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Carbon stocks and fluxes in the high latitudes: using site-level data to evaluate Earth system models

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	Abisko	Bayelva	Kytalyk	Samoylov	Zackenberg
Snow depth	2005-2012	1998-2009	2011-2013	2002-2013	1997-2010
Soil temperature	2005-2012	1998-2009	2011-2013	2002-2005	1995-2010
Soil moisture	2015	2009-2013	-	2002-2005	2006-2013
CO_2 flux	2012-2013	2008-2009	2003-2013	2002-2013	2000-2008

Table S1 | Years for which observational data were used for mean annual cycles.

	Abisko	Bayelva	Kytalyk	Samoylov	Zackenberg
JSBACH	12.4	8.2	16.9	16.9	29.4
JULES	26.7	0.05	3.2	3.6	1.2
JULES fixed veg	35.0	4.9	12.9	13.1	5.5
ORCHIDEE	8.9	1.0	7.5	14.8	0.05
ORCHIDEE no mixing	12.6	1.6	10.1	17.4	0.0
Observations	**39**	**8**	**29**	**24**	**13**

Table S2 | Total soil carbon in top 1m (kgm^{-3}). Observations are based on the flux tower footprint areas. Models are averaged for 1990-2013 inclusive (note they do not change substantially over the simulation period).

