

Evaluation of the snow regime in dynamic vegetation models

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Evaluation of the snow regime in dynamic vegetation land surface models using field measurements

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Abstract

An increasing number of studies have demonstrated the significant climatic and ecological changes occurring in the northern latitudes over the past decades. As coupled, earth-system models attempt to describe and simulate the dynamics and complex feed-backs of the Arctic environment, it is important to reduce their uncertainties in short-term predictions by improving the description of both the systems processes and its initial state. This study focuses on snow-related variables and extensively utilizes a historical data set (1966–1996) of field snow measurements acquired across the extend of the Former Soviet Union (FSU) to evaluate a range of simulated snow metrics produced by a variety of land surface models, most of them embedded in IPCC-standard climate models. We reveal model-specific issues in simulating snow dynamics such as magnitude and timings of SWE as well as evolution of snow density. We further employ the field snow measurements alongside novel and model-independent methodologies to extract for the first time (i) a fresh snow density value ($57\text{--}117\text{ kg m}^{-3}$) for the region and (ii) mean monthly snowpack sublimation estimates across a grassland-dominated western (November–February) [9.2, 6.1, 9.15, 15.25] mm and forested eastern sub-sector (November–March) [1.53, 1.52, 3.05, 3.80, 12.20] mm; we subsequently use the retrieved values to assess relevant model outputs. The discussion session consists of two parts. The first describes a sensitivity study where field data of snow depth and snow density are forced directly into the surface heat exchange formulation of a land surface model to evaluate how inaccuracies in simulating snow metrics affect important modeled variables and carbon fluxes such as soil temperature, thaw depth and soil carbon decomposition. The second part showcases how the field data can be assimilated with ready-available optimization techniques to pinpoint model issues and improve their performance.

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1 Introduction

Although data covering the last 125 yr indicate no significant surface–air temperature trends in the Arctic (Polyakov et al., 2002), warming is observed in recent decades (Serreze et al., 2000), and there are many associated indicators of change, such as the expansion of shrub cover (Sturm et al., 2001), decrease in Arctic sea ice extent (Stroeve et al., 2007, 2012), reduction in spring snow cover (Derksen and Brown, 2012) and warming of permafrost (Osterkamp, 2007). Model projections suggest that Arctic surface air temperatures will increase by as much as 0.25–0.75 °C per decade over the next 100 yr (Serreze and Francis, 2006), with associated increases in precipitation (Christensen et al., 2007). Since these regions hold a third of the global terrestrial carbon (McGuire et al., 1995) and half of the global below-ground organic carbon (Tarnocai et al., 2009), most of it locked in permafrost soils, the importance of recording, monitoring and understanding the complex dynamics and feedbacks of the Arctic climate is evident.

Earth system models, either land, ocean, atmospheric or coupled, can offer insights into the multi-level interactions within the system, provided they include and adequately describe the important processes involved. However, significant differences are found in the projections of change made by different models (Friedlingstein et al., 2006). Lack of knowledge of the initial state of the system has been identified as a major cause of uncertainty in decadal projections from climate models (Cox and Stephenson, 2007). This has motivated a major effort to compile datasets of Essential Climate Variables (ECVs) from field and Earth Observational (EO) data, under the guidelines of the Global Climate Observing System (GCOS, 2004, 2010).

This study focuses on land surface models and their ability to simulate snow-related variables and processes, which play a prominent role at high latitudes. Changes in albedo associated with the reduced extent of spring snow cover in the Northern Hemisphere over the past 20 yr has affected the radiative balance and helps explain why the long-term (20th century) increase in surface air temperature over Northern Hemisphere

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land has been greater in spring than in any other season relative to the inter-annual variability (Groisman et al., 1994). Snow is also a hydrological storage pool, collecting precipitation throughout the winter which is then released during the spring, and is essential for local ecosystems and populations. The high latent heat required to melt snow provides a significant cooling effect affecting atmospheric circulation (Vernekar et al., 1995). In addition, the thermal properties of the snowpack make it an efficient insulator, acting as a key control during winter on the heat transfer between the carbonaceous boreal soils and the ambient air, thus affecting permafrost dynamics and soil carbon decomposition. Realistic representation of the snowpack is therefore essential if a land surface model is to provide an accurate and complete formulation of the water cycle and energy flows at high latitudes.

After a description of the datasets and models in Sect. 2, Sect. 3 shows how a range of snow properties (snow water equivalent, snow density and depth, and snowpack sublimation) can be estimated from field data at scales suitable for comparison with land surface models. In Sect. 4, these derived variables are used to assess whether several state-of-the-art land surface models, some of them embedded in IPCC-standard climate models, correctly describe and quantify the snow regime in the Arctic and boreal latitudes. The discussion in Sect. 5 deals with two issues: we firstly investigate how the observed inaccuracies in modeled snow variables affect simulated soil temperatures, and consequently permafrost extent and heterotrophic respiration produced by a land surface model; secondly we describe how field data can readily be used to optimize model parameters in order to improve its simulation of snow dynamics.

2 Description of datasets

2.1 Field data

The field data used in this study is the FSU Hydrological Snow Surveys dataset (Krenke, 2004), which consists of snow measurements carried out from 1966–1996

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in the proximity of 1345 World Meteorological Organisation (WMO) stations spread throughout the FSU (Fig. 1). The number of stations in the data set is reduced from 1345 in 1990 to about 200 in 1991, and the majority of them are located in the European sector of the FSU, with very little coverage over northern Siberia. Each station exhibits different and varying sampling frequency, but measurements are usually taken every 5 or 10 days. Three products are provided. The most detailed is the *synoptic* product which contains the records of transect measurements made in the vicinity of the stations. Several variables were recorded, of which the most important for this study were snow depth, snow density and snow course type (field, forest, gully or unknown) with field and forest types being the most frequent. The *transect* product contains the measurements from the *synoptic* product (with snow water equivalent in place of snow density) that occurred during the 3 most common days of measurements (the 10th, 20th and final day of each month), so these data are standardised to a common time-frame and hence are easier to handle than the synoptic product. Finally, the *station* product contains the mean snow depth records for the first 10 days, the second 10 days and the remaining days of the month at the locations of the weather stations.

2.2 Earth observation data

Globsnow v.1.3 (Luoju et al., 2011) is a monthly snow water equivalent (SWE) dataset covering the period 1979–2010. Its novel algorithm (Pulliainen, 2006) uses forward modeling of brightness temperatures from three satellite-based radiometers (Nimbus-7 SMMR (1979–1996), DMSP SSM/I (1987–2002) and Aqua AMSR-E (for 2002–2010)) and assimilates field data on snow, some of which includes field data used in this study, to produce a global SWE product. Its accuracy has been demonstrated, and it has been used to estimate trends of SWE over the last 30 yr (Takala et al., 2011).

