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1 **The effect of soil on the summertime surface energy budget of a humid**

2 **Subarctic tundra in northern Quebec, Canada**

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ABSTRACT

31 Rising temperatures in the southern Arctic region are leading to shrub expansion and permafrost
32 degradation. The objective of this study is to analyse the surface energy budget (SEB) of a subarctic
33 shrub tundra site that is subject to these changes, on the east coast of Hudson Bay in eastern Canada.
34 We focus on the turbulent heat fluxes, as they have been poorly quantified in this region. This
35 study is based on data collected by a flux tower using the eddy covariance approach and focused
36 on snow-free periods. Furthermore, we compare our results with those from 6 Fluxnet sites in
37 the Arctic region and analyse the performance of two land surface models, SVS and ISBA, in
38 simulating soil moisture and turbulent heat fluxes. We found that 22.4% of the net radiation was
39 converted into latent heat flux at our site, 33% was used for sensible heat flux and about 15% for
40 ground heat flux. These results were surprising considering our site was by far the wettest site
41 among those studied, and most of the net radiation at the other Arctic sites was consumed by the
42 latent heat flux. We attribute this behaviour to the high hydraulic conductivity of the soil (littoral
43 and intertidal sediments), typical of what is found in the coastal regions of the eastern Canadian
44 Arctic. Land surface models overestimated the surface water content of those soils but were able to
45 accurately simulate the turbulent heat flux, particularly the sensible heat flux and to a lesser extent,
46 the latent heat flux.

47 *Significance statement.* Isostatic uplift after the last deglaciation led to the emergence of beaches,
48 which represent a large area in the Canadian Arctic. We studied the surface energy budget of
49 such a beach emerged 6000 years ago in northeastern Canada. Results suggest that evaporation
50 is up to 20% less than at previously studied sites which is explained by sandy soils which retain
51 little moisture despite abundant precipitation. Deployed numerical models showed difficulties in
52 simulating the soil conditions but proved successful in simulating the surface energy budget after
53 manual adjustments of the soil conditions. These simulation difficulties probably apply to other
54 parts of the Canadian Arctic, possibly leading to errors in meteorological and climate forecasting.

55 **1. Introduction**

56 The increase in global air temperature, which is twice as pronounced in polar regions than in
57 all other parts of the world (Chylek et al. 2009), is changing the distribution of vegetation zones
58 (Myers-Smith and Hik 2018). This change is particularly notable in the forest-tundra ecotone
59 (Payette et al. 2001) at the interface between the boreal forest and the Arctic shrub tundra. This
60 region is characterized by a mosaic of forest patches, usually restricted to humid, wind-sheltered
61 areas, surrounded by shrub tundra that occupies the well-drained and wind-exposed sites. Rising
62 temperatures favor the progressive expansion of these forest patches, which lead to the northward
63 expansion of the boreal forest (Tape et al. 2006).

64 The disappearance of shrub tundra at the southern Arctic border is leading to changes on multiple
65 levels. Heat and moisture exchanges at the interface between the land surface and the atmosphere
66 are being altered. As such, the surface energy budget (SEB) as a whole is also changing due
67 to shifts in leaf area index, land surface albedo, and evapotranspiration. These changes then
68 create a feedback affecting the climate at multiple scales by modifying ecosystem processes and

69 weather patterns (Baldocchi and Ma 2013; Pielke et al. 2011). Due to the circumpolar extent of
70 the forest-tundra ecotone, its potential impact on the climate demands our attention.

71 Moreover, the surface energy budget influences the fate of the permafrost as it controls how much
72 heat is stored in or released from the ground. Permafrost in this region has become discontinuous
73 to sporadic and is degrading quickly (Lemieux et al. 2016). In order to model carbon emissions
74 due to permafrost thaw (Schuur et al. 2015), information on the partitioning of the net radiation
75 into sensible, latent, and soil heat fluxes is key. Thus, an improved understanding of the energy
76 balance can ultimately lead to more accurate predictions of permafrost thawing and the resulting
77 greenhouse gas emissions.

78 Due to the harsh conditions and limited accessibility in the Subarctic region, few in-situ obser-
79 vations have been conducted. Available datasets come from sites located in central and western
80 North America, Russia and Europe (Lafleur et al. 1992; Eugster et al. 2000; Langer et al. 2011;
81 Stiegler et al. 2016). The topography of these sites often consists of plains or gentle slopes, which
82 tend to reduce drainage efficiency and provide wetter soils. The latent heat flux at these sites tends
83 to dominate the SEB over the other terms. For example, Eugster et al. (2000) reported latent heat
84 flux to net radiation ratios in the range of 0.44 to 0.72 at several low Arctic sites in Alaska, and
85 values between 0.48 to 0.65 for treeline shrub tundra across central Canada, Alaska, and Russia.
86 One aspect that has so far been poorly documented is the impact of soil properties on the SEB
87 of subarctic sites, as the vegetation tended to be considered the dominant factor on the issue. In
88 the eastern region of the Canadian Subarctic, for instance, soils are often composed of coarse
89 quaternary deposits such as glacial till and marine sediments (Lemieux et al. 2016). These soils
90 are usually very well-drained even on small inclines, limiting water availability and making this
91 region distinct from those of previous studies.

92 In addition to the quantification of surface fluxes, a pressing issue is whether current land surface
93 models (LSMs) such as ISBA (Noilhan and Planton 1989) and SVS (Husain et al. 2016; Alavi et al.
94 2016) are able to adequately simulate the SEB in Subarctic regions. LSMs are used in a variety
95 of applications, such as numerical weather prediction models and terrestrial ecosystem models.
96 Although the surface of the forest-tundra environment is snow-covered for most of the year, the
97 adequate modeling of the SEB during snow-free period is equally important because this period
98 regulates the growth of vegetation and to a larger extent, the water cycle. Studies have highlighted
99 difficulties in modeling dry sites, stating that the models capture energy limitations better than
100 water limitations (Engstrom et al. 2006), thus leading to increased errors for simulated latent heat
101 fluxes.

102 In this study, we use eddy covariance data to analyse the Arctic surface energy balance and
103 examine the performance of two LSMs: SVS and ISBA. SVS will soon replace the old version
104 of ISBA and be used as an operational weather-forecasting model in Canada. This LSM has
105 been proven to be competent at simulating latent and sensible heat fluxes at six sites in arid,
106 Mediterranean, and tropical climates (Leonardini et al. 2020). This makes it particularly interesting
107 to compare whether SVS performs better in simulating northern Canadian regions than the current
108 version of ISBA. This study presents a comprehensive analysis of observed and modeled surface
109 energy fluxes during three consecutive snow-free periods in a forest-tundra ecotone site in northern
110 Quebec in eastern Canada. The objectives of this study are to (i) use eddy covariance observations
111 to identify key factors that influence the energy partitioning at the surface, (ii) determine the
112 interannual variability of the partitioning, (iii) contrast the observations with those from other
113 Subarctic sites across the globe, and (iv) assess the capability of the current LSMs, ISBA and SVS,
114 to simulate surface heat fluxes in this environment.

115 2. Methods

116 a. Study Site

117 The study site is located in the Tasiapik valley close to the community of Umiujaq, on the eastern
118 shore of Hudson Bay in Quebec, Canada (56°33'31"N, 76°28'56"W; Umiujaq in Figure 1a). The
119 valley is 4.5 km long and 1.3 km wide and surrounded by *cuestas* (steep ridges) with a maximum
120 height of nearly 400 m. The lower part of the valley borders the Tasiujaq Lake at an elevation of
121 0 m and is connected to Hudson Bay by an inlet. The vegetation cover in the valley is typical of a
122 forest-tundra ecotone, with shrub and some lichen tundra in its upper region and forest tundra in
123 the lower region. While the dwarf birch *Betula glandulosa* is the dominant vegetation type in the
124 shrub tundra, the prevailing tree species in the forest-tundra are black and white spruces (*Picea*
125 *mariana* [Mill.] BSP and *Picea glauca* [Moench] Voss) (Paradis et al. 2016). Following Gagnon
126 et al. (2019), vegetation in the vicinity of the flux tower could be classified as one of three types:
127 pure lichen cover, small shrubs with lichen understory and a low LAI (presumably <0.5) and small
128 shrubs with moss understory. Lichens covers 20 to 30% of the surface while small shrubs with
129 differing understory make up the rest (See Figure A3).

130 In the valley, permafrost is discontinuous to sporadic (Lemieux et al. 2016) with rapidly degrading
131 permafrost mounds that are ≈ 100 m in diameter (Fortier and Aubé-Maurice 2008). Close to the
132 study site, permafrost is only present to the south at a distance of about 20 m. The active layer
133 thickness is about 1.5 to 2 m. This active layer depth together with the small inclination of a
134 few degrees of the surface presumably provides a rather undisturbed drainage of precipitation.
135 There are no complete long-term meteorological records available for Umiujaq. Instead, recent
136 observations from the research station in the Tasiapik valley were used to estimate climatological
137 conditions starting in 2012 for air temperature and from 2015 for precipitation. Umiujaq exhibits a

138 Subarctic climate with a mean annual temperature of -4.0°C . Their winters are long and cold and
139 their summers are cool. There is usually continuous snow cover from late October to early June
140 with a maximum thickness between 0.6 m and 1.0 m. Monthly temperatures and precipitation are
141 shown in Figure 2.

142 Weather patterns are strongly influenced by Hudson Bay to the west and whether or not it is
143 covered with ice. Hudson Bay usually remains frozen until mid-June, and then stores heat in
144 summer that is released in the fall. This results in relatively lower air temperatures in summer and
145 higher air temperatures in the fall. The average total annual precipitation is about 800 - 1000 mm,
146 out of which about 50% falls as snow and the precipitation cycle shows a strong correlation with
147 the freezing of Hudson Bay. After the freeze-up in mid to late December, precipitation remains
148 low from January to May/June. Precipitation subsequently increases and peaks in late summer and
149 autumn. Fog is common throughout the summer (Robichaud and Mullock 2001). Advection fog
150 forms when warmer air moves over the cold Hudson Bay, mostly in July and August.

151 The geology of the valley is thoroughly described in Lemieux et al. (2020). The bedrock is
152 covered by a 10- to 50-m thick layer of moraine deposits and fluvio-glacial sediments. Above the
153 fluvio-glacial sediments, there is a layer of marine sediments (mainly silt) that is $\approx 30\text{-m}$ thick, but
154 it does not cover the entire valley. The topsoil layer is sand (littoral and intertidal sediments, see
155 Figure 3), with some limited exceptions.

156 Gagnon et al. (2019) found that the soil covered with lichen is made up almost entirely of sand
157 (90% for the first 5 cm and almost 100% for deeper layers up to 30 cm). They reported a soil
158 organic content of about 1.4 kg m^{-2} under lichen and 4.3 kg m^{-2} for medium height shrubs tundra
159 with lichen understory. Most of the soil organic content is found in the first 2 cm for the lichen
160 covered areas and 5 cm for medium shrub covered areas.

161 *b. Instrumental Setup*

162 In June 2017, a 10-m tower was equipped with a sonic anemometer and a CO₂/H₂O gas analyzer
163 that was installed 4.2 m above the ground (IRGASON, Campbell Scientific, USA), facing eastwards
164 towards a plateau with a small slope of about 3° (Figure 1 b). The tower holding the instrument
165 and a steep ridge are to the west, in the opposite direction.

166 The tower also features general meteorological sensors, such as temperature and relative humidity
167 (model HMP45, Vaisala, Finland) and wind direction/speed (model 05103, R.M. Young, USA)
168 probes. Some 10 m west of the flux tower, another meteorological station was installed in 2012,
169 which also measures basic environmental variables such as air temperature and wind speed. It
170 is equipped with a 4-component radiometer (CNR4, Kipp and Zonen, The Netherlands). In the
171 proximity of the two stations, soil temperature and water content (5TM, Meter Group, USA)
172 are measured at several locations and under various vegetation types directly at the tower and
173 several meters distance, at 5 depths from 5 cm to 50 cm, in order to capture spatial variability.
174 Precipitation is also measured next to the tower with a T200B precipitation gauge (GEONOR,
175 USA) equipped with a single Alter shield. Raw precipitation observations were corrected for
176 wind-induced undercatch following Kochendorfer et al. (2018). Flux data was recorded using a
177 CR3000 datalogger while the other data was recorded using a CR1000 datalogger (both Campbell
178 Scientific, USA). The site was visited twice a year for maintenance and data collection.