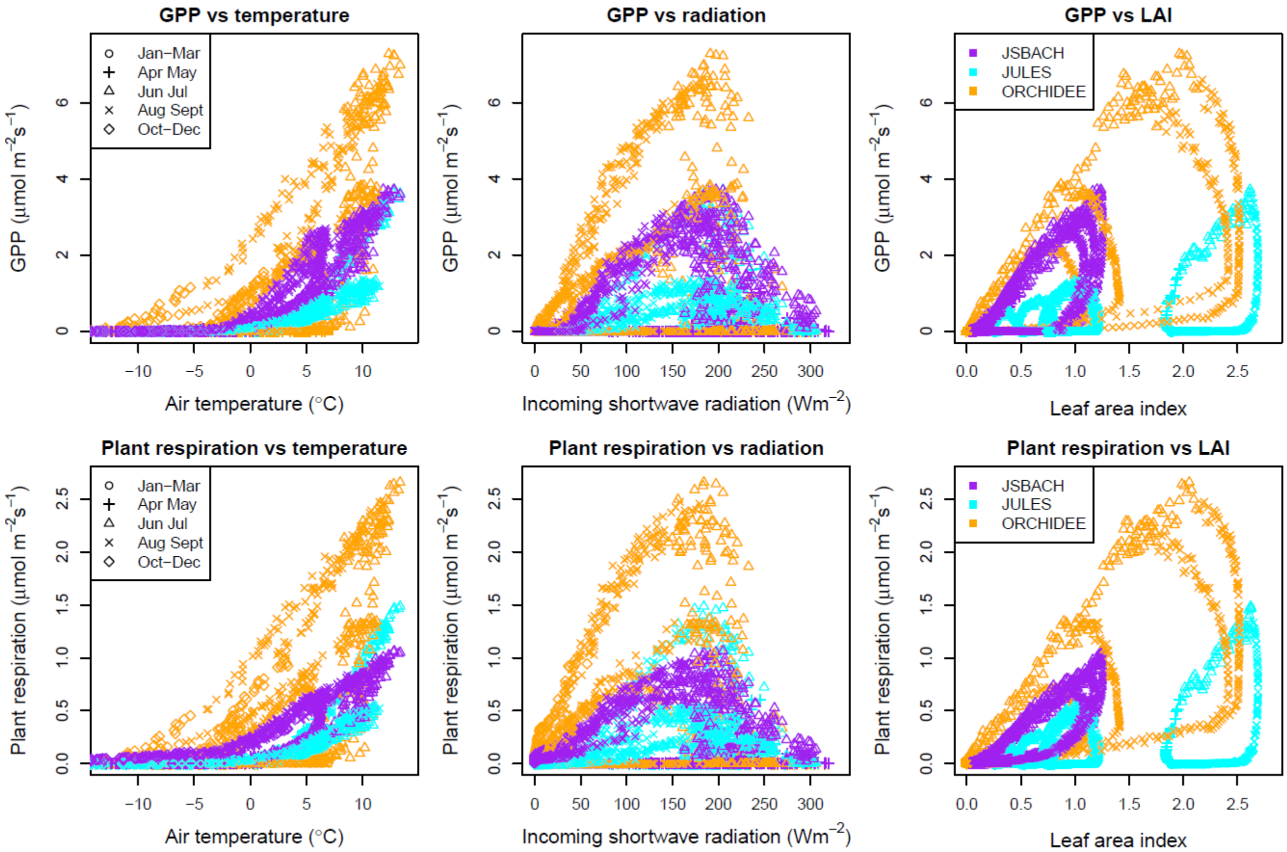

Figure S1 | Relationship of modelled GPP and plant respiration with different variables: Air temperature, incoming shortwave radiation and LAI. All sites are included on each plot. Shapes of points correspond to different parts of the season.

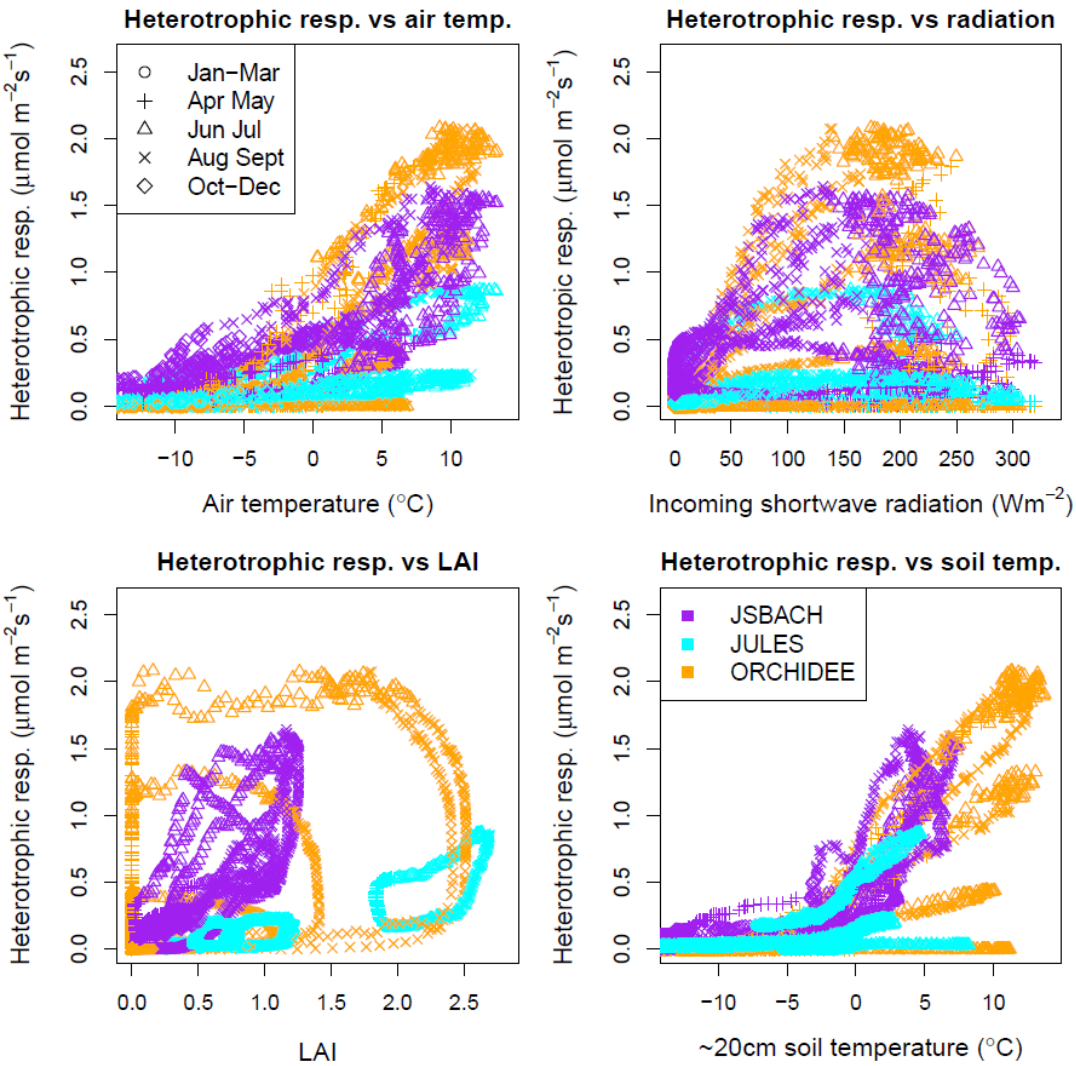

Figure S2 | Relationship of modelled soil respiration with different variables: Air temperature, incoming shortwave radiation, LAI and 20cm soil temperature. All sites are included on each plot. Shapes of points correspond to different parts of the season.