The LEGOS SWE retrievals are based on a dynamic algorithm (Mognard and Josberger, 2002) that calculates the snow depth from the spectral gradient of SSM/I brightness temperatures with additional inputs of air-snow interface temperatures from the National Centres for Environmental Prediction (NCEP) global reanalysis (Kalnay

et al., 1996) and modelled snow-ground interface temperatures and snow density derived from the Interactions between Soil–Biosphere–Atmosphere (ISBA) model (Boone et al., 2006).

2.3 Land surface models

Four state-of-the-art land surface models were evaluated, some of which have been coupled to General Circulation Models used to investigate the land–atmosphere–ocean carbon exchange. CLM4CN (Lawrence et al., 2011) is an updated version of CLM4 with a Carbon-Nitrogen biochemical model and constitutes the land component in the Community Earth System Model (CESM) (Collins et al., 2006). CLM4CN incorporates the Snow and Ice Aerosol Radiation (SNICAR) model (Flanner et al., 2007) to determine snow albedo while the properties of the snowpack, such as density, evolve following metamorphic processes based on a microphysical model (Flanner and Zender, 2006). CLM4CN includes boreal Plant Functional Types (PFTs) and has specific parameterizations for the thermal and hydraulic properties of organic soil (Lawrence and Slater, 2008); its soil temperature profiles and permafrost distribution compare favorably to observations (Lawrence et al., 2008).

JULES (Best et al., 2011) is the land surface component used in the Hadley Centre climate models and is based on the Met Office Surface Exchange Scheme (MOSES) (Essery and Clark, 2003). JULES does not include important parameterizations for northern latitudes, such as organic soils or endemic PFTs, but contains a complex snowpack sub-model with varying snow density, sublimation and snow interception by the canopy (Best, 2009); these processes are also present in CLM4CN.

LPJ-WM (Wania et al., 2009) is a version of LPJ tailored to northern latitudes by including boreal-specific PFTs and incorporating peatland hydrology, organic soils and permafrost dynamics. The original hydrology (Gerten et al., 2004) was modified to allow more soil layers, while permafrost extent and active layer depth are obtained by modeling the soil temperature profile as a function of depth using a one-dimensional energy flow formulation. However, its representation of snow properties is very simple,

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with snow density being held constant for the first three quarters of the snow season and increased as a linear function of time during the last quarter, following findings by Oelke et al. (2003); Ling and Zhang (2004). It does not include canopy interception, sublimation or any of the snow metamorphic processes found in CLM4CN and JULES.

The Sheffield Dynamic Global Vegetation Model (SDGVM) (Woodward and Lomas, 2004; Woodward et al., 1995) has been extensively used in DVM comparison studies (Cramer et al., 2001; Le Quere et al., 2009; Piao et al., 2009), and produces magnitudes and trends of carbon fluxes typical of those from other DVMs at global and regional scales. Nevertheless, it lacks the complexity and many of the processes found in both CLM4CN and JULES. It does not consider the thermal properties of soil, a key drawback when considering permafrost soils in boreal latitudes, has no specific adaptations for northern latitudes and does not simulate snow interception by the canopy. Nevertheless, it includes sublimation, thus providing a more realistic and complete water balance than LPJ-WM. It is also an in-house model that can be readily modified, and thus is particularly valuable in the optimization of sublimation discussed in Sect. 5.2.

Spatial resolution varies between the datasets and models. GlobSnow and LEGOS SWE data are provided in 1° grid-cells, while the grid-cell spacing of the models is defined by their climate drivers. LPJ-WM is driven by the CRU TS 3.0 (Climate Research Unit Time-Series) (Mitchell and Jones, 2005) with 0.5° grid-cells, and JULES by WATCH (Water & Global Change) (Harding and Wernaars, 2011), also at 0.5° . SDGVM is driven by CRU TS 3.0, but at 1° grid-spacing, while CLM4CN uses the CRU + NCEP climatology, based on CRU 2.0 and the NCEP reanalysis (Kanamitsu et al., 2002) at $0.9375^\circ \times 1.25^\circ$. All model outputs are given as averaged monthly values.

3 Methods

In this section we describe the methods used to derive five snow variables from the transect data:

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1. Snow Water Equivalent
2. Fresh snow density
3. Snow density evolution (monthly and daily time-steps)
4. Snow depth
5. Snowpack sublimation

The details of how these variables are estimated are in part motivated by the need to compare them with values calculated in models, as discussed in Sect. 4.

3.1 Snow water equivalent

Monthly averaged SWE outputs were acquired from the low resolution *transect* product. Usually, and depending on the spatial resolution of the dataset, more than one station falls within the spatial extent of each grid-cell. Since the cover type for each station was provided, this allowed the SWE for forest and non-forest areas to be estimated. The SWE for the whole grid-cell was then calculated as the weighted average of these two values, where the weights were given by the fractions of the grid-cell covered by forest and non-forest. These fractions were derived from the MODIS Vegetation Continuous Fields (MODIS VCF) MOD44B land cover product (Hansen et al., 2003). (A similar approach was applied for the other snow variables described in the following sub-sections.)

3.2 Snow density

The thermal conductivity of the snowpack is proportional to snow density (Saito et al., 2012; Sturm, 1992; Sturm et al., 1997), making this variable important for the energy balance of the soil column. Since its conductivity tends to be small, the snowpack acts as a thermal insulator, and reduces the magnitude of fluctuations in temperature and

propagating heat from the air to the soil. Snow density also has a small effect on the radiation balance because of its effects on albedo (Bohren and Beschta, 1979), but the grain size is more important (Perovich, 2007). Estimating snow density requires both the initial value for fresh snow (Sect. 3.2.1) and a procedure to track its development up to melting (Sect. 3.2.2).

3.2.1 Fresh snow density

The density of fresh snow provides the starting point for snow evolution, so is an important property of the snowpack. In order to estimate fresh snow density from the field records, we scanned the entire *synoptic* dataset, approximately 600 000 measurements, for pairs of measurements that were taken less than N days apart, in which the earlier record showed no snow but the latter had positive values of snow depth. It was then assumed that the age of the snow would be uniformly distributed on the interval $[0, N]$ with an expected value of $N/2$ and so the latter record represented the value of fresh snow density $N/2$ days after the initial snowfall. This is less likely to be correct as N increases because snow undergoes densification immediately following its deposition, while newly deposited snow can dilute the snowpack and decrease its overall density. To reduce the effect of such perturbations, only measurements acquired up to $N = 3$ days apart were considered. This was possible because, although most of the records in the synoptic dataset were collected with a 5 or 10 day step, there were 60 pairs of measurements at the beginning of the snow season taken less than 4 days apart in which the earlier one had snow depth equal to zero and the later a positive value.