179 *c. Data Processing*

180 Raw 10-Hz eddy-covariance data were processed using the EddyPro[®] (version 7.0.3; Li-COR
181 Biosciences, USA) software package. This software includes multiple corrections and QA/QC
182 procedures such as turbulent fluctuation detrending based on a running mean, covariance maxi-
183 mization, density fluctuation compensation (Webb et al. 1980), and analytic correction of high-pass

184 and low-pass filtering effects (Moncrieff et al. 1997). Turbulence data were rotated into a stream-
 185 line coordinate system using the sector-wise planar-fit method (Wilczak et al. 2001). The random
 186 uncertainty quantification approach from Finkelstein and Sims (2001) was used to detect outliers,
 187 spikes, and other artifacts. The 0-1-2 quality scheme from Mauder et al. (2013) was applied to
 188 the data, and only segments that were flagged as 0 and 1 were kept in the study. These values
 189 correspond to good- and medium-quality observations, which are both suitable for the analysis of
 190 seasonal energy budgets.

191 Ground heat flux was computed using a time series of a measured soil temperature profile
 192 $T(z_1, t), T(z_2, t)$ and $T(z_3, t)$ at three depths $z_1 < z_2 < z_3$ and the one-dimensional heat equation

$$\frac{\partial T(t, z)}{\partial t} = D \frac{\partial^2 T(t, z)}{\partial z^2}, \quad (1)$$

193 where D is the thermal diffusivity (Westermann et al. 2009). As the thermal diffusivity is defined
 194 by the thermal conductivity K and the heat capacity c_h by $D = K/c_h$ we calculated D through
 195 K and c_h using empirical equations which are based on the soil density and water content (Chen
 196 2008; Bittelli et al. 2015). Then, the heat equation was solved numerically in Python using two
 197 of the temperature sensors in the soil ($T(z_1, t)$ and $T(z_3, t)$) as Dirichlet boundary conditions. The
 198 result is a temperature distribution in the considered spatial domain. It was checked against the
 199 third soil temperature measurement $T(z_2, t)$ for correctness ($R^2 = 0.99$, RMSE = 0.21°C). Then,
 200 the temperature gradient at the upper boundary z_1 was calculated. The ground heat flux at depth
 201 z_1 is then given by the Fourier equation:

$$G_{z_1} = K \left. \frac{\partial T(z, t)}{\partial z} \right|_{z=z_1} \quad (2)$$

202 The heat storage change G_s in the soil is accounted for by taking the change of the mean temperature
 203 of the soil above z_1 . The mean temperature is obtained by extrapolating the temperature distribution
 204 given by the heat equation. The final ground heat flux is then given by $G = G_{z_1} + G_s$. This method

205 was chosen as measurements under different vegetation types were available making it possible to
206 analyze the spatial variability. Furthermore, problems have been reported of alternative methods
207 such as using heat flux plates (Ochsner et al. 2006).

208 Post-processing was also necessary to sort out the remaining outliers and to fill data gaps. This
209 was achieved using the program PyFluxPro (Isaac et al. 2017). This program has six processing
210 levels, uses EddyPro output files as input and produces a continuous time series for all fluxes. For
211 the first three processing levels, data are read, quality controlled, and finally auxiliary measurements
212 are merged when gaps are present. The quality control includes (i) range checks based on user-
213 defined limits, (ii) spike detection, (iii) manual removal for specific dates and (iv) data rejection
214 based on other variables. For the data rejection, CO₂ and H₂O signal strengths from the IRGA as
215 well as internal error codes from both the sonic anemometer and the IRGA were used to remove
216 erroneous flux data. For the fourth processing level, meteorological variables were gap filled with
217 ERA5 data, a reanalysis product from the European Centre for Medium-Range Weather Forecast.
218 These data provide hourly estimates for various meteorological and soil variables beginning in
219 1979 at a spatial resolution of 30 km ([https://www.ecmwf.int/en/forecasts/datasets/
220 reanalysis-datasets/era5](https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5)). Each variable was bias-corrected using a linear fit between ERA5
221 and flux tower observations during periods when both were available.

222 Finally, the fluxes were gap-filled using a self-organizing linear output map (SOLO), a type of
223 artificial neural network (ANN) (see Hsu et al. (2002) and Abramowitz (2005)). SOLO requires a
224 set of environmental drivers such as air temperature, radiation and humidity, as well as the fluxes
225 as inputs. SOLO first constructs relationships between the environmental drivers by applying an
226 ANN-equivalent of a principal component analysis. It then uses an ANN-equivalent of a multiple
227 linear regression to make connections between the drivers and the fluxes. ANN together with

228 marginal distribution sampling (MDS, Reichstein et al. (2005)) were shown to be the best choices
229 for gap-filling flux data (Moffat et al. 2007).

230 On average, 56.1% of the hourly values of the sensible heat flux (2017: 50.8%; 2018: 63.5%:
231 2019: 54.2%) and 47.8% of the latent heat flux (2017: 46.7%; 2018: 54.2%; 2019: 42.4%) passed
232 all quality control steps.

233 This study focused on the summer period, defined here as snow-free periods. It started on the
234 date that snow disappeared and ended when the first snowfall occurred in order to avoid any effects
235 associated with the presence of snow on the ground. In order to facilitate inter-annual comparisons,
236 we applied these criteria to the three measurement years and defined common start and end dates:
237 July 1st to September 5th (66 days).

238 *d. Comparison Sites*

239 To give the energy balance measurements from the Umiujaq site context, data from six other flux
240 tower sites across the Arctic were selected from the FLUXNET (Pastorello et al. 2020) database.
241 These sites were located in Canada, the United States, and Russia and were selected within latitudes
242 ranging from 56°N to 70°N (Figure 4). FLUXNET is a global network of micrometeorological
243 tower sites that collects eddy covariance observations. We have chosen sites that show some
244 similarities concerning the vegetation (shrubland) and are not considered as permanent wetland.

245 The comparison sites cover a variety of biomes located in Arctic and Subarctic regions (Figure
246 4). Their characteristics are detailed in Table 1. Mean annual temperatures range from -14.3°C
247 in eastern Russia (RU-COK), by far the coldest site, to -1°C in Alaska (US-EML). Mean annual
248 precipitation varies considerably, from the dry Alaskan site (122 mm; US-AN) to more humid sites
249 in western Russia (538 mm; RU-VRK) and eastern Canada, with Umiujaq being by far the most
250 humid site (800 - 1000 mm).

251 Vegetation covers at all the sites, except for RU-COK, is classified as open shrubland. Those
252 sites are covered by all evergreen or deciduous woody vegetation less than 2 m tall at coverage
253 between 10% and 60%. The vegetation at RU-COK is closed shrubland, given that the canopy
254 cover exceeds 60%.

255 Data processing was similar to that described in section 2c. Quality assurance and quality control
256 methods follow those in Pastorello et al. (2014), where meteorological variables were gap-filled
257 using ERA-Interim reanalysis data (Vuichard and Papale 2015), and flux data were gap-filled with
258 marginal distribution sampling (Reichstein et al. 2005).

259 As for most of the eddy covariance sites worldwide, all studied sites experienced non-closure of
260 the energy budget. Since not all terms of the energy budget were available for each site and given
261 the ongoing debate on how to best distribute the residual energy budget (De Roo et al. 2018), no
262 corrections were applied to any of the energy budgets.

263 *e. Model Description*

264 To simulate the SEB, two LSMs were used on offline mode: ISBA and SVS. ISBA was included
265 in SURFEX v8.1 (SURFace EXternalisée), a modeling platform developed by Météo-France
266 (<http://www.umr-cnrm.fr/surfex/>) that comprises several other models for various types of
267 land and water surfaces. SVS is a surface model built by Environment and Climate Change Canada
268 for operational hydrometeorological applications.

269 Both models require meteorological data as inputs. These data can be obtained from an atmo-
270 spheric model when operated in coupled mode, or from observations in stand-alone mode (offline).
271 In this study, the stand-alone option was used to ensure the quality of the driving data and to avoid
272 errors that may be introduced by the atmospheric model itself.

273 1) ISBA-MEB

274 ISBA (Noilhan and Planton 1989) simulates all water and energy exchanges between the atmo-
275 sphere, vegetation, and soil. The water and energy budgets of soil are computed by solving the
276 one-dimensional Fourier law and a mixed-form of the Richards equation explicitly (Boone et al.
277 2000; Decharme et al. 2011). The soil is therefore divided into layers of increasing depth down
278 to 12 m. Each layer has specific hydraulic and thermal properties, which are either calculated
279 internally based on the soil texture or forced externally. The phase change between the solid and
280 liquid phases of water in the soil is computed using the Gibbs free-energy method (Boone et al.
281 2000; Decharme et al. 2016). This method calculates the temperature that governs the phase
282 change based on the soil texture. During any phase-change, the total amount of water in each layer
283 remains unchanged. Therefore when the ice content increases, the water content decreases by the
284 same amount, and vice versa.

285 Vegetation characteristics can be automatically selected from 19 available categories us-
286 ing the site coordinates and ECOclimap ([https://opensource.umr-cnrm.fr/projects/
287 ecoclimap-sg/wiki](https://opensource.umr-cnrm.fr/projects/ecoclimap-sg/wiki)), or they can be specified by the user (vegetation type, LAI, canopy height
288 and albedo).

289 In order to specifically include vegetation in ISBA, Boone et al. (2017) recently added what is
290 called the multi-energy balance (MEB) parametrization. MEB largely follows the representation
291 of vegetation detailed in the Rossby Centre Regional Climate Model (Samuelsson et al. 2011).
292 The most important additions (relative to the standard version of ISBA) are the presence of canopy
293 turbulence, a new radiation transfer scheme, and a litter layer. Within MEB, the sensible and latent
294 heat fluxes between multiple components are calculated separately and then combined to form the

295 final heat flux. These components are the vegetation, the bare ground, a ground-based snowpack,
296 and the canopy air space.

297 2) SVS

298 The Soil, Vegetation and Snow scheme (SVS) (Husain et al. 2016; Alavi et al. 2016) is partly
299 based on an early version of ISBA (Bélair et al. 2003a,b), in which a force-restore method is used
300 for the energy budget, as well as a single layer interaction between the canopy and the atmosphere.
301 A notable improvement of this model is the different land surface tiling and vertical layering, as
302 well as having separate energy budgets for bare soil and (low and high) vegetation tiles. Another
303 difference between the two LSMs concerns soil hydrology. The earlier version of ISBA monitored
304 soil moisture at a superficial layer and the rooting-depth layer. In SVS, a certain soil water content
305 is attributed to each of the N soil layers and its evolution is modeled based on Darcian flow.
306 However, in SVS soil temperature is only defined at the soil surface and the associated variation is
307 determined using the single-layer approach mentioned above.

308 3) METEOROLOGICAL FORCING DATA

309 Both ISBA and SVS require the same input data: air temperature, specific humidity, wind speed,
310 incoming shortwave and longwave radiation, atmospheric pressure, and (solid and liquid) precipi-
311 tation rates. Observational data for these variables are available from 2012, except for atmospheric
312 pressure, for which the data start from June 2017 and precipitation where measurements started in
313 2015. Before this date, ERA5 atmospheric pressure and precipitation was used. A simple linear
314 regression was conducted between the measured data and the ERA5 data in order to correct for the
315 atmospheric pressure data from ERA5 and make it consistent with *in situ* observations. A fixed
316 threshold of 0°C was applied to separate the solid and liquid phases of precipitation. This simple

317 threshold was implemented because we were only studying the summer period, and thus a more
318 elaborate phase partitioning strategy was unnecessary.

319 4) MODEL SETUP

320 Both models were set up in a similar fashion to maximize the comparability of the results.
321 However, we have opted to use the soil discretization that is most commonly used for each model
322 and which happen to be different from each other. For ISBA, the soil column has a depth of 12 m
323 and is divided into 20 layers of increasing depth. The soil column for SVS is 10-m-deep and has 7
324 layers, also of increasing depth.