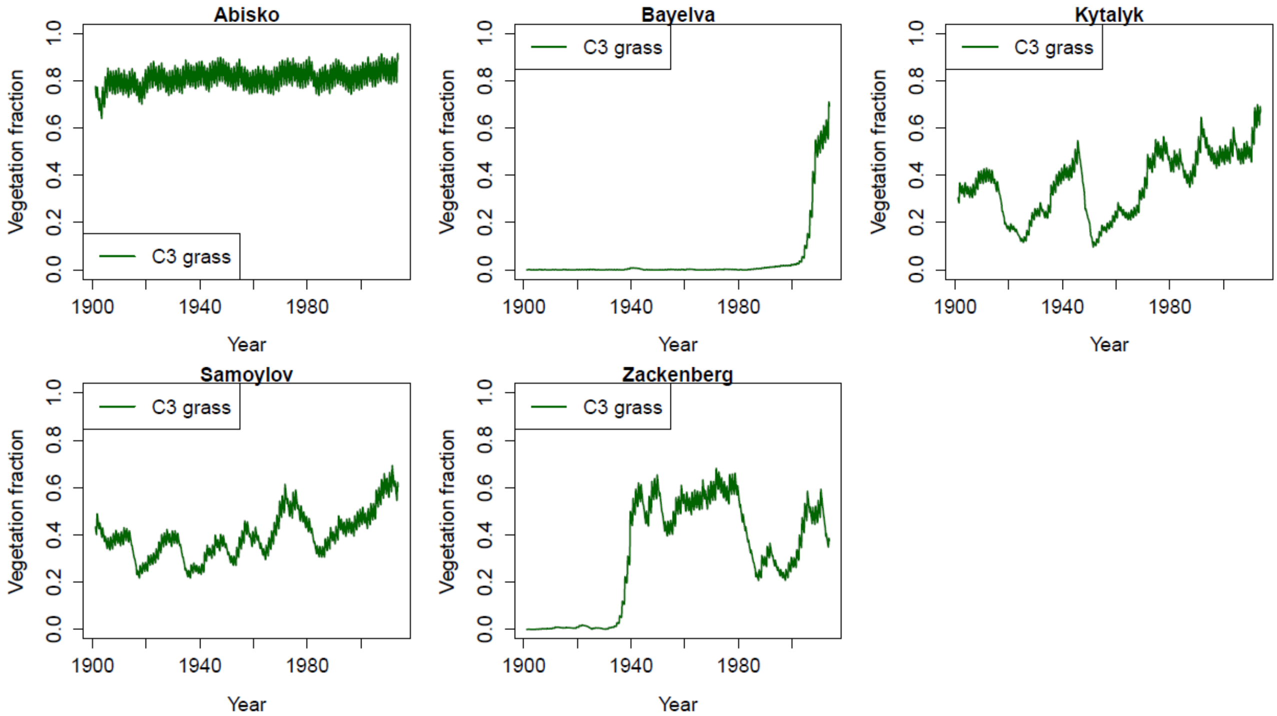

Figure S3 | Vegetation fractions in JULES as simulated by the dynamic vegetation model, TRIFFID. There are 9 PFT's that can potentially grow, but only C3 grass is able to grow at these sites.