3.2.2 Snow density evolution (monthly time-step)

After deposition, overburden pressure and prevailing weather conditions, such as temperature and wind, cause continual evolution of the snowpack density (Anderson, 1976; Liston and Elder, 2006), with a tendency of increasing over the winter period, reaching

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a maximum before or during melt. Using the same approach as for SWE, monthly values of density were estimated using the *transect* product as a weighted average of values for forest and non-forest areas.

3.3 Snow depth and density (daily time-step)

5 Daily time series for both snow depth and density had to be created from the data records for exploitation by LPJ-WM in the permafrost analysis in Sect. 5.1. Only stations overlying organic soils which had nearly complete temporal coverage with few data gaps were selected, in order to ensure a more accurate construction of daily time series. This led to a dataset containing 9 stations, all located either in south-central
10 Eurasia, where the permafrost boundary lies, or Eastern Eurasia, where low air temperatures contribute to predominantly continuous permafrost; there were none in the western sector where permafrost is generally absent.

The *station* dataset contains records of mean snow depth at the location of each station on the 10th, 20th and last day of each month, while the *transect* dataset contains records of snow depth and snow density on the same days, but acquired along transects in the vicinity of the station. In the former, snow depth was assigned a zero value when snow was absent, so no gaps occur, even during the summer and spring. In contrast, the transect measurements took place only when snow was present; in its
15 absence, the records contain no data, limiting our ability to detect the start and end dates of the snow season. Hence we utilized both sets, acquiring snow depth from the *station* records and snow density from the *transect* records, after first ensuring that the snow depth for each station from the two datasets was well correlated ($\rho \geq 0.9$) and had no offset.

To form daily snow depth fields between the 10 day interval of the *station* records, an
25 increase between two dates was attributed entirely to snowfall on the second date (step function), while decreases were calculated using linear interpolation between observed values. Snow density was linearly interpolated both when increasing and decreasing, but a lower limit of 100 kg m^{-3} was imposed during the early part of the snow season,

and at the end of the season it was assigned the value of the last available record before snow depth reached zero.

3.4 Snowpack sublimation

Estimation of snowpack dynamics needs to include both source and ablation processes, where the latter describes all processes that remove snow. While the source terms can be estimated from climate (precipitation) data, there are inherent difficulties in estimating snow ablation and distinguishing amongst its different elements and particular sublimation (Lundberg and Halldin, 2001). The available in situ sublimation data are sparse and confined in spatial extent, with the available information being restricted to point measurements, usually obtained by a variety of weighting techniques, on-site modeling (Hood et al., 1999), flux towers (Harding and Pomeroy, 1996) and pan methods (Zhang et al., 2005), or any combination of the above (Gelfan et al., 2004). Hence we developed a novel approach that exploits the high resolution *transect* records to estimate snowpack sublimation across regions of Eurasia.

The snow mass balance equation can be written as

$$P - S_s - S_c - M - T - T_s = 0 \quad (1)$$

where P is precipitation (snowfall), S_s and S_c are snowpack and canopy sublimation respectively, M is melt, T is wind transport of snow (blown snow) and T_s is sublimation of the blowing snow. The magnitudes of the blowing snow terms (T and T_s) are much smaller than sublimation for the continental parts of northern latitudes (Dery and Yau, 2002; Yang et al., 2010), hence are neglected. Even though locally and in finer scales, e.g. coastal or high altitude regions, those two terms can significantly alter the snow mass balance due to prevailing winds, for an eastern and western sector examined in this study (Fig. 1), modeling results by Dery and Yau (2002) showed that blowing snow sublimation and snow divergence (blown snow) are negligible; nevertheless it should be mentioned that the modeling approach by Yang et. al. (2010) produced slightly higher values for those terms. For overall mass balance, canopy sublimation is important since

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differences in meteorological conditions and incoming radiation fluxes cause its magnitude to exceed that of the snowpack, especially late in the snow season (Molotch et al., 2007; Lundberg and Halldin, 2001). However, as the available field data only contain information about the snowpack, we focus only on the snowpack sublimation term, S_s .

Let SWE_A and SWE_B , with $SWE_A \geq SWE_B$, be consecutive non-increasing snow water equivalent measurements from the *synoptic* transect dataset around a particular station on Julian days t_A and t_B , with $t_B - t_A \leq 10$. An estimate of the monthly sublimation in month m is then provided by

$$S_s = -d(m) \frac{SWE_A - SWE_B}{t_B - t_A} \quad (2)$$

where $d(m)$ is the number of days in month m . Whenever a pair of records meets the above conditions, a value for S_s is obtained. For each station, the representative monthly value for S_s is then taken as the median value of all these estimates over the months and years for which data are available. (Since the resulting probability distributions are approximately non-negative exponential, the median is preferable to the mean as it is a better measure of tendency.) Two important conditions were enforced to remove seasonal and episodic melting and ensure that the remaining ablation process responsible for non-positive changes in SWE is snowpack sublimation:

1. Only months with no seasonal melting were included. Seasonal melting varies geographically, with earlier melt and shorter snow seasons in western than eastern Eurasia, so we used estimates of the dates of snow cover appearance and disappearance derived from the Special Sensor Microwave/Imager (SSM/I) (Kouraev and Mognard, 2010).
2. Anomalous high values of estimated sublimation caused by episodic melting were excluded by only using S_s values up to the 80th percentile of the distribution in each month.

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4 Results

In this Section we display the values derived by the methods set out in Sect. 3 and compare them with corresponding values calculated by the models. Different formulations of snow processes are used by each model, and this is reflected in the models discussed in each subsection.

4.1 Snow Water Equivalent Comparisons

Snowpack SWE is calculated by all four models and is also given by the Globsnow and LEGOS EO products. In order to compare these six datasets with the SWE values derived from the transect data, for every grid-cell we defined vectors $\mathbf{A}_k$, $k = 1-6$, and $\mathbf{B}$, whose elements are the monthly values of SWE for the six datasets and transect measurements respectively. These were used to define two statistics:

1. ρ_k , the correlation coefficient between the time series for dataset k and the transect data, i.e., between $\mathbf{A}_k$ and $\mathbf{B}$;

2.
$$\Delta_k = (\overline{\mathbf{A}_k} - \overline{\mathbf{B}}) / \overline{\mathbf{B}}, \quad (3)$$

which quantifies the bias in dataset k relative to the observed mean SWE, $\overline{\mathbf{B}}$; here the overbar denotes average.

All comparisons between models and field records were carried out for the full available temporal range of the transect data (usually 1966–1990 or 1966–1996), while for the EO products the range was restricted to their current availability: GlobSnow from 1980 to 1996 and LEGOS from 1988 to 1996 but only for latitudes north of 50° N.