325 Models were initialized with a spin-up to ensure an equilibrium of soil moisture and temperature.
326 We used 5 complete years of observations (2012-2017) as the spin-up, which was sufficient for
327 both models to reach an equilibrium. We analyzed the melt and onset of the snow cover in the
328 model to make sure that snow did not interfere with our study period. We used a time step of 1
329 hour for both models.

330 The soil composition that was used for both models was 95% sand and 5% silt (Gagnon et al.
331 2019). Hydraulic and thermal parameters were for the most part calculated by the models them-
332 selves, based on the soil composition provided. A modified version of both ISBA and SVS models
333 were also tested where hydraulic parameters were adjusted. The adjusted parameters were the
334 saturated soil water content and the field capacity, both of which were based on soil water content
335 observations at several depths. For both parameters, a profile that decreased with increasing soil
336 depth was selected. The saturated soil water capacity was set to $0.3 \text{ m}^3 \text{ m}^{-3}$ at the surface and
337 $0.08 \text{ m}^3 \text{ m}^{-3}$ at depths of 1 m and more. The field capacity at the surface was $0.18 \text{ m}^3 \text{ m}^{-3}$ in both
338 models and was decreased to a minimal value of $0.05 \text{ m}^3 \text{ m}^{-3}$ at soil layers deeper than 1 m. For

339 comparison, we also ran both models without changing any of the parameters. This is referred to
340 as the default version of the model.

341 5) MODEL EVALUATION

342 Only a subset of the whole dataset was used for model evaluation. We selected data with energy
343 budget closures (EBC) between 0.8 and 1.2, where EBC is defined as:

$$EBC = \frac{H + LE}{R_n + G}. \quad (3)$$

344 In equation 3, H and LE are the turbulent fluxes of sensible and latent heat, G is the ground heat flux
345 and R_n is the net radiation. We applied this procedure to minimize possible measurement errors
346 and assure comparability between the models, as their energy budget closures are equal to 1 by
347 definition. In order to evaluate and visualize the performance of ISBA and SVS in simulating H and
348 LE , we used Taylor diagrams (Taylor 2001), which can effectively illustrate several performance
349 metrics on the same plot. The normalized standard deviation for a dataset from its mean is given
350 by the radial distance from the origin at (0|0). The angle between the point of interest, and the
351 abscissa defines the correlation of the model with the observations. The normalized RMSD of
352 the model is shown by the distance to the point indicated as Observation at (1|0). The standard
353 deviation and RMSD are both normalized using the standard deviation of the observations.

354 3. Results

355 a. General Conditions in Umiujaq

356 Table 2 shows the conditions during the three summers of interest at the Umiujaq site. These
357 conditions varied considerably in terms of air and soil temperatures, with summer 2019 being the
358 warmest and summer 2018 being the coldest. Also, precipitation differed considerably, although

359 there was no dry period in neither of the three summer, with a total of 237 mm in 2017 and 369
360 mm in 2019. Another marked difference, particularly in the beginning of summer, is the amount
361 of fallen snow from the preceding winter. Summer 2018 was delayed by more than two weeks due
362 to an exceptionally thick snow cover (≈ 1.4 m) from the previous winter. This was not the case
363 for the other two summers, as the maximum snow depths from those winters were more typical
364 (0.6-1.0 m).

365 Studying the one site over three summers allows us to investigate an interesting range of inter-
366 annual variability, with total seasonal precipitation varying by up to 132 mm (factor of 1.56) and
367 mean air temperatures varying by up to 2°C . In general, we can classify the summer of 2017 as
368 dry, the summer of 2018 as cold, and the summer of 2019 as warm and humid.

369 *b. Evolution of the Surface Energy Budget*

370 Figure 5 illustrates the cumulative values of net radiation R_n , sensible heat flux H , latent heat
371 flux LE , ground heat flux G , and residual flux for each of the three summers. Only data with an
372 hourly EBC between 0.2 and 1.5 were used to omit obvious outliers. Our results indicate that H
373 plays a more important role than LE in dissipating the net radiation. Figure 5 shows that over
374 the three summers, H accounted for about 33% of the net radiation while LE represented roughly
375 22.4%. The ground heat flux played a far less important role, making up around 15% of the net
376 radiation. The residual flux was 29.5% on average and thus comparable to H and LE . For this
377 reason, a sensitivity test confirmed that the relative importance of H and LE does not vary with
378 the applied EBC criterion. The average Bowen ratio BR (H/LE) fluctuated between 1.4 and 1.54
379 for and EBC ranges between 0.2 to 1.5 and 0.8 to 1.2. The energy budget closure was best in 2019,
380 with a mean closure of 76.7%. Indeed, in 2019, there were more frequent winds from areas less
381 affected by topography, and therefore more favorable to the use of the eddy covariance technique.

382 Although conditions varied considerably between the three summers, the general seasonal trends
383 for H and LE were comparable (see Figure A2). H exhibited a strong increase directly after
384 snowmelt and peaked in late June, with maximum values of 300 W m^{-2} . It then decreased
385 throughout the summer before settling on a magnitude comparable to that of LE at the end of the
386 summer. Unlike the clear peak and fall seen for H , LE did not follow a clear seasonal trend (Figure
387 A2). It stayed rather constant throughout the season and shows only a small peak towards the middle
388 of the summer in the last two years, reaching maximum values of nearly 200 W m^{-2} . Consequently,
389 BR is greater than 1 at the beginning of summer and subsequently decreases, dropping below 1 at
390 the end of the summer.

391 By comparing the three years with one another, it is evident that a delayed start to the summer
392 season in 2018 did not have a large impact on the overall patterns of turbulent heat fluxes. H in
393 2019 showed a similar pattern to 2017, only differing in the second half of July due to an extended
394 warm spell (Figure A1). In summary, it appears that H evolves in a consistent way and follows the
395 same trend after snowmelt, with some differences as a result of meteorological conditions.

396 *c. Evapotranspiration and Soil Properties*

397 In this section, we look at one of the limiting factors for evapotranspiration: water availability.
398 To further investigate whether a lack of water restricts the latent heat flux, Figure 6 displays the
399 precipitation, evapotranspiration (ET), and soil water content in summer 2017 (summer 2018 and
400 2019 are available in the supplementary material).

401 Compared to other sites at similar latitudes, Umiujaq experiences high levels of precipitation
402 (Figure 4), meaning there is no general lack of water supply at the surface. Rainfall in Umiujaq is
403 also fairly equally distributed over the entire summer (Figure 2), therefore no distinct dry or wet
404 periods were observed for any of the three summers (the corresponding figures for 2018 and 2019

405 are available in the supplementary material). However, as the red curve in Figure 6 indicates, the top
406 soil volumetric water content under the lichen θ stayed rather low even during precipitation events,
407 with peaks below $0.15 \text{ m}^3 \text{ m}^{-3}$ in 2017 and $0.2 \text{ m}^3 \text{ m}^{-3}$ for the three summers (the corresponding
408 figures to Figure 6 are available in the supplementary material). Following rainfall, θ dropped
409 rapidly to well below $0.1 \text{ m}^3 \text{ m}^{-3}$, and at times dropped to $0.05 \text{ m}^3 \text{ m}^{-3}$. This demonstrates that
410 even though there is abundant water in the form of rain, the soil can only hold small amounts of the
411 water. This is because the water drains rapidly due to the large particle size (Gagnon et al. 2019)
412 of the soil and the associated high hydraulic conductivity.

413 Two post-rain periods in Figure 6, one at the end of June and the other in the beginning of July,
414 very clearly demonstrate the influence of soil water content on ET. Approximately two days after the
415 drop in surface water content, ET declined from about 2 mm day^{-1} to less than 0.5 mm day^{-1} and
416 then rose again in response to the next rainfall event. The presence of a ≈ 2 -day lag time between
417 both variables is particularly visible until mid-July as illustrated by the 3-day running mean of ET.
418 The running mean of ET also highlights the influence of soil moisture on ET. Consistently, the
419 opposite behaviour was observed for BR, which increased after the rainy periods.

420 *d. Comparing Umiujaq to Other High Latitude Sites*

421 Figure 7 illustrates the mean net radiation and sensible and latent heat fluxes for the comparison
422 sites (see Table 1), exclusively during daytime ($R_n > 0$). The observed net radiation peaked at the
423 beginning of the study period or shortly after, and subsequently decreased. Maximum net radiation
424 values tend to drop at higher latitudes (CA-NS recorded more than 300 W m^{-2} at 56°N and the
425 maximum mean value at RU-COK was 180 W m^{-2} at nearly 71°N).

426 The general behavior of H and LE can be roughly divided into three groups: LE that is typically
427 higher than H (BR<1; RU-Cok, US-ICH, RU-VRK, US-EML), LE and H that are approximately

428 equal ($BR \approx 1$; US-AN and CA-NS) and finally LE that is lower than H ($BR > 1$; CA-UM (Umiujaq)).
429 We argue that this general behavior of H and LE is usually influenced by three main factors: (i)
430 meteorological conditions, (ii) vegetation cover, and (iii) soil type. Factors (i) and (ii) are discussed
431 in the remainder of this section, while factor (iii) is discussed in section 3e.

432 (i) There were considerable differences in the mean annual precipitation for all sites analyzed,
433 ranging from 140 mm (US-AN) to more than 800 mm in Umiujaq. However, there is no correlation
434 between the annual precipitation and the Bowen ratio or the magnitude of the latent heat flux. This
435 can be seen at the drier sites in Alaska, which have BRs that are similar to or lower than the much
436 more humid RU-VRK site. Moreover, the highest BR values were observed at Umiujaq, the most
437 humid site. Thus, the annual mean precipitation does not limit LE , although in the aftermath of a
438 rain event, LE levels will rise as seen for Umiujaq before.

439 (ii) Vegetation cover also alters the BR through transpiration. Through the shade they provide,
440 shrubs cool ground temperature by reducing incoming radiation and consequently influence the
441 relative importance of H and LE . The relative importance G was not analyzed as it was not
442 available for all sites. Even though all the sites were classified as shrubland, the composition of
443 the vegetation at these sites varied. Table 1 provides a brief overview of the vegetation at each
444 site, from the sparse tussock tundra at US-AN with only a few shrubs, to the CA-NS site's dense
445 shrub vegetation with small trees. However, in regards to precipitation, no correlation between the
446 vegetation type and the Bowen ratio was found.

447 *e. Soil Types*

448 As discussed above, the influence of precipitation and vegetation on seasonal LE and its behaviour
449 relative to H is limited. This leaves soil as the dominant factor that controls the BR at Umiujaq.
450 Figure 8 shows the approximate position of each site in a soil texture diagram. The estimate for

451 Umiujaq is based on Gagnon et al. (2019), while the others are based on the description of the soil
452 from the references presented in Table 1.

453 Although the positioning of each site within the soil texture triangle is highly uncertain, there is
454 a clear distinction between Umiujaq and the other sites: all the other sites exhibit more clay and
455 silt-rich soils compared to Umiujaq. Sand-rich soils are known to hold less water, drain quicker, and
456 are thus often drier compared to soils rich in clay and silt. For instance, the mean soil moisture for
457 the sites US-ICH ($0.39 - 0.52 \text{ m}^3 \text{ m}^{-3}$) and US-AN ($0.56 \text{ m}^3 \text{ m}^{-3}$) are considerably higher than for
458 Umiujaq ($0.1 - 0.2 \text{ m}^3 \text{ m}^{-3}$). According to Clapp and Hornberger (1978), the saturated volumetric
459 water content of sand is $\approx 0.1 \text{ m}^3 \text{ m}^{-3}$ less than for clay or silty clay. The largest difference is in the
460 saturated hydraulic conductivity: $1.056 \text{ cm min}^{-1}$ for sand compared to $0.0062 \text{ cm min}^{-1}$ for silty
461 clay (Clapp and Hornberger 1978).

462 Permafrost, which forms an impermeable layer, is also an important factor at each site (see Table
463 1). Sites that have a shallow active layer, defined as the soil layer that thaws in summer, have a
464 topsoil layer that is often wet, thus favoring *LE*. At the majority of the comparison sites, the active
465 layer is thin, ranging from 15 cm to 70 cm. At Umiujaq however, there is discontinuous/intermittent
466 permafrost and an active layer that is quite deep (100 - 150 cm), which favors the percolation of
467 surface water.