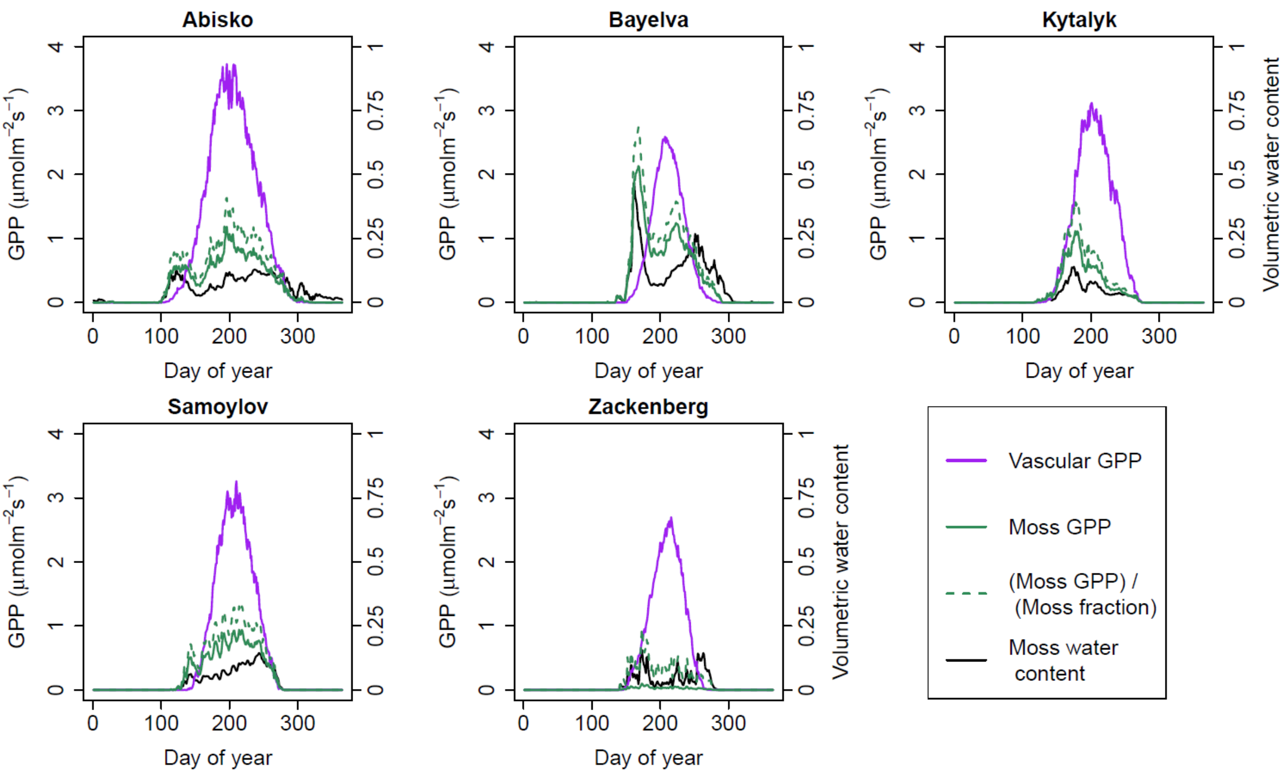

Figure S4 | Annual cycle of moss GPP and vascular plant GPP in JSBACH, showing the difference in their seasonal cycles. The moss moisture content is also shown to demonstrate the link between moss GPP and water content.

Sensitivity to snow

Representativeness of snow depths

In flat, open tundra landscapes, the snow is heavily affected by wind blowing, with the consequence that snow depth does not correspond directly to precipitation, and therefore using direct snowfall measurements is not possible in these landscapes. This scenario particularly applies to Kytalyk, Samoylov, and the Abisko mire. On a large scale, the snow can be quite even distributed due to the flat landscape (e.g. Blok et al., 2010, Table 2), but the microtopography at these sites (e.g. ice wedge polygons, palsas) leads to small-scale variability in snow depths. For example at Samoylov the depressed polygon centers have much deeper snow than the elevated rims (Boike et al., 2013). Thus a single point measurement of snow depth may not be representative of the whole flux tower footprint. At Abisko, however, several locations on the mire are averaged to give a representative sample.

At Bayelva and Zackenberg, the landscape is more mountainous, and there is more variation in snow depth around the area due to the topography of the land (and consequent differences in vegetation). At Zackenberg the snow is measured on transects across different vegetation types and the values range from snow-free to more than 1m of snow at a single time. However, the flux tower is situated in a fairly homogenous *cassiope* heath where snow surveys show the typical standard deviation of snow depth any one point in time is around +/-12cm (ZEROCALM1, https://data.g-e-m.dk/, average depth around 50cm). For this site, the point observation appears to be representative of the flux tower footprint. At Bayelva the snow depth varies by around +/-50% within the vicinity of the flux tower (Gisnås et al., 2014), and our point observation falls a little higher than the typical values for maximum snow height.