Comparisons between the *transect* product and the datasets are presented in Fig. 2 for each dataset, with the correlation coefficient ρ in the left column and Δ , the relative bias, on the right. The months May to August were excluded and grid-cells with

less than 10 pairs available for correlation are not shown. Globsnow achieved the best agreement with the transect data, with over 78 % of the grid-cells having a correlation coefficient exceeding 0.6; this is expected since the Eurasian data were used in the calibration of Globsnow. However, a region of low correlation occurs around the Caucasus, probably because topography causes problems in the retrieval. In contrast, only 56 % of the grid-cells had $\rho \geq 0.6$ in the LEGOS EO-based SWE product for which data were available only for latitudes north of 50° N. As has been demonstrated (Foster et al., 1997; Pulliainen, 2006; Takala et al., 2011), temporal and spatial biases in radiometry mean that a single algorithm, like that used at LEGOS, cannot provide accurate and global SWE estimates without assimilating ground data or using forward modeling, as in Globsnow. The similarities between the LPJ-WM and SDGVM correlation maps can be attributed to their use of the same climate drivers. For these two models, 72 % and 65 % respectively of grid-cells have $\rho \geq 0.6$, and both exhibit lower correlation in the south-west part of the European sector, as well as in the region east of Lake Baikal. Approximately 66 % of the CLM4CN grid-cells have $\rho \geq 0.6$, with poor correlation in eastern Siberia, while for JULES 74 % of the grid-cells have $\rho \geq 0.6$.

The relative bias, Δ , is shown in Fig. 2 (right) only for grid-cells with $\rho \geq 0.6$. GlobSnow again agrees well with the data, except for a small underestimate in the western sector. Fewer grid-cells are marked in the LEGOS dataset because of the poorer correlation, and underestimation occurs throughout Eurasia. LPJ-WM, SDGVM and CLM4CN all tend to overestimate SWE in the European sector but perform well in the other regions, both in terms of ρ and Δ . In contrast, JULES significantly and systematically underestimates SWE throughout Eurasia, in many regions by more than 50 %, despite the high values of the correlation coefficients.

Since all winter months were considered in estimating ρ , it is more a measure of seasonal than inter-annual variability: it essentially evaluates how well the datasets capture snow appearance and disappearance dates and the variations of SWE throughout the season. To investigate whether the datasets reproduce the inter-annual variability observed in the field data, correlation coefficients obtained using only the January values

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were considered for the available periods of each dataset. Globsnow and JULES both had 61 % of grid-cells with $\rho \geq 0.6$, but the corresponding values for LPJ-WM, SDGVM and CLM4CN were only 35 %, 25 % and 12 % respectively. There were insufficient data in the LEGOS product to make a significant estimate, since these records start only in 1988. This approach revealed strengths and weaknesses of each model in simulating SWE (e.g. JULES achieving exceptional timings but severe underestimation of magnitude) with the discussion continuing in Sect. 5.1 where the effects of such shortcomings are investigated and Sect. 5.2 where an optimization technique is presented.

4.2 Fresh snow density

The water-to-snow density ratio of fresh snow is often considered to follow the “10-1 rule”, i.e. fresh snow has a density of 100 kg m^{-3} , although factors such as atmospheric temperature can cause variations (Judson and Doesken, 2000; Roebber et al., 2003). This rule is followed by JULES, but LPJ-WM adopts a snow density of 150 kg m^{-3} for the first three quarters of the snow season, and CLM4CN defines fresh snow density as a function of air temperature according to Anderson (1976), varying from $50\text{--}170 \text{ kg m}^{-3}$ (SDGVM does not calculate snow density or snow depth, so is not relevant to this section).

The box plots of Fig. 3 show the mean, median, 25 % and 75 % percentiles and range of snow density acquired from a total of 60 pairs of transect measurements for days $N = 1, 2$ and 3 following a first snowfall, which according to the method detailed in Sect. 3.2.1 represent mean snow age at days 1/2, 1 and 3/2 respectively. Fresh snow density and density evolution in a range of models are also shown. The Canadian Land–Surface Scheme (CLASS) model (Verseghy et al., 1993) defines snow density evolution as an exponential function of time following

$$\rho(t) = [\rho_{\text{fr}} - \rho_{\text{max}}] e^{-Rt} + \rho_{\text{max}} \quad (4)$$

where $R = 0.24 \text{ day}^{-1}$ defines the rate of densification, ρ_{fr} is the fresh snow density and ρ_{max} the maximum attained snow density set at 100 kg m^{-3} and 300 kg m^{-3} respectively

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with time t in days. After applying a logarithmic transformation to Eq. (4), linear regression was used in order to acquire the best-fit estimates for the parameters ρ_{fr} and R by using the 3 snow density values as realizations of $\rho(t)$ for $t = 0.5, 1$ and 1.5 while preserving the value of ρ_{max} . After reversing the transform, the new attained, median-representing values were $\rho_{\text{fr}} = 91.4 \text{ kg m}^{-3}$ and $R = 0.37 \text{ day}^{-1}$ showing a higher rate of densification but a relatively smaller fresh snow density value. Figure 3 shows the original parameterization of CLASS as well as the one (CLASS*) where its parameters were derived by the regression described above.

By adopting the snow density evolution of CLASS and utilizing the *synoptic* dataset, Fig. 3 suggests that for the spatial coverage offered by the transect data in Eurasia, a value of fresh snow density within the range $57\text{--}117 \text{ kg m}^{-3}$ (80 % confidence interval for $t = 0$ days) is appropriate. This is within the lower half of the range proposed by CLM4CN and encompasses the 100 kg m^{-3} value set in JULES, but the initial value assumed by LPJ-WM appears too high, most likely used to compensate for the lack of snow density evolution by the model for the initial 3/4 of the snow season.

4.3 Snow density evolution

Monthly values of snow density were extracted from the *transect* dataset and compared with the monthly density obtained by JULES and CLM4CN for each month and gridcell possible over the period 1966–1996 and across the spatial extent of the data; these are the only two models containing processes controlling density.

CLM4CN performs very well in describing the increase of snow density throughout the year ($\rho = 0.60$) as shown in the plot of Fig. 4 (top). When individual months are examined, low correlation is observed in October and the model overestimates snow density in March; both issues can be attributed to inaccuracies in snow timings, since these two months mark the start and end of the snow season for the region. It was indeed found that most of the areas with poor snow density correlation also exhibited poor SWE correlation or overestimation of SWE (Fig. 2). Snow density is affected by overburden pressure, which is taken as proportional to SWE in CLM4CN, hence

improved representations of SWE magnitude and timings are likely to improve simulation of snow density evolution.

In contrast, snow density evolution in JULES is characterized by a very slow increase after deposition, especially during October and November. This does not agree with the *transect* dataset and causes JULES to underestimate density throughout the winter (Fig. 4, bottom), perhaps as a result of its underestimation of SWE seen in Fig. 2. JULES would clearly benefit from improved representations of both snow density evolution and SWE which, as discussed later, creates issues when soil temperature simulations are considered.