468 *f. Model Comparison*

469 1) SOIL MOISTURE

470 Results from the previous sections highlight a particularity of the Umiujaq site, that despite its
471 abundant precipitation, the presence of sandy soil produced a low latent heat flux. In light of this,
472 the ability of both models to simulate soil water content was first examined. As mentioned in
473 section 2e, two runs were performed for each model. A default run was performed with no changes

474 to the parameters of the model and another with adjusted hydraulic parameters was performed
475 in order to improve soil simulations. Figure 9 compares the observed soil water content at two
476 different depths with the outputs from ISBA and SVS with default parameters and Figure 10 is the
477 corresponding figure for the model run with adjusted parameters.

478 For the default simulation, only the soil composition was input into the models. ISBA grossly
479 overestimated the soil water content at 6 cm by a margin of $0.2 \text{ m}^3\text{m}^{-3}$ during peak runoff and
480 produced results that were comparable to the observations about one month later in the simulation.
481 Even after the soil water content dropped to a range that was comparable to the observations, the
482 model still slightly overestimated the soil water content (around $0.02 \text{ m}^3\text{m}^{-3}$). The observed soil
483 water content at 6 cm and 50 cm differed considerably, while only minor differences between these
484 two layers were observed in the estimates produced by the two models. The surface water content
485 at 6 cm was overestimated more by SVS than by ISBA, but the default simulation revealed more
486 pronounced differences between values at 6 cm and 50 cm.

487 After adjustments were made (see section 2e), the ISBA outputs more closely matched the obser-
488 vations. When applying the modified soil properties, the simulated SVS results were slightly less
489 accurate than ISBA at the surface when compared to the observations. SVS generally underesti-
490 mated the soil water content by about $0.03 \text{ m}^3\text{m}^{-3}$, while it showed high peaks during rain periods.
491 At 50 cm in depth, SVS produced results that were comparable to ISBA and to the observed value.
492 To simulate the turbulent heat fluxes, the adjusted models were used.

493 2) TURBULENT FLUXES

494 Figures 11 and 12 show observations compared with simulated sensible and latent heat fluxes
495 from ISBA and SVS, respectively.

496 ISBA was able to accurately simulate H with the exception of a few high values that reached up
497 to 400 W m^{-2} (Figure 11 b). The amount of scatter was generally quite low along the observed
498 energy scale. Color-coding revealed that model performance was somewhat dependent on the time
499 of the day for both models. Nighttime fluxes simulated by ISBA (light and dark blue dots) were
500 systematically overestimated, but only by about 10 W m^{-2} . Also for ISBA, there was a slight
501 tendency to overestimate H in the afternoon (14-18), while at the other times of the day, the points
502 were scattered more evenly around the 1:1 line. Otherwise, the simulated fluxes closely followed
503 the observed values. SVS simulations of H were more challenging. While the distribution of the
504 points show a slightly more positive trend, SVS still adequately predicted H on average over the
505 entire energy span, except for high values over 250 W m^{-2} , similar to ISBA. The mean bias for
506 ISBA was 11.0 W m^{-2} while it was 2.9 W m^{-2} for SVS. The problem of overestimated nighttime
507 fluxes that was observed with ISBA simulations did not occur with SVS. The modeled surface
508 temperature compares favourably to the observations ($R^2 = 0.85$ and 0.9 , for ISBA and SVS), but
509 has a positive bias in the case of ISBA (bias = 1.5°C) and a negative bias in the case of SVS (bias
510 = -1.3°C).

511 Neither model was able to simulate LE as accurately as H , with the most notable problem for
512 both models being the nighttime fluxes (Figure 12). ISBA rarely simulated very small negative
513 fluxes, while SVS did not simulate them at all, despite the observational dataset presenting a regular
514 occurrence of negative fluxes. A negative LE indicates the condensation of water vapor onto the
515 soil surface, and this process seems completely absent from the models. Apart from this, both
516 models yielded a data point distribution that exhibited a more positive trend compared to their H
517 simulations. For ISBA, the bias was 5.3 W m^{-2} and for SVS, it was 11.1 W m^{-2} . Particularly,
518 patterns with very low simulated fluxes and high observed fluxes were more frequent for LE than
519 for H . The q-q plot for ISBA showed only minor deviations from the 1:1 line for fluxes with

520 an amplitude of less than 100 W m^{-2} . In the region above 100 W m^{-2} , the number of points
521 dramatically decreased, with only a few points responsible for the high deviations seen in the q-q
522 plot. For SVS, the datapoints diverge from the 1:1 line at a lower observed flux ($\approx 70 \text{ W m}^{-2}$) and
523 the slope following the divergence is higher than for ISBA. Thus, a significant portion of the *LE*
524 data points were overestimated, with deviations as large as 150 W m^{-2} for observed fluxes in the
525 100 W m^{-2} region.

526 The left side of Figure 13 shows that ISBA performed slightly better in simulating *H* than SVS.
527 The correlation (correlation coefficient of 0.91 for SVS versus 0.97 for ISBA) and the normalized
528 RMSD (0.44 for SVS versus 0.27 for ISBA) show better results for ISBA than for SVS. Also, the
529 standard deviation for ISBA was similar to that of the observations, and ISBA slightly overestimated
530 the variability in *H*, while the overestimation of SVS is a bit higher (1.08 for SVS versus 1.02 for
531 ISBA). These results are consistent with the more dispersed distribution of the data points in the
532 scatter plot that indicate the underestimation of *H* by SVS.

533 The Taylor diagram on the right of Figure 13 illustrates the relatively poor simulation performance
534 of the two models for *LE* compared to *H* which affected both models to a similar extent. (ISBA:
535 correlation coefficient of 0.84, standard deviation of 1.2 and normalized RMSD of 0.64; SVS:
536 correlation coefficient of 0.74, standard deviation of 1.12 and normalized RMSD of 1.64)

537 **4. Discussion**

538 *a. Energy Balance Closure and Measurement Error*

539 A typical issue with eddy covariance measurements is the lack of energy budget closure (Foken
540 2008). Throughout most meteorological conditions, the energy budget remained unclosed at the
541 Umiujaq site (on average 70.5%). However, certain wind directions can lead to better closure than

542 others (see Appendix Figure A4). The most probable reasons for the SEB imbalance at this site
543 are: (I) the complex topography and (II) the mismatch between the footprint areas of the different
544 flux sensors.

545 (I): As seen in Figure 1b, the flux tower was located to the east of a steep ridge. The cliff acted as
546 a backward-facing step, potentially leading to flow detachment in the presence of westerly winds.
547 The eddy covariance approach is based on the assumption that there is a surface located upstream
548 that is flat and homogeneous, which is not the case here. However, in recent years, we have
549 noticed an increase in studies conducted in non-ideal terrain (e.g. Geissbühler et al. (2000); Hiller
550 et al. (2008); Etzold et al. (2010); Nadeau et al. (2013); Stiperski and Rotach (2015)). However,
551 considering that the closure of the energy balance exceeded 55% for westerly wind direction
552 (compare Figure A4), most of these data were retained. Foken (2008) argues that in heterogeneous
553 terrains, large eddies can substantially contribute to the energy exchange when generated at the
554 boundaries between different land uses that are excluded from the flux measurements.

555 (II): The footprint of the eddy covariance system naturally varies depending on wind speed and
556 direction, covering an area of up to a few thousand square meters whereof 20 - 30% is dominated
557 by lichen, and the rest by shrubs (Figure A3). Radiation measurements cover areas on the order
558 of a few square meters, whereas the ground heat flux measurement is a point measurement. Thus,
559 radiation and ground heat flux measurements that do not represent the flux footprint inevitably,
560 which leads to an unbalanced budget. The measurements of R_n and G do not match those of H
561 and LE taken in the footprint area, which may therefore lead to a budget imbalance. The radiation
562 and ground heat flux measurements were derived from lichen-covered surfaces, but the footprint
563 of the eddy covariance system covered a mixture of lichen and shrubs of varying heights, with
564 either lichen or moss understory (Gagnon et al. 2019). This was however, difficult to quantify as
565 knowledge on the distribution of each vegetation type would have been required for every 30-min

566 flux footprint area. However, we observed a worse energy budget closure (50% to 65%) for periods
567 during which winds blew in from the south, where shrubs tended to be taller and the topography
568 was more complex.

569 To explore the spatial variability of the surface albedo around the flux tower, point measurements
570 of spectral albedo were made over the two dominant vegetation types (Belke-Brea et al. 2019),
571 lichen and shrubs. The results showed slightly higher albedo levels over lichen, on the order of a few
572 percentage points over the entire spectral range. Additionally, the ground heat flux was calculated
573 under different vegetation types and only minor differences were observed, which alter the overall
574 EBC by 1 or 2%. The 5TM probe did not cover the organic layer at the surface thus an additional
575 error of the ground heat flux is probable. This is because the hydraulic and thermal properties
576 of the organic layer differ from those of the mineral soil. Therefore, although the measurement
577 footprint mismatch certainly contributes to the energy budget imbalance, it is not the sole cause.

578 Furthermore, measurement errors possibly contributed to the degradation of the energy budget.
579 Mauder et al. (2013) reported errors of 10%-15% for the eddy covariance data for flags 0 and 1,
580 which were used in this study. Radiation and ground heat flux measurements yielded errors of up to
581 10% (compare <https://www.campbellsci.com/cnr4>). Although the energy balance closure
582 was rather poor, taking climatic conditions into account, it was still in the range of other carefully
583 designed experiments (e.g. Cui and Chui (2019)). The energy budget closure could most probably
584 be improved by additional radiation measurements and by relocating the tower further away from
585 the steep ridge to reduce perturbations of the mean flow for westerly winds.

586 *b. Importance of the Soil*

587 As rather sparse vegetation prevails within the eddy covariance footprint (20 - 30% lichen and 60
588 - 70% small shrubs with low LAI) we conclude that it is not the dominant factor which controls the

589 SEB. Thus, the soil properties take a more central role in partitioning the turbulent heat fluxes. The
590 coupling of soil moisture and evapotranspiration is emphasized by similar variation seen in Figure
591 6. The delay of evapotranspiration relative to soil moisture can likely be explained by variations
592 in air temperature which also show an increased importance once the changes in soil moisture are
593 rather small as observed in the second half of the period depicted in Figure 6.

594 Lemieux et al. (2020) investigated the groundwater dynamics of the Tasiapik valley and used
595 empirical formulas to estimate that 40% of the total precipitation returned to the atmosphere as
596 evapotranspiration. In our study, 22% to 35% of the precipitated water evaporated in the three
597 summers of interest. There are likely two probable reasons for this difference. One was the use of
598 empirical formulas in the Lemieux et al. (2020) study that were not specifically developed for the
599 climatic conditions in Umiujaq. The second reason is the fact that the watershed contained areas
600 with clay-rich soil, which can retain more water than sandy soil (Clapp and Hornberger 1978).
601 Thus, the evapotranspiration rates for some parts of the Tasiapik valley are likely to be higher than
602 those found around the flux tower used in our study.

603 ISBA performed slightly better in simulating soil conditions than SVS, revealing weaknesses of
604 using a simplified soil scheme. However, both models exhibited problems when simulating the
605 vertical profile of the soil water content and showed no or only a slight gradient of the volumetric
606 water content in the default configurations (Figure 9). Thus, the models accounted for only a
607 fraction of the actual changes in the soil properties that occurred with depth.

608 *c. Turbulent Fluxes and Summer Energy Balance*

609 Overall, the sensible heat flux is the dominant consumer of the energy from radiation due to
610 the striking dependence of the latent heat flux on the soil water content. On average, the sensible
611 heat flux is 61.3 W m^{-2} during summer, showing a decline from 81.5 W m^{-2} at the beginning of

612 summer to 48.3 W m^{-2} at the end. As previously discussed, the latent heat flux is more constant
613 throughout summer with an average value of 39 W m^{-2} . The net input of energy from the radiation
614 on average is 187.5 W m^{-2} .