Even for sites where a point measurement of snow depth is representative of the flux tower area, the snowfall timeseries is derived using an assumed density and could be better parameterised using snow density measurements.

Sensitivity study

To investigate the impact of the variability and uncertainty in snow depth, we performed a sensitivity study. The observations suggest that increasing and decreasing the snow depth by 50% from the model simulated values would capture the range of observed snow depths in each of these landscapes. Since the snow depth is dynamically simulated rather than input to the models, we approximated the change in snow depth by increasing and decreasing the snowfall forcing by 50%. Two of the models (JSBACH and JULES) were then re-run (including spin-up) in these two different configurations. Snow depth in these simulations now spans a range that includes the point observations (Figure S5).

As expected, increased snow depth leads to an overall warming of the soil for every site, and reduced snow depth leads to a cooling (Figure S6). However, most of the change happens in winter, where it will have less impact on the carbon cycle since the vegetation and soil decomposition processes take place mainly in summer (JJA) (Figure S6).

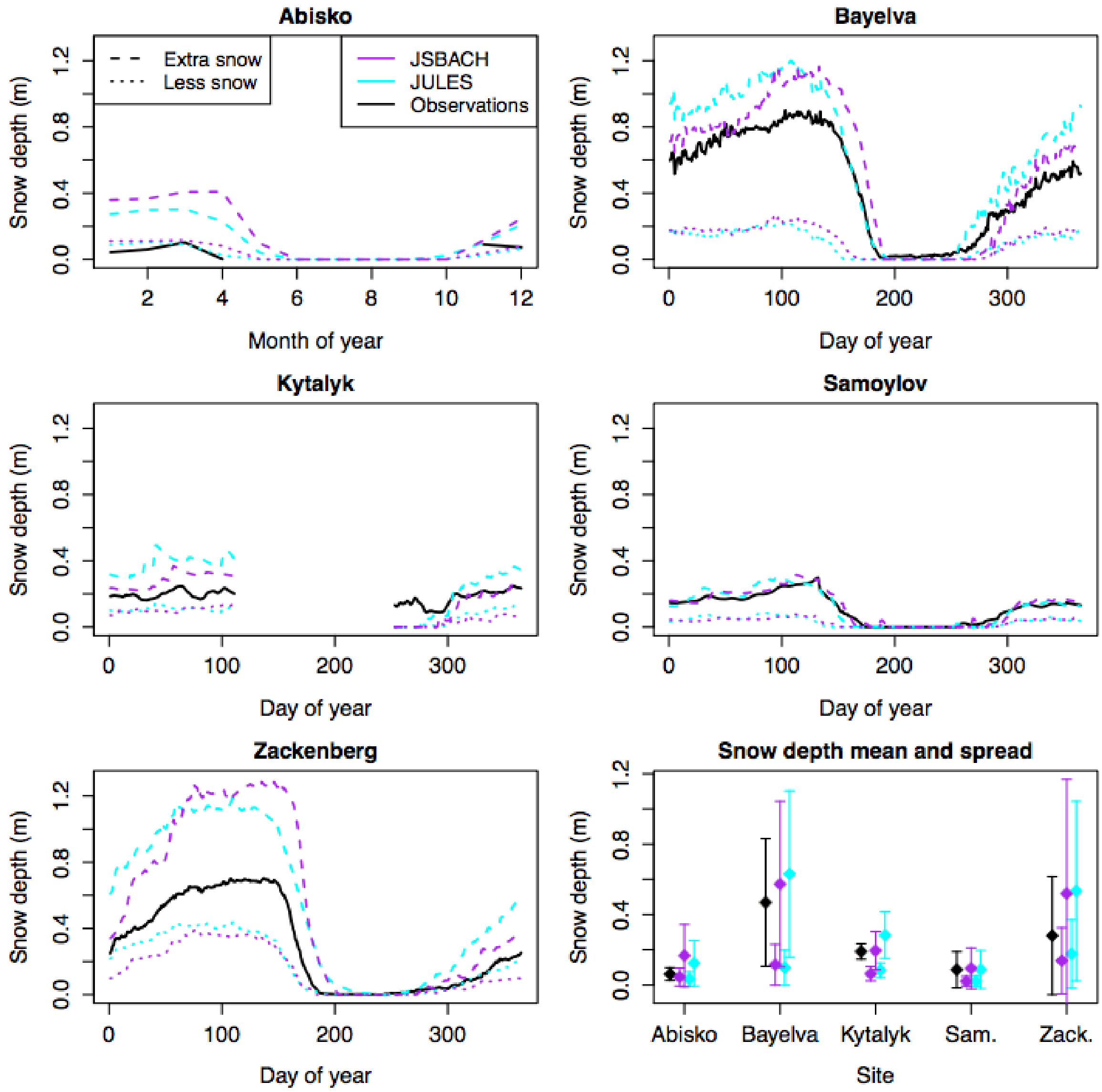

Figure S5 | Mean annual cycles of snow depth (as in Fig. 1 in main manuscript) showing simulations with increased and reduced snowfall in JSBACH and JULES.

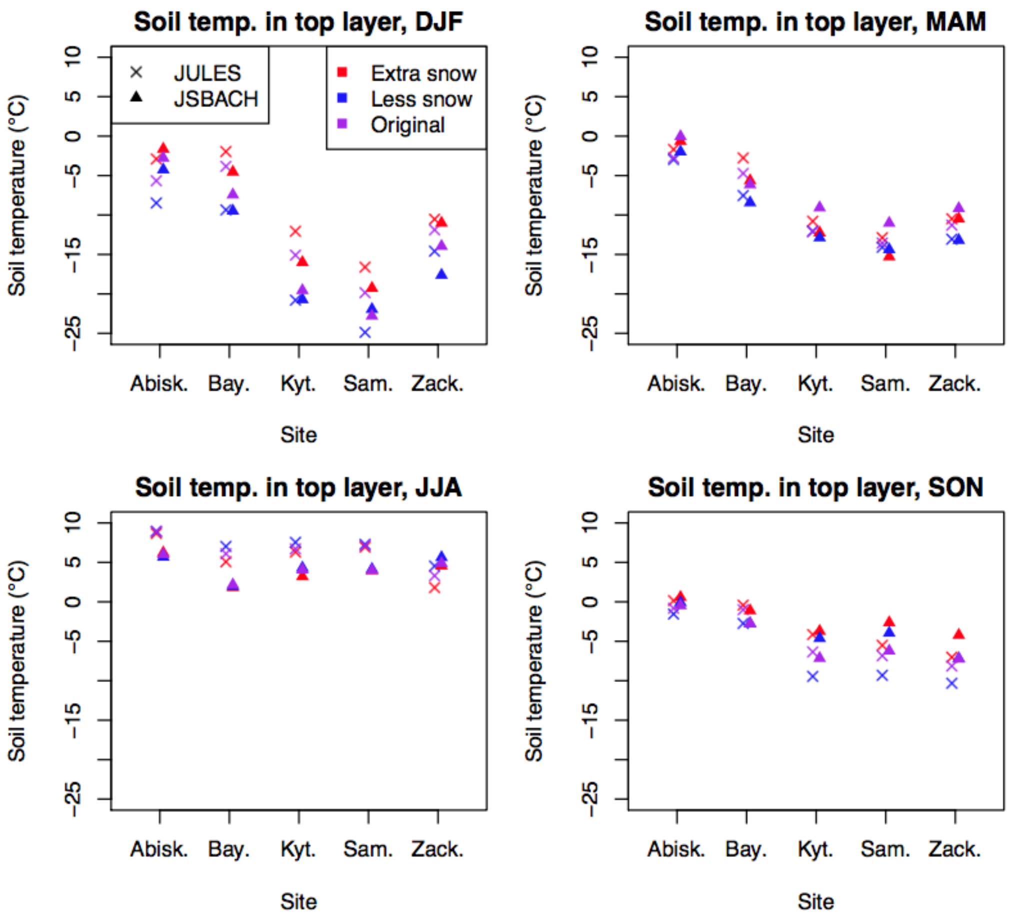

Figure S6 | Mean soil temperature in different seaons, showing simulations with increased and reduced snow for JULES and JSBACH. (DJF=December, January, February. MAM=March, April, May. JJA=June, July, August. SON = September, October, November.)