4.4 Snowpack sublimation

Table 1 gives the monthly snowpack sublimation, S_s , derived using the methodology described in Sect. 3.4, for the western and eastern sectors shown in Fig. 1, both of which have a high density of stations. The values for the JULES and SDGVM models also shown (note that LPJ-WM does not consider sublimation and CLM4CN does not include it in its standard list of outputs). The eastern sector experiences harsher winters with very low temperatures and a snow season spanning November to late April, compared with late November to March in the west as indicated by the SSM/I snow cover product. Both the VCF land cover product (see Sect. 3.1) and the metadata for the *synoptic* records indicate that the western sector consists mainly of herbaceous cover, while the eastern sector is a mix of deciduous forests and herbaceous cover. By comparing adjacent forest and field sites, Zhang et al. (2005) and Reba (2012) showed that S_s is smaller in forest, mainly because of lower wind and net radiation, the latter of which is very sensitive to canopy characteristics (Davis et al., 1997). Furthermore, as Dery and Yau (2002) showed, the relative air humidity with respect to ice is lower in western Siberia than the east hence sublimation is likely to be higher in the western sector. This is consistent with the monthly estimated values of snowpack sublimation presented in Table 1; furthermore since there is little forest cover in the western sector, the retrieved values of S_s are estimates of overall sublimation.

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Only limited independent information is available to assess the estimates in Table 1. Zhang et al. (2005) measured snowpack sublimation in the Mogot Basin (55.5° N, 124.7° E) in eastern Siberia in a field and a larch forest site, and for the period 13 March–22 April of 2002 found sublimation rates of 15.7 mm and 12.1 mm for the field and forest sites respectively. 11 stations were found in a 6° × 6° grid centered on the Mogot Basin whose estimated monthly snowpack sublimation values are presented in Table 2. The stations are ordered from highest to lowest fraction of herbaceous cover, as indicated by the rightmost column. Although the field measurements from Zhang et al. (2005) were restricted to a specific basin and the late part of the snow season for a specific year, their measured sublimation values are comparable with our estimates for March. The overestimation observed for the herbaceous cover sites could be attributed to seasonal or episodic melting events that were not eliminated by the 80th percentile rule (see Sect. 3.4).

For western Eurasia, no field measurements were found in the literature to evaluate the sublimation retrieval, but global estimates of annual snowpack sublimation have been provided by Dery and Yau (2002), using process modeling and the 6 hourly European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA15) dataset at 2.5° resolution for 1979 to 1993 (Gibson et al., 1997). Their published values range from 5 to 25 mm yr⁻¹ for eastern Russia, and from 50 to 75 mm yr⁻¹ for western Russia, compared with 22.1 mm and 39.7 mm respectively from our methodology (excluding the melting season). Both approaches agree that sublimation is higher in the west, though this difference is smaller in our estimates than in the modeling approach of Dery and Yau (2002).

As data is sparse to provide a proper evaluation of the snowpack sublimation retrieval presented in this study, comparisons described above are encouraging and justify the use of the retrieved values to assess model performance. For the eastern subset and months November to April, JULES gives a small overestimate in the eastern sector (Table 1), possibly because of high November sublimation, but for the other months values produced agree reasonably well with the transect-based estimates both

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in magnitude and variability ($\rho = 0.86$, p value = 0.06). Similarly, for the western subset, JULES agrees well with the results of the retrieval ($\rho = 0.82$, p value = 0.17) and is at the lower limit of the values produced by Dery and Yau (2002). SDGVM captures the variability of sublimation ($\rho = 0.90$, p value = 0.03 for the eastern sector; $\rho = 0.93$, p value = 0.06 for the western sector) but systematically underestimates its magnitude by a factor of 4 and fails to capture land cover effects, giving similar values for the east and west. The optimization method of Sect. 5.2 explains how these shortcomings of SDGVM can be partly alleviated.

5 Discussion

From an Earth-system perspective, these results prompt the question of how an inaccurate description of the snow regime in a land surface model can affect other elements of the land surface system it is attempting to simulate. Despite the relatively long snow season found at these latitudes, climate data, such as CRU TS 3.0, show that the bulk of precipitation occurs during the spring and summer months. Even though inaccuracies in the models would affect runoff during the melting season, this will have little impact on the overall water balance, which is mostly affected by the spring and summer rainfalls and evapotranspiration during those months. During winter, though, the snowpack acts as a thermal buffer whose depth and density affect heat transfer between the atmosphere and soil; an inaccurate description of the snow regime can therefore affect simulations of soil temperature and permafrost extent. As land surface models start to include permafrost parameterizations, it is essential to understand the magnitude and nature of complications that arise when the problems with simulating snow processes mentioned earlier are left untreated. To investigate how large these effects could be, the atmosphere–snowpack–soil heat exchange formulation of LPJ-WM, the only land–surface model with a complete permafrost parameterization and organic soil consideration, was forced with snow variables acquired from the field data using the methodology detailed in Sect. 3.3.

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5.1 Driving LPJ-WM with snow-related field measurements

The 30 yr time-span and 5 to 10 day temporal spacing of the field records provide sufficient data on snow variables, such as depth and density, to force the heat diffusion formulation of LPJ-WM. Furthermore, since it treats snow density as constant except in the last quarter of the snow season, comparison of model calculations with and without imposing observed values of snow density offer an insight on the feedbacks of opting for a simplified snow density evolution.

Differences between modeled and field values of snow depth and density for three field stations are shown in Fig. 5, together with values of soil temperature at 25 cm depth estimated by the model in its unmodified form, and when driven by observed snow depth alone (LPJ-WM S) and by both depth and density (LPJ-WM SD). As expected, where LPJ-WM overestimates snow depth, modeled soil temperatures are lower in winter when the model is forced by observed snow depth. This is illustrated by Fig. 5 (top) for the Njuja station in eastern Siberia in a region of continuous permafrost. Winter soil temperatures dropped significantly, with an average decrease of 6°C at 25 cm depth during January for the period 1966–1996. However, there was little difference in soil temperature for the upper soil layers during the late spring and summer months. As a consequence, the integrated annual heterotrophic respiration decreased by approximately 7.4 %, but the summer thaw depth was unaffected. Note that, for the 9 stations examined, the biggest difference in snow depth between LPJ-WM and field data was at Njuja, and so the largest differences in heterotrophic respiration were observed here.