615 Thus, in Umiujaq, the sensible heat flux makes up 33% of the net radiation on average while
616 22.4% can be attributed to latent heat flux. At the comparison sites, the proportion of sensible
617 heat flux to radiation is 22% to 40% and the proportion of latent heat flux varies between 29%
618 and 55%. Thus, sensible heat fluxes in Arctic environments seems somewhat less variable, while
619 the differences in the latent heat flux are higher and depend more on local peculiarities such as
620 water availability. This makes latent heat fluxes more difficult to simulate. The poor simulation
621 of the soil at the Umiujaq site by the default configurations in both models, which leads to wetter
622 conditions, impacted simulations of the turbulent fluxes as the latent heat flux was larger due to
623 more water availability. After adjusting the soil properties, the ISBA simulations for these fluxes
624 improved, indicating the influence of soil moisture on the turbulent fluxes in both models. While
625 H was accurately simulated along the entire range of observed values, this was not the case for LE .
626 The simulated values for LE were overestimated, closer to the higher-end values of the observed
627 LE . This difference may be attributed to the fact that the parametrization of LE is less adapted to
628 the dry soils at that site.

629 Both SVS and ISBA apply the same parametrization for the above-canopy turbulent fluxes.
630 Therefore the reason for the more accurate H simulations in comparison to LE most probably
631 lies in the soil simulation. Both the calculations for the thermal and the hydraulic regime are
632 considerably simpler in SVS. As H and LE depend on the temperature and moisture content at the
633 surface, respectively, an error in those variables directly affects the calculation of H and LE .

634 **5. Conclusions**

635 We analysed the SEB of a Subarctic experimental site in eastern Canada using the eddy covariance
636 technique. We then compared the SEB to other Arctic and Subarctic sites and assessed two current
637 LSMs, ISBA and SVS, in their ability to simulate SEB at the experimental site.

638 Our results showed that the sensible heat flux was the most important way of dissipating the
639 energy from net radiation. The sensible heat flux accounted for about 33% of net radiation while
640 the latent heat flux made up only 22.4%. The experimental site near Umiujaq consisted of sandy
641 soil that resulted in very low soil water content at the surface, despite abundant precipitation for
642 this latitude. The dry soil conditions most likely constrained the latent heat flux by limiting the
643 water available for evaporation. This condition favored the dissipation of energy through sensible
644 heat flux rather than through latent heat flux. Thus, we conclude that soil has a major influence on
645 the partitioning of the turbulent heat fluxes in contrast to other studies that favor vegetation as the
646 dominant factor for this partitioning.

647 Other documented sites in the Arctic showed net radiation partitioning that was different. At
648 these sites, the latent heat flux was larger than the sensible heat flux. We attribute this difference
649 to soil properties that retain more water, allowing it to drain slower.

650 We have also shown that both LSMs either need to be tuned or are not sophisticated enough to
651 account for this particular situation concerning soil content. For this reason, we adjusted the models
652 to be able to account for this factor. However, once the soil water content was better simulated
653 through the modification of the models, the sensible heat flux in particular was very accurately
654 simulated. On the other hand, latent heat flux simulations were not as accurate. Nonetheless,
655 the results confirm that both LSMs are generally suitable for operational hydrometeorological
656 applications in the humid Subarctic tundra ecotone.

657 Due to the isostatic uplift that followed deglaciation and led to the emergence of beaches,
658 we stress that the sandy soil that covers a significant portion of Arctic Canada causes energy
659 partitioning that is very different from what is presented in other available studies. A priori soil
660 conditions found at the Tasiapik valley site seem uncommon. However, consulting this map
661 ([https://ftp.maps.canada.ca/pub/nrcan_rncan/publications/STPublications_](https://ftp.maps.canada.ca/pub/nrcan_rncan/publications/STPublications_PublicationsST/295/295462/gscgcm_195_b_2014_mn01p1.pdf)
662 [PublicationsST/295/295462/gscgcm_195_b_2014_mn01p1.pdf](https://ftp.maps.canada.ca/pub/nrcan_rncan/publications/STPublications_PublicationsST/295/295462/gscgcm_195_b_2014_mn01p1.pdf)), it can be seen that many
663 regions in the Arctic have similar soil conditions. Indeed, large parts of the coasts of Hudson
664 Bay and Ungava Bay, as well as other parts of the coast of numerous other regions in the eastern
665 and central Canadian Arctic are covered by coarse sediments such as sand and gravel. Moreover,
666 important parts of the Canadian Arctic comprise rocky outcrops that likely show even higher
667 Bowen ratios. The soil water content and latent heat flux might therefore be overestimated by
668 surface models across the Canadian Arctic. Furthermore, the disappearance of permafrost due to
669 climate change may lead to drier soil and subsequently, altered latent heat fluxes.

670 *Data availability statement.* Data available on request from the authors.

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676 APPENDIX

677 **Supplementary Figures**

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853 **LIST OF TABLES**

854 **Table 1.** Description of the study sites. MAT and MAP are the mean annual temperature
855 and the mean annual precipitation. C: Continuous, DC: Discontinuous, S:
856 Sporadic. 42

857 **Table 2.** Hydrometeorological conditions for the three summers of interest. Soil temper-
858 ature measurement depth is 5 cm. 43

Site	CA-UM	CA-NS	US-ICH	US-EML	US-AN	RU-VRK	RU-COK
Location	Quebec, Canada	Manitoba, Canada	Alaska, US	Alaska, US	Alaska, US	Vorkuta, Russia	Chokurdakh, Russia
Coordinates	56.33°N 76.28°E	55.92°N 98.96°E	68.61°N 149.30°E	63.88°N 149.25°E	68.93°N 150.27°E	67.05°N 62.94°W	70.83°N 147.49°W
Elevation (m ASL)	140	244	940	700	600	100	48
MAT (°C)	-4.0	-3.1	-7.4	-1.0	-11.0	-6.1	-14.3
MAP (mm)	800-1000	495	318	378	122	538	212
Vegetation	shrub tundra	heath tundra	tussok tundra	tussok tundra	shrub tundra	shrub tundra	shrub tundra
Soil Type	sand	clay	porous peat	till/ loess	n/a	loamy and peaty	fluvial silt
Permafrost	DC/S	DC/S	C	C	C	DC/S	C
Reference	This study	Goulden (2016)	Bret-Harte et al. (2016)	Belshe et al. (2012)	Rocha et al. (2016)	Biasi and Friborg (2016)	van der Molen et al. (2007)

TABLE 1: Description of the study sites. MAT and MAP are the mean annual temperature and the mean annual precipitation. C: Continuous, DC: Discontinuous, S: Sporadic.

	Period of interest	T_{mean} (°C)	T_{mean}^{Soil} (°C)	Precipitation (mm)	Maximum snow height (cm) in previous winter
2017	1 July - 5 September	9.6	11.5	237	65
2018		8.1	9.3	290	140
2019		10.1	11.7	369	100
Mean		9.3	10.8	298	101

TABLE 2: Hydrometeorological conditions for the three summers of interest. Soil temperature measurement depth is 5 cm.

859 **LIST OF FIGURES**

860 **Fig. 1.** a) Location of the study site and the experimental setup in the Tasiapik valley close to the
861 village of Umiujaq on the eastern shore of Hudson Bay, Canada and b) photo of the upper
862 portion of the valley where the flux tower is installed. The angle of view in panel b) is shown
863 in panel a). Note the presence of a cuesta west of the tower. 46

864 **Fig. 2.** Mean, high and low 2-m air temperatures as well as the mean amount of snow and rain per
865 month. The shaded area indicates the period of interest for this study. 47

866 **Fig. 3.** Soil profile under lichen cover in the vicinity of the flux tower. 48

867 **Fig. 4.** FLUXNET site locations and spatial pattern of circumpolar yearly precipitation. Source:
868 Harris et al. (2014) 49

869 **Fig. 5.** Partitioning of net radiation into H , LE , and G for the three summers. The residual Res
870 after the partitioning is also given. Only data with EBC between 0.2 and 1.5 were used. 50

871 **Fig. 6.** Daily sum of evapotranspiration (ET) and precipitation and daily mean of soil water content
872 (SWC) at 6-cm depth, for summer 2017 (June to August). 51

873 **Fig. 7.** Mean net radiation (black lines), sensible and latent heat fluxes (red and blue lines; uncor-
874 rected for energy imbalance) of all available summers and selected study sites. The shaded
875 areas illustrates the mean standard deviation. Only data during daytime ($R_n > 0$) were used.
876 Sites are presented in ascending order of latitude. The bottom right plot shows the mean of
877 Umiujaq compared to the mean for all sites. For RU-COK only one year of data was available. . . 52

878 **Fig. 8.** Ternary diagram of soil textures according to the USDA classification. Circles indicate the
879 estimates for each site. For site US-An, no data about the soil was available. The color of
880 the circles informs on the BR: purple: $BR < 1$; orange: $BR \approx 1$ and green: $BR > 1$. Source: see
881 references in Table 1. 53

882 **Fig. 9.** Observed and modeled soil water content at 2 different depths for 2017 for models with
883 default parameters. 54

884 **Fig. 10.** Observed and modeled soil water content at 2 different depths for 2017 for models with
885 adjusted parameters. 55

886 **Fig. 11.** Left column: Scatter plots comparing observed and modeled sensible heat fluxes at the
887 Umiujaq site for all summers with the colors indicating the hour of day for (a) ISBA and (c)
888 SVS. Right column: q-q plots, for which the quantiles of the distributions are plotted against
889 each other, for (b) ISBA and (d) SVS. 56

890 **Fig. 12.** Same as Figure 11 but for latent heat flux 57

891 **Fig. 13.** Taylor diagrams indicating how closely modeled sensible heat flux (left) and latent heat flux
892 (right) match observations. 58

893 **Fig. A1.** Weekly running mean of the air temperature for summer 2017, 2018 and 2019. 59

894 **Fig. A2.** Weekly running mean of summer sensible and latent heat fluxes for all three study years.
895 Summer 2017 starts later as data were available only from mid June. 60

896 **Fig. A3.** Drone image of the study site with the flux footprint climatology of summer 2017, indicating
897 80% of the source area. 61

898 **Fig. A4.** Dependence of the energy budget closure on the wind direction for all three summers. 62