Vegetation growth is not directly impacted by snow or soil temperature changes in these models. However, the change in winter snowfall also leads to changes in soil moisture during summer, which does affect vegetation growth. An increase in snow should lead to an increase in water infiltration into the soil in spring and thus an increase in the available soil moisture. In JULES, however, for two of the sites (Zackenberg and Bayelva) the opposite effect is seen, where increased snow depth leads to less soil moisture in summer, and vice versa (Figure S7). In JULES, the changes in soil moisture are reflected in the GPP, ecosystem respiration (Reco) and vegetated fraction, which all increase with higher soil moisture and reduce with lower soil moisture (Figure S7). At many of the sites these are significant changes (although they still leave the model with low values of GPP/Reco compared to observed fluxes). The impact of any change in GPP is amplified by the resulting changes in vegetation fraction. In JSBACH, however, the changes in soil moisture, GPP and Reco are not significant (Figure S7).

Soil carbon stocks are impacted directly by the soil thermal state (as well as soil moisture, and inputs from vegetation). For JSBACH, while the vegetation fluxes do not show any noticeable sensitivity to snowfall (Figure S7), the soil carbon has a small but consistent trend towards lower soil carbon in the simulations with increased snow (Figure S8), which – since the other influencing variables have not significantly changed – is most likely due to consistently higher soil temperatures when more snow is present. For JULES, however, any changes in decomposition due to soil temperature are obscured by larger differences of vegetation inputs, particularly for Kytalyk and Samoylov sites (Figure S8), where the vegetation fractions are very different during spinup for the different sensitivity tests, and thus the rate of soil carbon accumulation changes significantly.