The Tevriz station is located in central Siberia and here LPJ-WM places the permafrost boundary at a depth just below 25 cm (Fig. 5, middle). Here there is less consistency between modeled and observed snow depth than at Njuja, with good agreement in some years but overestimates in others. The soil response is also less consistent: significant overestimates of snow depth do not always yield reduced winter temperatures (for example, in 1983 or 1984). In fact, for those particular years, even though

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field data had indeed less snow than the model, they also showed an earlier start to the snow season which compensated for the reduced snow depth later on and prevented a lower soil temperature. Although temperature differences up to 5 °C are observed for January in 1980 and 1981 for the Tevriz station, the summer temperature shows only a small decrease, so again the thaw depth is unaffected while heterotrophic respiration reduces by 7.3 %.

At Olekminsk station in eastern Siberia, the magnitude and timing of snow depth are very similar in LPJ-WM and field data (Fig. 5, bottom) so forcing the model with field data has little effect on the simulated soil temperatures, and causes integrated annual heterotrophic respiration to reduce by only 1.3 %. Additionally, as for the other 2 stations, it was found that forcing the model with snow depth and snow density made little to no difference to soil temperatures than when forcing it with snow depth alone. It was therefore demonstrated that even though simulations of soil temperature and consequently soil carbon decomposition in LPJ-WM are sensitive to inaccuracies of both snow magnitude (Fig. 2, right) and timing (Fig. 2, left), soil temperature simulations are robust towards snow density and the simplified approach of LPJ-WM is deemed adequate.

5.2 Optimization of the SDGVM model

While results of Sect. 4 revealed the varying shortcomings of each model in capturing the snow dynamics, sensitivity experiments of Sect. 5.1 demonstrated how they can manifest into perturbations in the simulation of important carbon fluxes in northern latitudes such as soil carbon decomposition. Justifiably, a ready, model-independent optimization method was sought out which would employ the field records of snow water equivalent available in this study to improve model snow parameterization and consequently simulations. A fast model was desirable as any multi-dimensional optimization routine would likely require hundreds of computations of an error function, the production of which would require a model run. For this study the SDGVM was used as it is the fastest model to run from our model suite due to its relatively large time step (daily

as opposed to hourly) and simple process descriptions. For the observational data, monthly averages were produced from the *transect* data which contains 1345 sites. The optimization process consists of finding the minimum of an error function that describes the difference between model and observation. The error function was taken to be the average absolute monthly difference in snow months between observation and model, averaged over all sites. Where, a snow month is defined as any month having a positive value of SWE, either observed or modeled. For the optimization, a modified version of the “Downhill Simplex Method in Multi-dimensions” (Press et al., 1992) was employed. The algorithm was modified to allow for sensible constraints to be applied to the physical optimization parameters.

Four parameters were chosen which were thought to play a major role in the timings and dynamics of the snow pack. Firstly, the average daily temperature that snow occurs. This parameter determines the start date of the snow pack and is set to 0°C in SDGVM. Secondly, the average daily temperature at which snow melts. This parameter impacts the end date of the snow pack and is also set to 0°C. Thirdly, the rate at which snow melts which also impacts on the end date of the snow pack. This parameter is proportional to temperature. Fourthly, the sublimation rate which affects the overall size of the snowpack; this parameter is proportional to potential evapotranspiration. The system is biased towards Western Europe as it contains a high concentration of observational sites (Fig. 1). This has been negated to some degree by restricting the sites used for the optimization to one per 2° grid square of latitude and longitude, and using the one most observed where multiple sites occur in such a grid square.

5.2.1 Optimization results

After the optimization process was applied the average absolute error in monthly SWE between SDGVM and transect measurements reduced from 33.9 mm to 28.3 mm with the reduction of SWE in Western sector being largely responsible where the correlation was also improved (Fig. 2, SDGVM-Opt). The optimized parameters for SDGVM are

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given in Table 3 with large differences in all parameters particularly the temperature limit for snow start and sublimation rate.

5 The significant reduction in the snow start temperature limit points to a shorter snow season and consequently less snow will be produced over the year for all sites. Similar effect will have the increase of the sublimation rate by almost a factor of 4; this is indicative of the underestimation of sublimation in SDGVM and agrees with the findings of Sect. 4.4. Both of these changes lead to reduce the overestimation of SWE by the model and improve its description of the snow dynamics in the west. As it was demonstrated earlier, all models suffer from inaccuracies in describing the snow regime which vary by model. By utilizing the field measurements, hotspots of uncertainties can be located and ready-available optimization techniques such as the one described here can pinpoint and correct problematic model processes and parameters. While in this study the climate drivers of SDGVM remained unaffected and the optimization was carried out on model parameters, Brun et. al. (2013), by employing the same field measurements, optimized the performance of the Interactions between Soil–Biosphere–Atmosphere (ISBA) model (Noilhan and Mahfouf, 1996) by retaining model parameterization but using different climate drivers which they assessed based on best-fit statistics between simulated snow variables and the field measurements. Both these methods can be used independently or in tandem to optimize snow simulations in a land surface model.

6 Conclusions

The first part of the present study extensively utilizes field transect measurements to evaluate the performance of a range of Earth Observation and land surface models in describing snow dynamics across the extend of the FSU. Findings suggest:

- 25 – GlobSnow offers an accurate SWE product which, due to its global coverage, resolution and 30 yr time span, is invaluable for benchmarking SWE retrievals in

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land surface models, in case field data is insufficient, by pinpointing regions where improvement is needed.

- LPJ-WM, SDGVM and CLM4CN all perform well in reproducing the temporal dynamics of SWE in Eurasia, although there are model-specific regions where there are systematic and significant anomalies, particularly in the western sector, where none of the models captures either the variability or magnitude of SWE.
- JULES correctly simulates both the seasonal and inter-annual variability of SWE throughout Eurasia but systematically underestimates SWE throughout the region.
- A novel approach demonstrated that viable values of fresh snow density for the region range from 57–117 kg m⁻³. CLM4CN captures the snow density evolution, while JULES underestimates snow density throughout the season. In both cases, better simulation of the magnitude of SWE may improve the simulated evolution of snow density.

The analysis on snowpack sublimation bestows confidence both in the related processes and modeled values produced by JULES as well as the methodology detailed in this study for retrieving snowpack sublimation from the transect records. The latter is further being exploited to produce a relevant and unique dataset across Eurasia shedding light on the dynamics of sublimation with specific interest on its inter-annual variability and trends as well as its correlation with land cover and prevailing meteorological conditions acquired from station data.

By driving the atmosphere-snowpack-soil heat exchange formulation of LPJ-WM with the snow metrics acquired from field measurements, it was revealed how inaccuracies in snow variables in a land surface model can feedback both to the simulated heterotrophic respiration and the winter soil temperatures although thaw depth seemed to remain unaffected. As showed, for locations where LPJ-WM overestimated SWE, winter soil temperatures were much lower than the ones originally produced by the model.