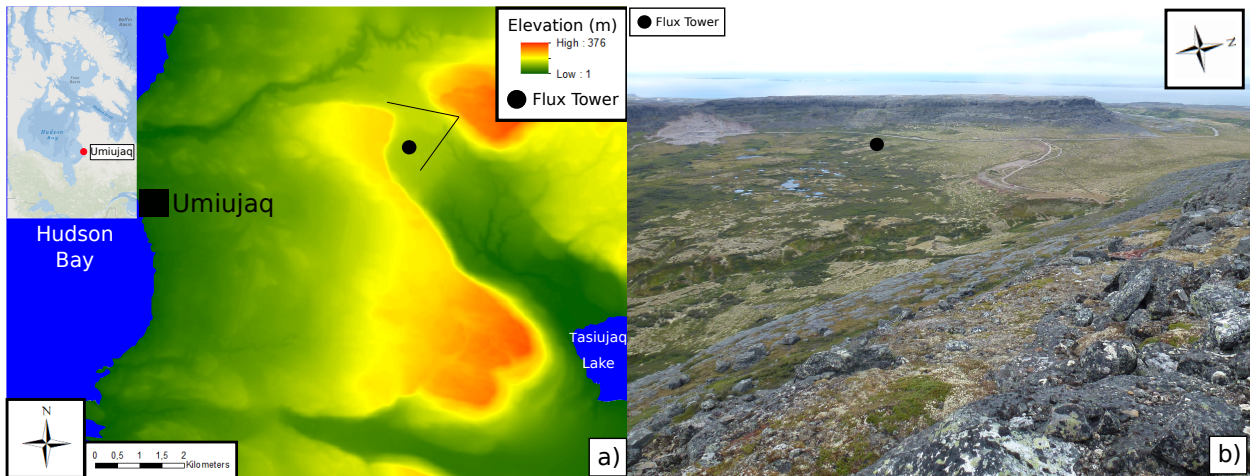


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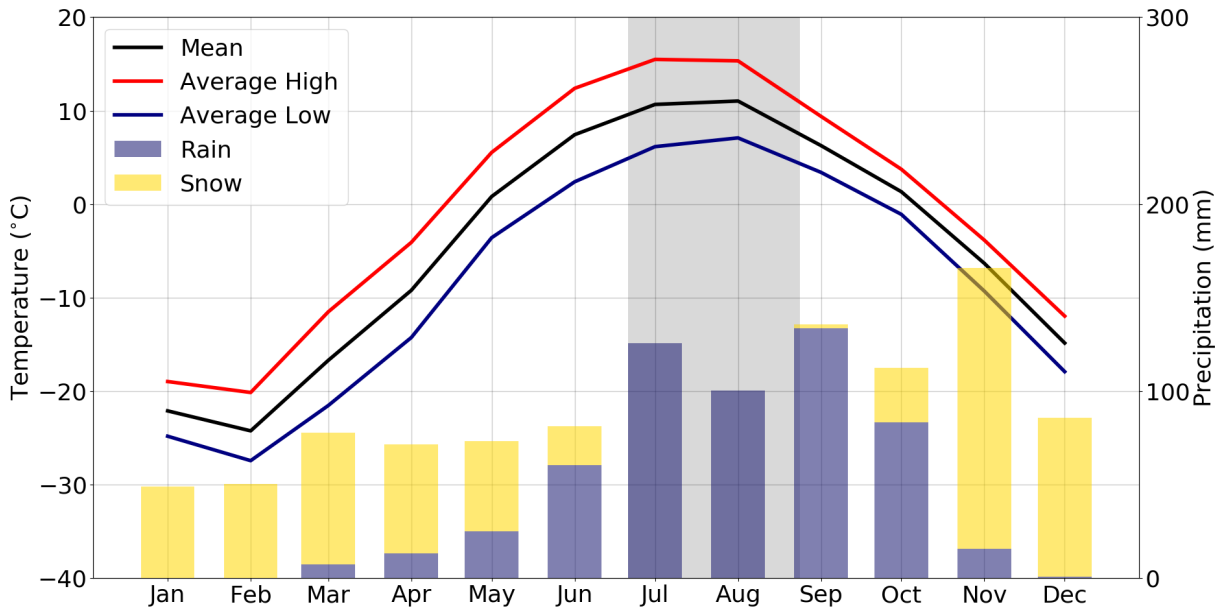


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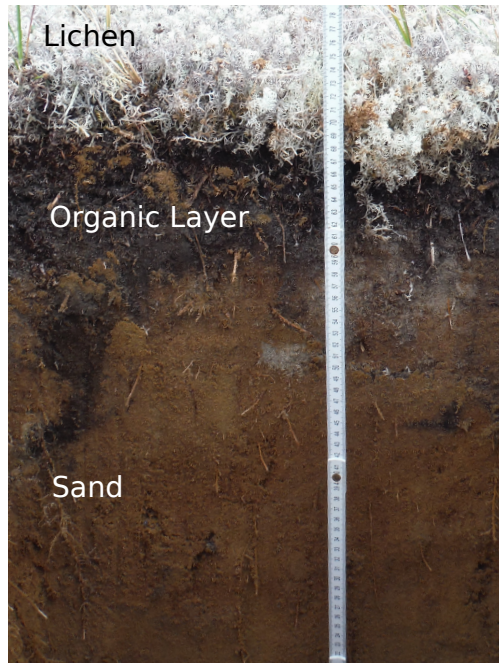


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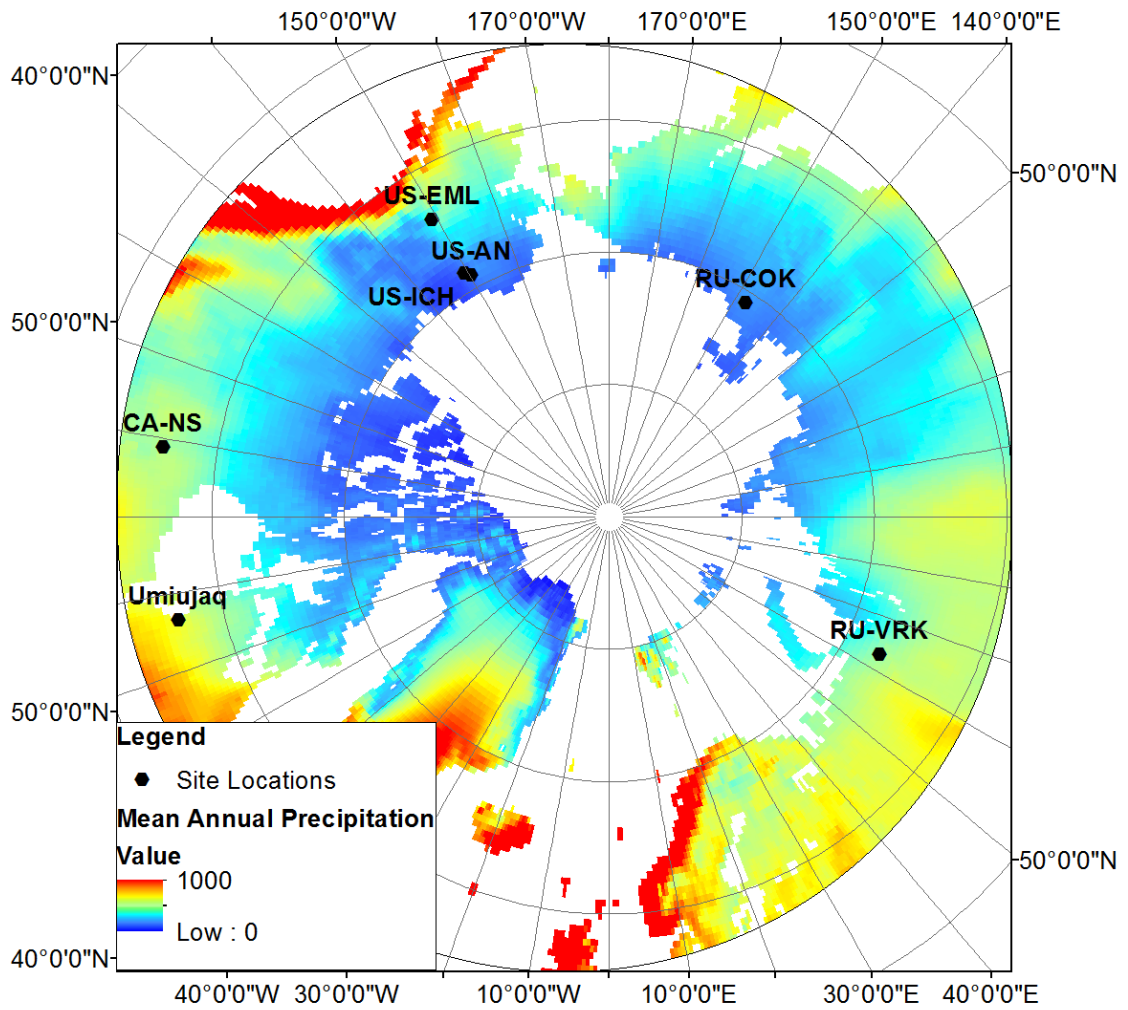


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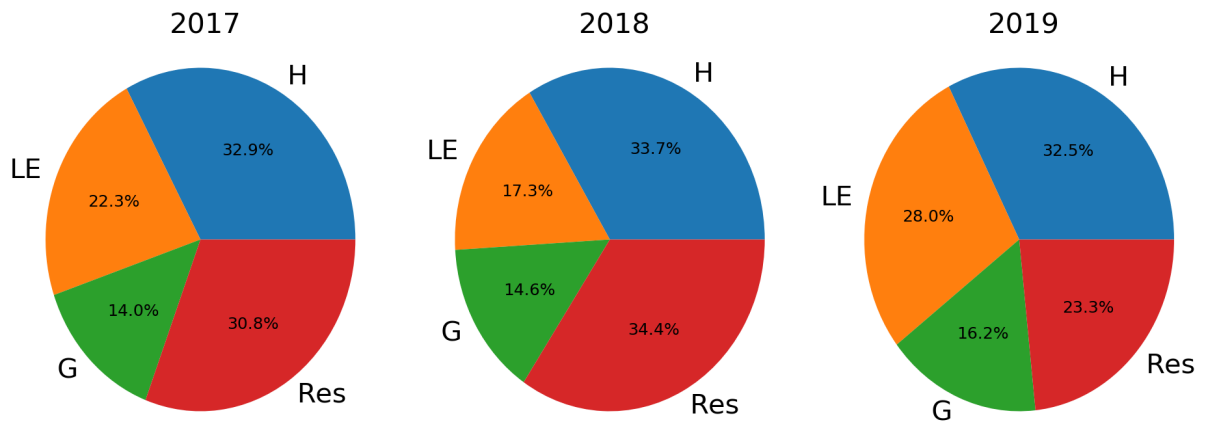


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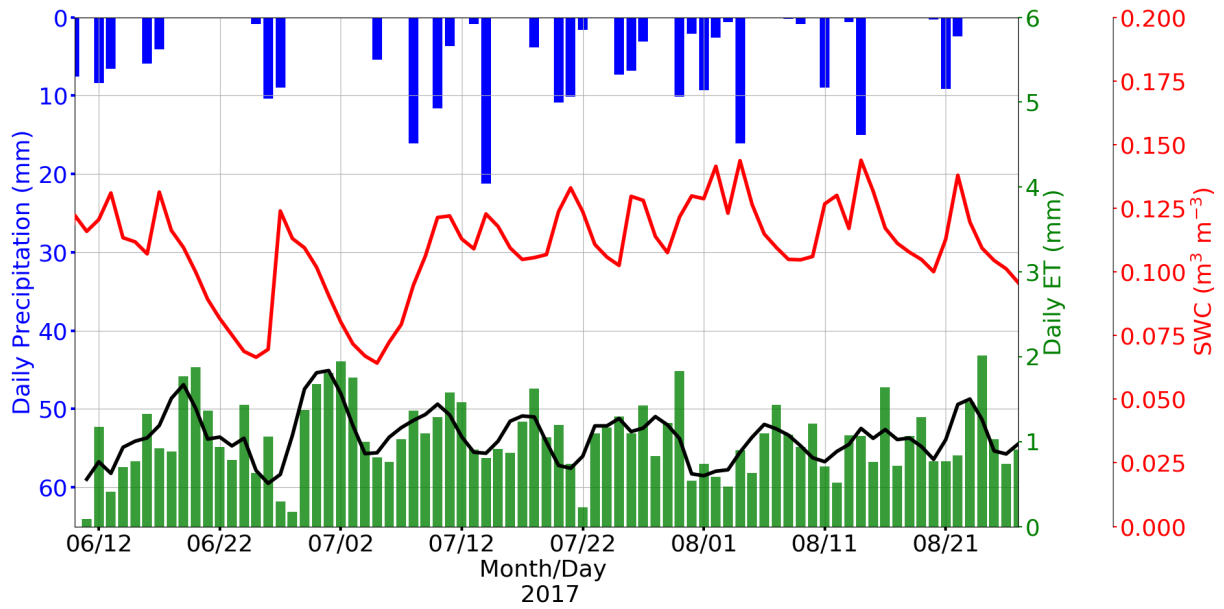


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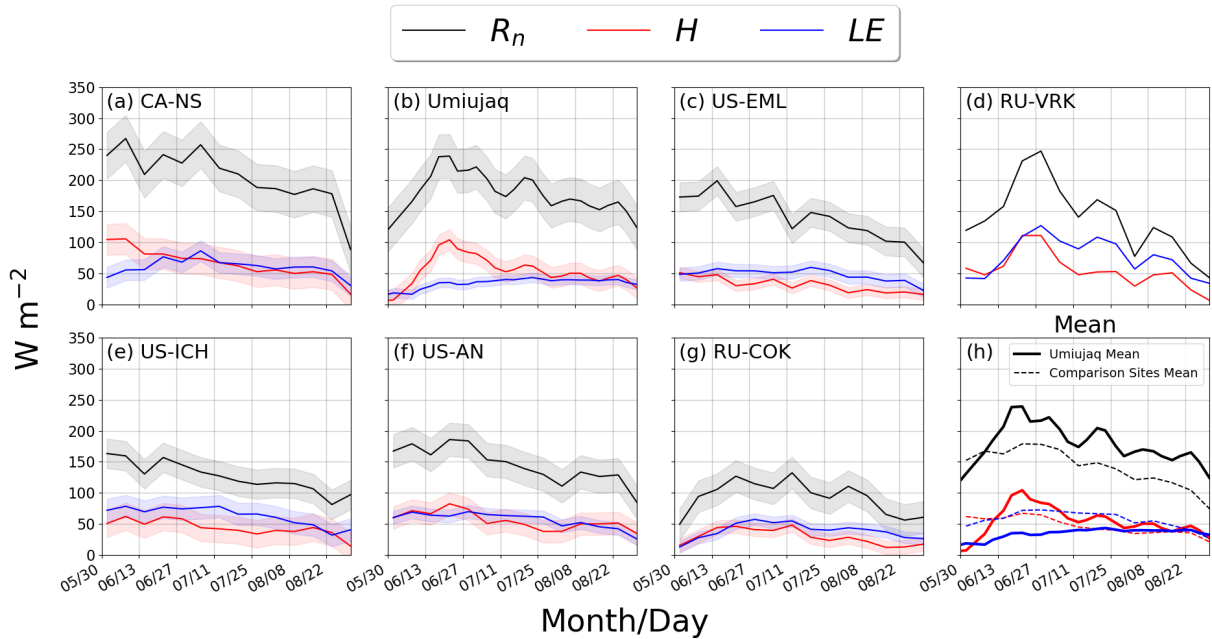