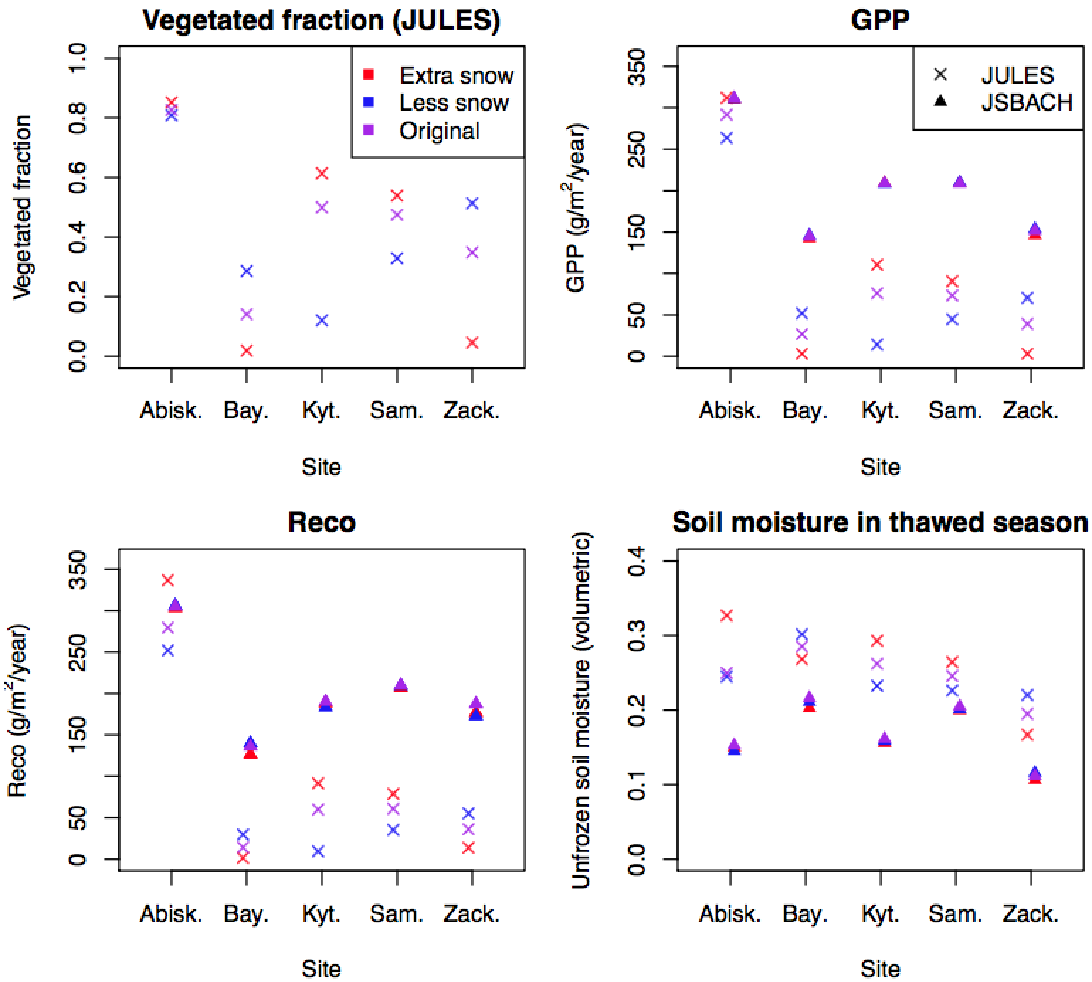

Figure S7 | Impacts of increased/reduced snowfall on soil moisture and carbon-cycle related variables (GPP, ecosystem respiration, and vegetated fraction), in JSBACH and JULES.

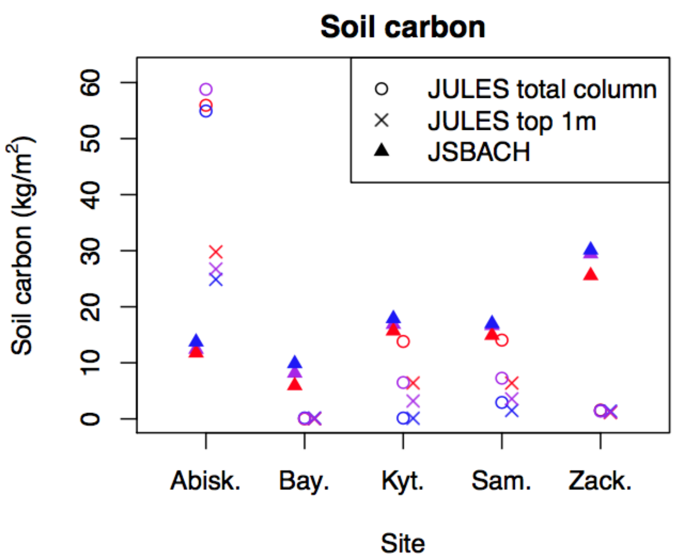

Figure S8 | Impact of increased/reduced snowfall on soil carbon stocks in JSBACH and JULES.

Discussion

Our sensitivity study has shown a high sensitivity of surface soil temperature to a 50% change in snow depth of up to 5ºC or more, seasonally. This is in line with observations (Gisnås et al., 2014). Soil carbon decomposition is sensitive to these soil temperature changes, resulting in lower carbon stores for the warmer simulations in JSBACH (Fig. S8), which is in line with studies such as Lund et al. (2012) which showed that snow affected the carbon budget at Zackenberg by warming the soil and increasing soil respiration. However, the impact of snow on soil moisture is not in line with observed behaviour: in general, more snow should lead to increased soil moisture availability in summer (see for example Litaor et al., 2008). However, in JULES for two of the sites, the summer soil moisture is reduced with additional snowfall, and in JSBACH there are no significant changes. This supports the conclusion that more work is needed on the hydrology schemes in these models. Furthermore, the models are missing some snow-vegetation interactions such as preventing vegetation growth when covered by snow, or protection from damage in winter.

It is also important for the models to better represent the profile of snow thermal conductivity: for example the models do not simulate the low-conductivity 'depth-hoar' layer that can form at the base of the snowpack (Domine et al., 2016). For this, monitoring of snow temperature at different heights can be valuable to improve the models (Barrere et al., 2017). It is also useful to compare snow density in models and observations. For example, recent work shows that including wind compaction is essential to capture high snow density at Samoylov (Gouttevin et al., 2017), and indeed our models show a snow density closer to the 'default' model in Gouttevin et al. (2017), which is too low due to omission of wind effects.

In large-scale modelling, it is certainly important to represent variability in snow depth, which is only coarsely included in land surface models in most cases (e.g. snow depth varies only between surface tiles (Essery et al., 2003)). For recent developments towards this, see for example Gisnås et al. (2014).

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