Dankers et al. (2011) ran a version of JULES which incorporated permafrost dynamics in order to study and evaluate the simulated soil temperatures. Their finding was a significant negative bias of soil temperatures which they attributed to inaccuracies in JULES SWE noting that the bias was much higher in winter than in summer. This is consistent with this study where it was found that JULES systematically underestimates SWE to a significant degree which in tandem with the finding regarding LPJ-WM would suggest that for the winter months JULES would indeed produce much lower soil temperatures. In their study Dankers et al. (2011) relied on a restricted number of field data around the Arctic. The abundance of the transect records spread across Eurasia along with the methodologies presented in this study offer the opportunity not only to investigate, optimize and correct for JULES underestimation of SWE but also for its density formulation which as showed was problematic although the LPJ-WM runs demonstrated that it will likely have a minimal effect on simulated soil temperatures.

Overall, this study proves beyond reasonable doubt the importance of systematic field measurements as means to evaluate land surface models. From a single dataset it was possible to evaluate the SWE simulated by prominent land surface models, infer about fresh snow density and snow density evolution, quantify perturbations of uncertainties of SWE in simulated soil temperatures and evaluate modeled snowpack sublimation. Nevertheless, despite the ability of field and earth observational datasets to pinpoint weaknesses in model formulations, efforts to optimize the ECVs they produce are often overlooked especially in favor of the addition of new processes. For example, the step of benchmarking SWE and snow density in JULES before adding permafrost formulations was not followed which is the cause of its observed negative bias in soil temperatures. Furthermore, without any sublimation processes included, LPJ-WM considered the more complex organic soils and permafrost dynamics with the use of an over-simplified snow density evolution. Additionally, even though SDGVM includes sublimation processes, a bibliographic review would have showed the problems which this study revealed in detail. These findings highlight the need for better integration of already existing, detailed and accurate datasets (field or EO) of ECVs in the community

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of land surface models with the scope of benchmarking the simulation of essential aspects of the earth system and reduce the uncertainties imposed on attempted short and long term predictions.

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Table 1. Estimates of monthly snowpack sublimation (November to April) and total sublimation obtained by the method of Sect. 3.4 and respective values for the JULES and SDGVM models for the eastern and western Eurasian subsets shown in Fig. 1. For the retrievals, the summation (^a) of sublimation values excludes the melting month(s) for each sector when sublimation values cannot be obtained (–).

Snowpack Sublimation (mm) Eastern Subset							
	Nov	Jan	Feb	Mar	Apr	Sum	
Estimated	1.53	1.52	3.05	3.80	12.20	–	22.1 ^a
JULES	5.7	2.9	2.9	6.4	10.2	5.2	33.3
SDGVM	1.35	0.73	0.68	1.15	2.61	4.6	11.12
Snowpack Sublimation (mm) Western Subset							
	Nov	Dec	Jan	Feb	Mar	Apr	Sum
Estimated	9.2	6.1	9.15	15.25	–	–	39.7 ^a
JULES	5.7	7.2	7.3	11.3	16.4	1.2	49.1
SDGVM	1.0	0.9	0.9	1.2	2.8	4.4	11.2

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Table 2. Estimated monthly (November to March) snowpack sublimation values (mm) for 11 meteorological stations around the Mogot basin; also given are the altitude (m) and the fraction of transect records with herbaceous and forest cover for each station.

Station Name	Nov	Dec	Jan	Feb	Mar	Alt. (m)	Fractional Cover (Herbaceous-Forest)
Zeja	–	3.05	4.58	10.2	18.3	232	1.0–0.0
Nagornyj	6.1	3.05	7.63	18.3	19.8	862	0.81–0.19
Tygda	–	–	–	11.6	24.4	314	0.78–0.22
Skovorodino	3.05	3.35	7.63	6.1	18.3	400	0.77–0.23
Tynda	–	–	–	3.0	15.3	513	0.75–0.25
Magdagaci	–	–	–	–	9.15	384	0.39–0.61
Ignasino	–	–	–	1.52	6.1	295	0.27–0.73
Cul'man	–	–	–	22.3	12.2	664	0.15–0.85
Urusa	–	–	–	6.86	15.2	454	0.0–1.0
Dzalinda	–	–	–	1.7	12.2	267	0.0–1.0
Unaha	–	–	–	2.12	6.1	453	0.0–1.0

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Table 3. Original and optimized parameters of the SDGVM model: snow start temperature limit (°C), snow melt temperature limit (°C), snow melt rate, sublimation rate.

Parameter	SDGVM	SDGVM-Opt
Snow start Temperature limit	0 °C	−4.38 °C
Snow melt Temperature limit	0 °C	0.61 °C
Snow Melt Rate	1	0.31
Sublimation Rate	1	3.79

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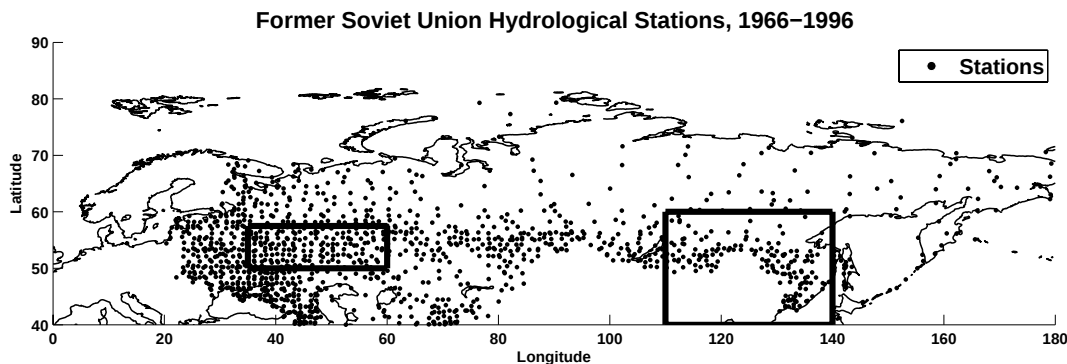


Fig. 1. Location of hydrological stations considered in this study; note the scarcity over northern Siberia. The limits of the western and eastern geographical subsets used in Sect. 4.4 are indicated by the rectangles.

