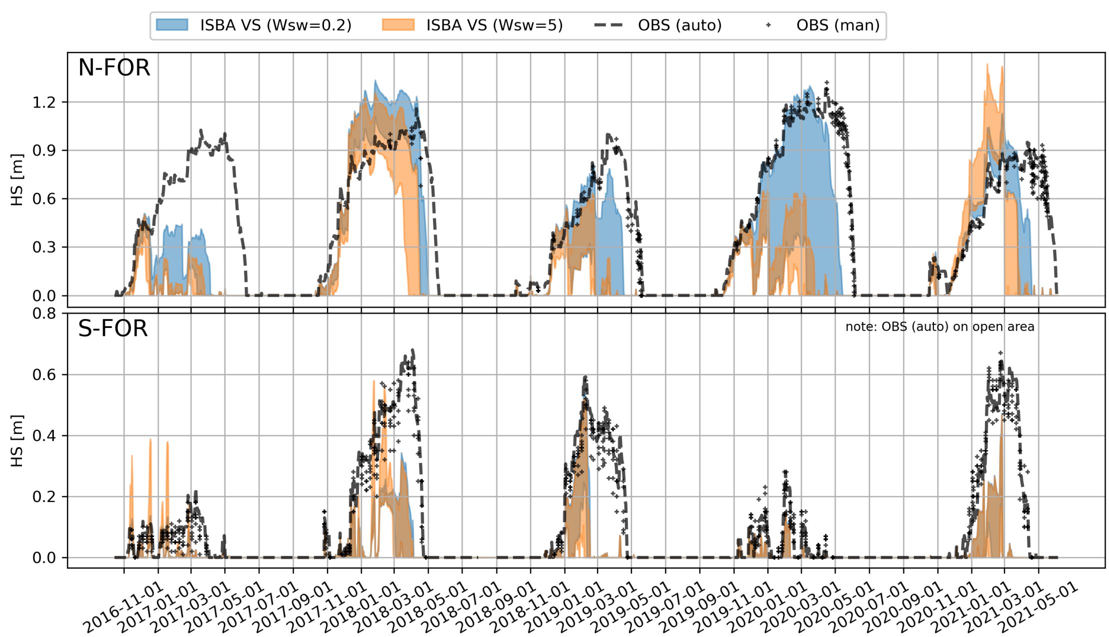

Figure C1. Effect of w_{sw} parameter on snow depths simulated by ISBA-VS. The envelopes visualize the corresponding ensemble spreads between minimum and maximum values.

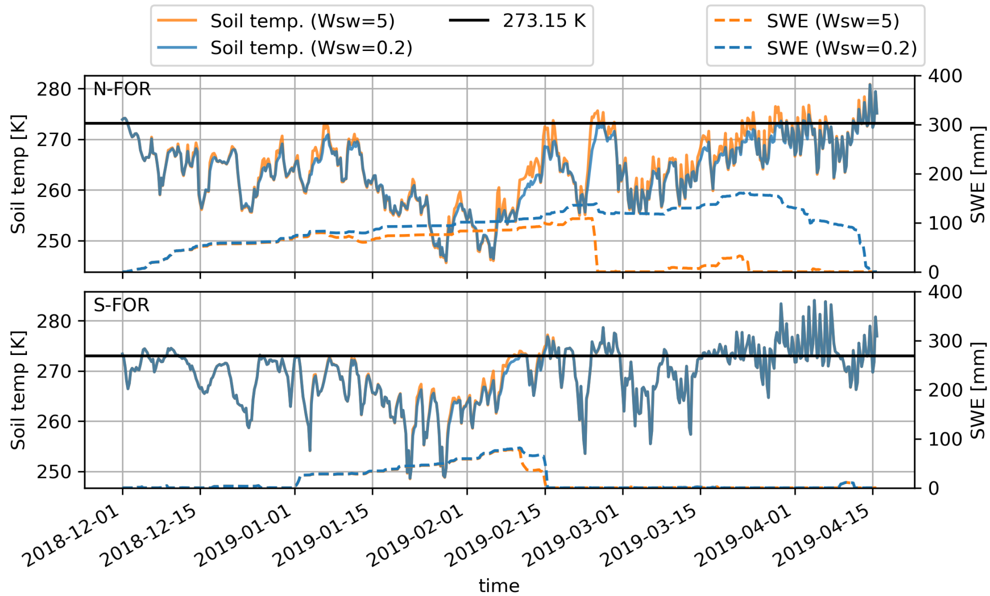

Figure C2. The sensitivity of w_{sw} parameter on 6-hour surface soil temperature simulations and snow water equivalent (SWE). Simulations are represented by one member of the ESCROC-E_2 ISBA VS.

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