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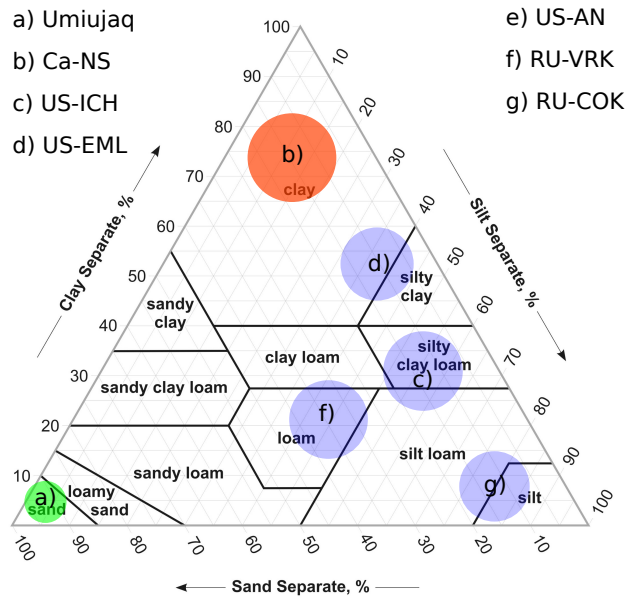


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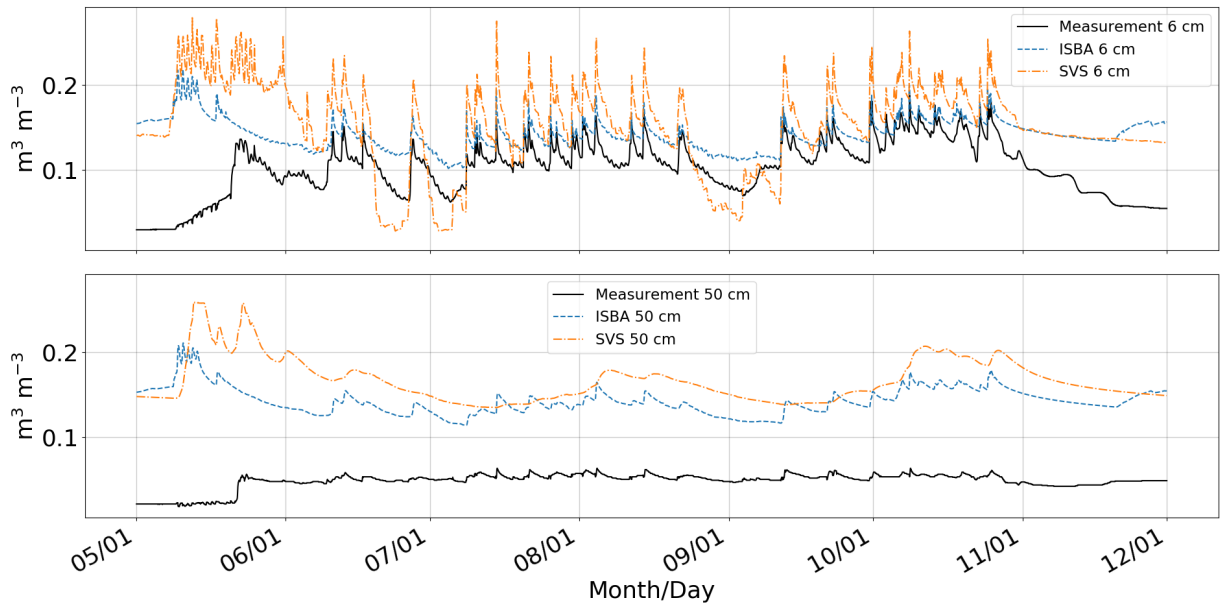


FIG. 9: Observed and modeled soil water content at 2 different depths for 2017 for models with default parameters.

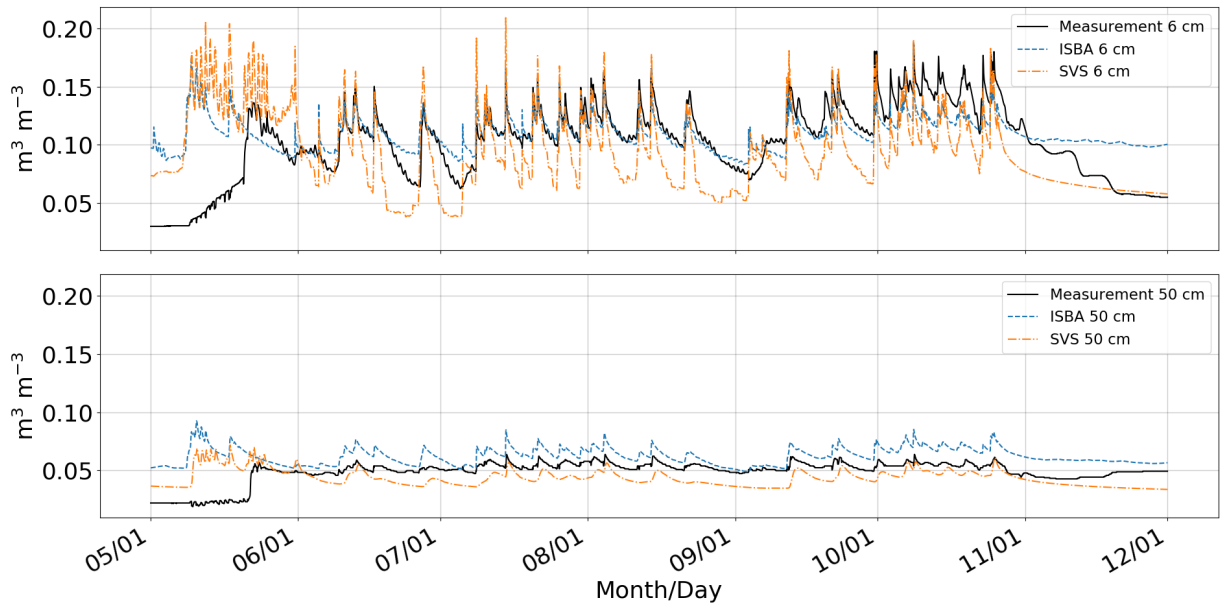


FIG. 10: Observed and modeled soil water content at 2 different depths for 2017 for models with adjusted parameters.

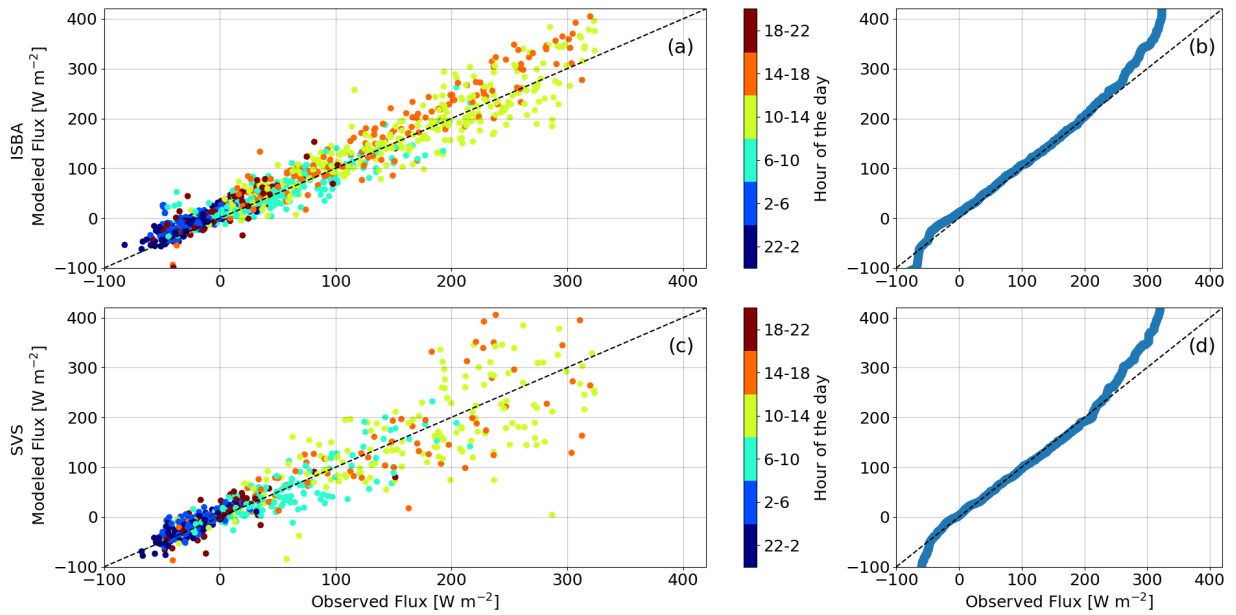


FIG. 11: Left column: Scatter plots comparing observed and modeled sensible heat fluxes at the Umiujaq site for all summers with the colors indicating the hour of day for (a) ISBA and (c) SVS. Right column: q-q plots, for which the quantiles of the distributions are plotted against each other, for (b) ISBA and (d) SVS.