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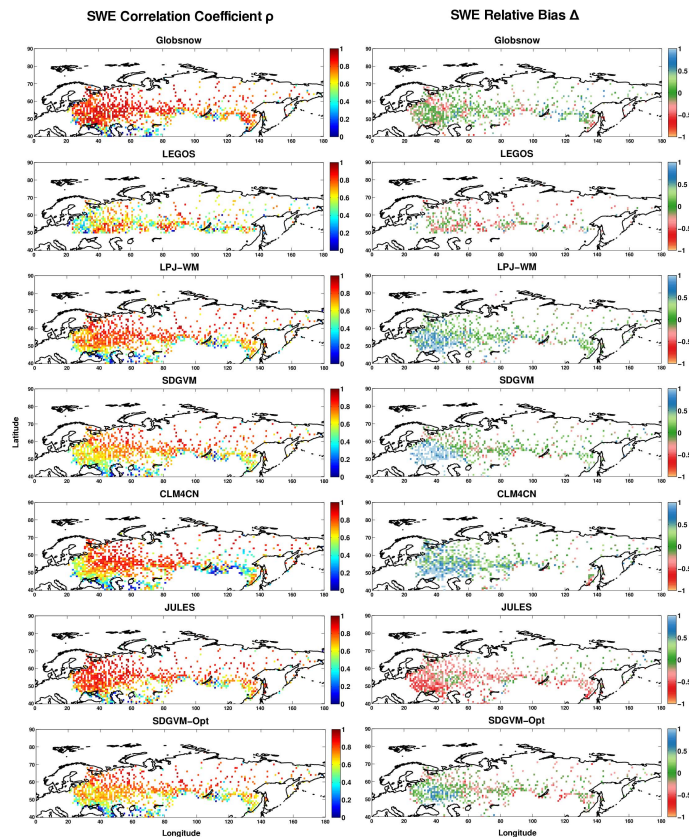


Fig. 2. (left) Correlation coefficient of monthly SWE, ρ , between the *transect* record and the six datasets for 1966–1996 across the extent of the FSU. (right) Relative bias, Δ , between the *transect* records and the datasets for grid-cells with $\rho \geq 0.6$. Figures titled SDGVM-Opt relate to the results of the optimization experiment detailed in Sect. 5.2.

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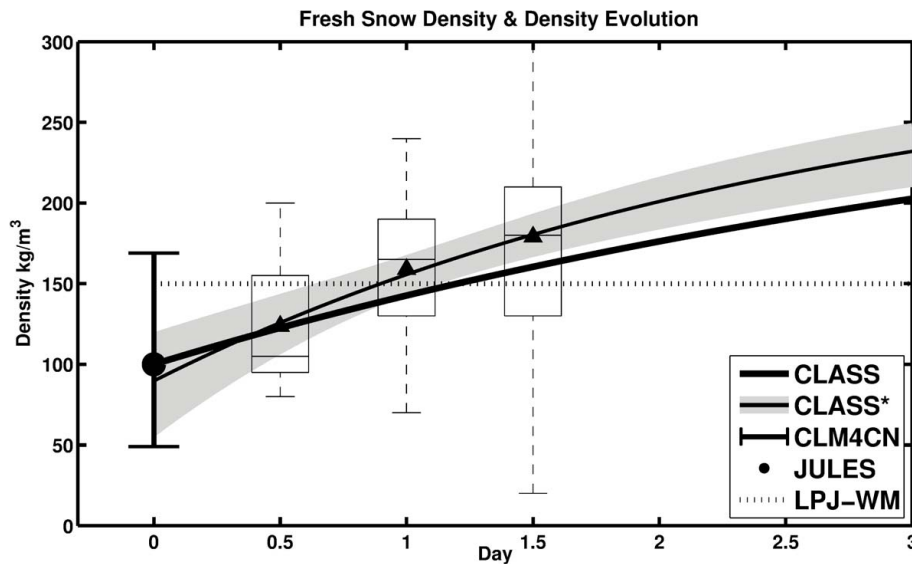


Fig. 3. Fresh snow density (day = 0) in CLM4CN, JULES and LPJ-WM as well as snow density evolution in the CLASS model. Box plots show snow density statistics extracted from the transect data with the triangle showing the mean value. CLASS* shows the snow density evolution in the CLASS model but with parameters obtained through regression analysis; gray area depicts 80 % confident intervals.

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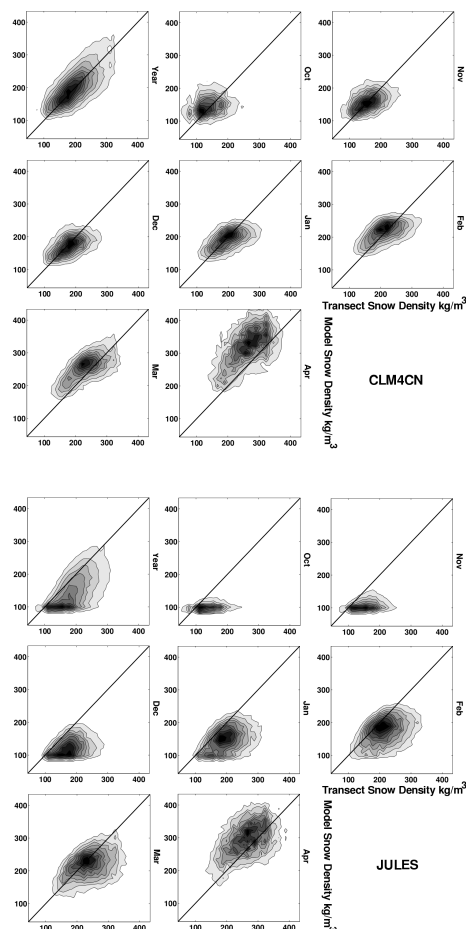


Fig. 4. Density histograms of monthly snow density from CLM4CN (top) and JULES (bottom) compared to the *transect* dataset for the entire snow season (year) and for individual months, across the full spatial and temporal extent of the *transect* dataset.

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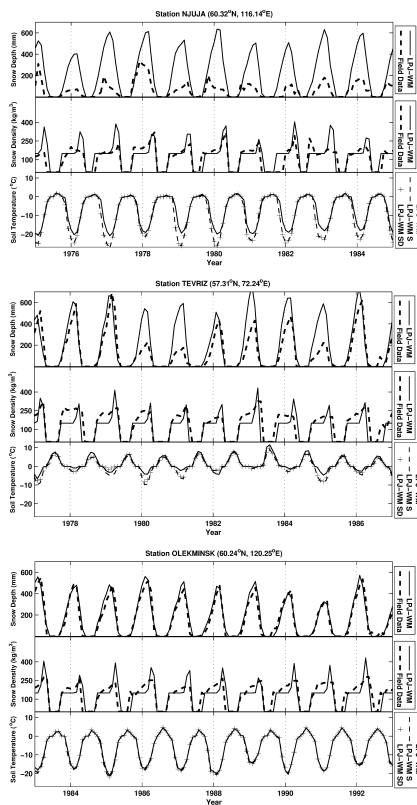


Fig. 5. Snow depth (mm) and snow density (kg m^{-3}) acquired from LPJ-WM and field records for 3 WMO stations located in eastern and central Siberia. The bottom subplot in each figure shows the monthly soil temperature ($^{\circ}\text{C}$) at a depth of 25 cm produced by LPJ-WM, along with the corresponding temperature when the model is driven by snow depth alone (LPJ-WM S) and both snow depth and density (LPJ-WM SD) using values taken from the field records.