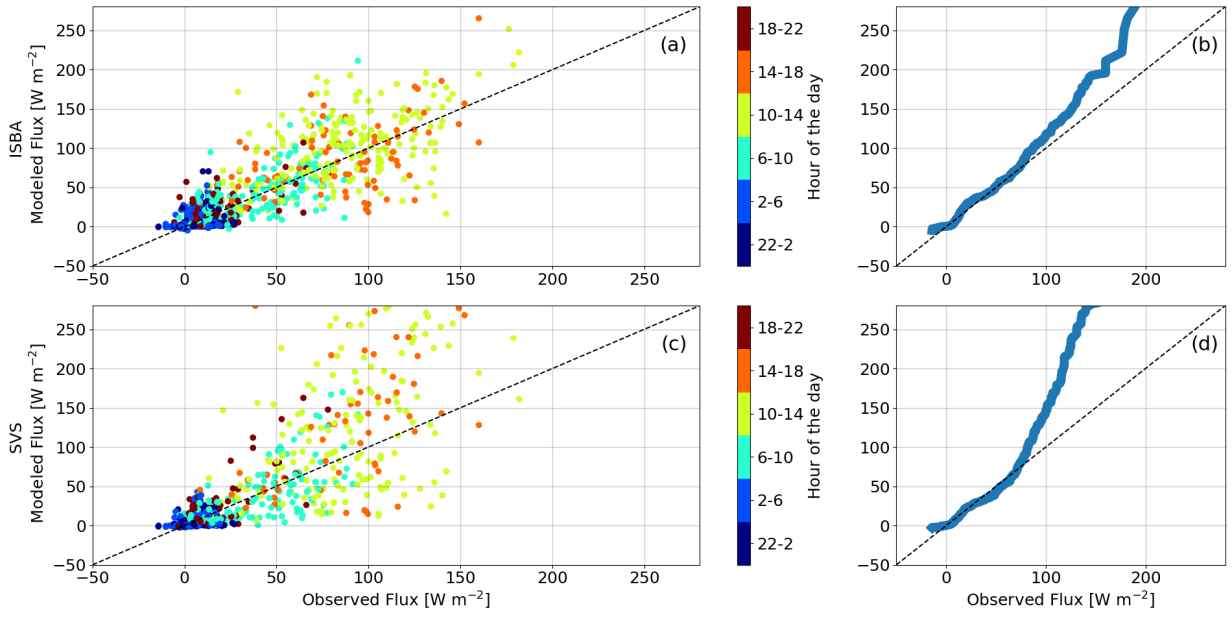


FIG. 12: Same as Figure 11 but for latent heat flux

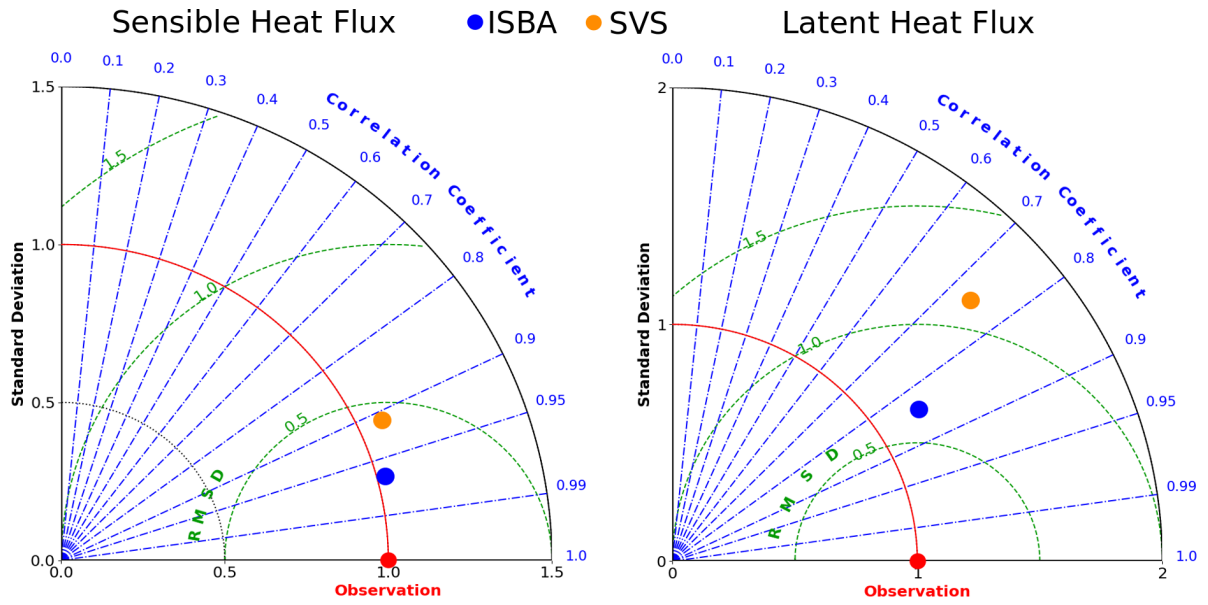


FIG. 13: Taylor diagrams indicating how closely modeled sensible heat flux (left) and latent heat flux (right) match observations.

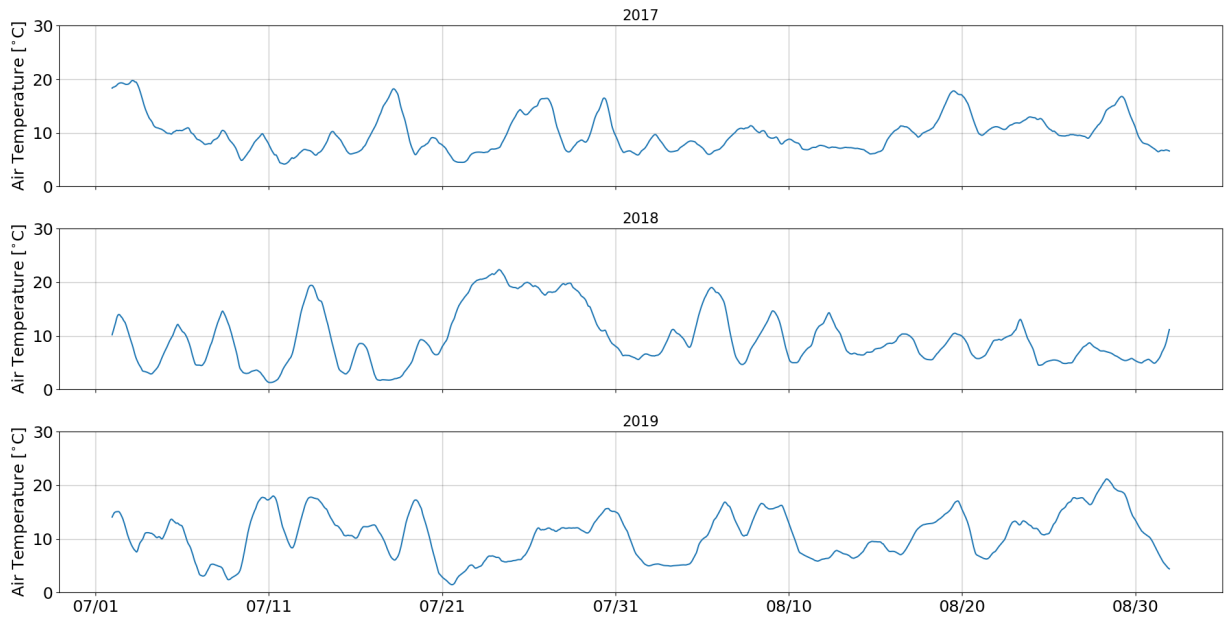


Fig. A1: Weekly running mean of the air temperature for summer 2017, 2018 and 2019.

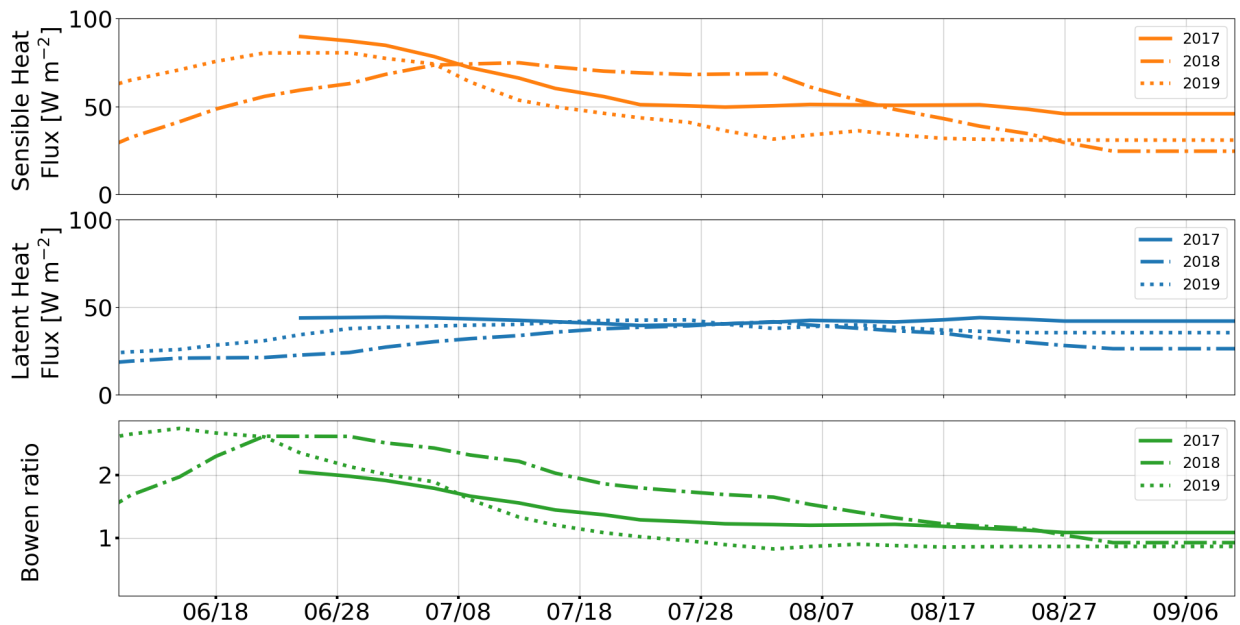


Fig. A2: Weekly running mean of summer sensible and latent heat fluxes for all three study years. Summer 2017 starts later as data were available only from mid June.

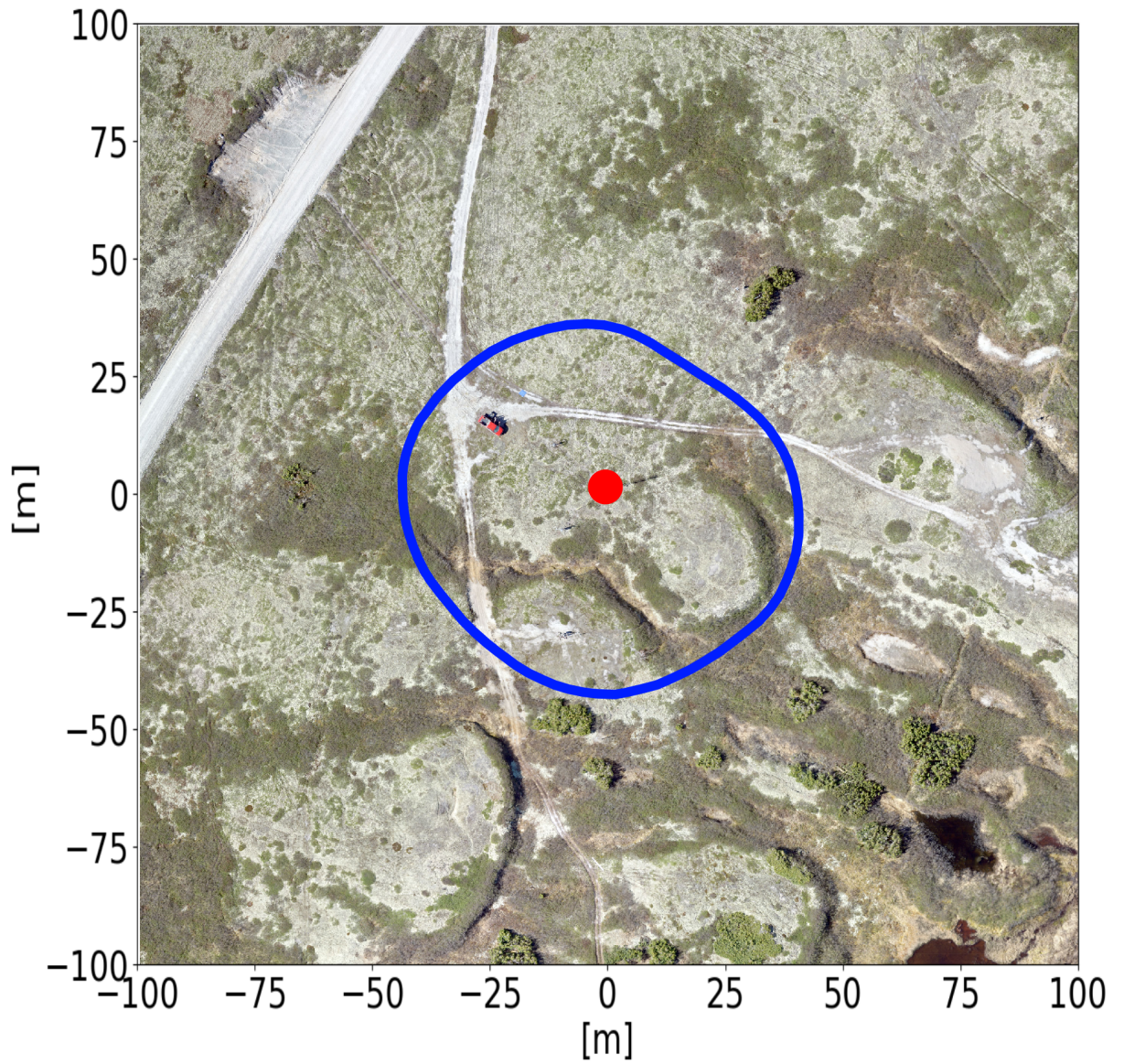


Fig. A3: Drone image of the study site with the flux footprint climatology of summer 2017, indicating 80% of the source area.



Fig. A4: Dependence of the energy budget closure on the wind direction for all three summers.