

1 Changing Characteristics of Runoff and Freshwater  
2 Export From Watersheds Draining Northern  
3 Alaska

4 Michael A. Rawlins<sup>1</sup>, Lei Cai<sup>2</sup>, Svetlana L. Stuefer<sup>3</sup>, and Dmitry  
5 Nicolsky<sup>4</sup>

6 <sup>1</sup>Department of Geosciences, University of Massachusetts, Amherst,  
7 MA 01003, USA

8 <sup>2</sup>International Arctic Research Center, University of Alaska Fairbanks,  
9 Fairbanks, AK 99775

10 <sup>3</sup>Civil and Environmental Engineering, College of Engineering and  
11 Mines, University of Alaska Fairbanks, Fairbanks, AK 99775 USA

12 <sup>4</sup>Geophysical Institute, University of Alaska Fairbanks, Fairbanks, AK  
13 99775, USA

14 Corresponding author: Michael A. Rawlins <rawlins@geo.umass.edu>

15 **Abstract**

16 The quantity and quality of river discharge in arctic regions is influenced  
17 by many processes including climate, watershed attributes and, increasingly,  
18 hydrological cycle intensification and permafrost thaw. We used a hydrological  
19 model to quantify baseline conditions and investigate the changing character  
20 of hydrological elements for Arctic watersheds between Point Barrow and just  
21 west of Mackenzie River over the period 1981–2010. A synthesis of measure-  
22 ments and model simulations shows that the region exports  $31.9 \text{ km}^3 \text{ yr}^{-1}$  of  
23 freshwater via river discharge, with 57.7% ( $18.4 \text{ km}^3 \text{ yr}^{-1}$ ) coming collectively  
24 from the Colville, Kuparuk, and Sagavanirktok rivers. The simulations point  
25 to significant ( $p < 0.05$ ) increases (134–212% of average) in cold season dis-  
26 charge (CSD) for several large North Slope rivers including the Colville and

27 Kuparuk, and for the region as a whole. A significant increase in the proportion  
28 of subsurface runoff to total runoff is noted for the region and for 24 of the 42  
29 study basins, with the change most prevalent across the northern foothills of  
30 the Brooks Range. Relatively large increases in simulated active-layer thick-  
31 ness (ALT) suggest a physical connection between warming climate, permafrost  
32 degradation, and increasing subsurface flow to streams and rivers. A decline in  
33 terrestrial water storage (TWS) is attributed to losses in soil ice that outweigh  
34 gains in soil liquid water storage. Over the 30 yr period the timing of peak  
35 spring (freshet) discharge shifts earlier by 4.5 days, though the time trend is  
36 only marginally ( $p = 0.1$ ) significant. These changing characteristics of Arctic  
37 rivers have important implications for water, carbon, and nutrient cycling in  
38 coastal environments.

39 **KEYWORDS:** Arctic; runoff; river discharge; permafrost; subsurface flow

## 40 **1 Introduction**

41 The arctic water cycle is central to a range of climatic processes and to the  
42 transfer of carbon, energy, and other materials from the land mass to coastal waters  
43 of the Arctic Ocean. Freshwater export to the Arctic Ocean is high relative to the  
44 ocean's area (Shiklomanov et al., 2000), and dominated by river discharge (Serreze  
45 et al., 2006), which serves as a conveyance for carbon and heat across the land-  
46 ocean boundary. Syntheses of data and models have advanced understanding of key  
47 linkages and feedbacks in the Arctic system (Francis et al., 2009), mean freshwater  
48 budgets across the land, atmosphere and ocean domains (Serreze et al., 2006), and  
49 time trends in observations and model estimates over the latter decades of the 20<sup>th</sup>  
50 century (Rawlins et al., 2010).

51 A warming climate is expected to lead to intensification of the hydrological cy-  
52 cle, including increases in net precipitation (P) at high latitudes, and evidence of  
53 broad-scale intensification is emerging (Peterson et al., 2002, 2006; Rawlins et al.,  
54 2010; Zhang et al., 2013; Bring et al., 2016). A more vigorous water cycle is related  
55 in part to both the amount of moisture air can hold and changes in atmospheric  
56 dynamics. Shorter ice duration on lakes and longer seasons for evaporation are also  
57 manifestations of warming on the Arctic hydrological cycle. Much of the increase in  
58 net P is expected to occur during winter (Kattsov et al., 2007), potentially through  
59 intensified local surface evaporation driven by retreating winter sea ice, and enhanced  
60 moisture inflow from lower latitudes (Zhang et al., 2013; Bintanja and Selten, 2014).  
61 An increase in river discharge from Eurasia to the Arctic Ocean was noted in sim-  
62 ulations with the HadCM3 general circulation model (Wu et al., 2005), illustrating

63 the potential for increased winter net P to influence freshwater export. Positive  
64 trends in column-integrated precipitable water over the region north of 70°N, linked  
65 to positive anomalies in air and sea surface temperature and negative anomalies in  
66 end-of-summer sea ice extent (Serreze et al., 2012), support the future model pro-  
67 jections. Rivers form a primary conduit for transferring terrestrial materials to the  
68 coastal ocean, and these materials exert a strong influence on marine ecosystems and  
69 carbon processing.

70 Permafrost warming and degradation has been observed over parts of Alaska,  
71 Russia, and Canada (Brown and Romanovsky, 2008; Romanovsky et al., 2010; Smith  
72 et al., 2010). In one study permafrost area is projected to decrease by more than  
73 40%, assuming climate stabilization at 2°C above pre-industrial (Chadburn et al.,  
74 2017). Warming and permafrost degradation is expected to cause a shift in arctic  
75 environments from a surface water-dominated system to a groundwater-dominated  
76 system (Frey and McClelland, 2009; Bring et al., 2016). There is increasing evidence  
77 of impacts of permafrost degradation on biogeochemical cycles on land and in aquatic  
78 systems. Recent reported increases in baseflow in arctic rivers are suggestive of  
79 increased hydrological connectivity due to permafrost thaw (Walvoord and Striegl,  
80 2007; Bense et al., 2009; Walvoord and Kurylyk, 2016; St. Jacques and Sauchyn,  
81 2009). Groundwater processes have a dominant role in controlling carbon export from  
82 the land to streams in permafrost terrain (Frey and McClelland, 2009; Neilson et al.,  
83 2018). In areas where much of the landscape is defined by the absence of permafrost,  
84 runoff generation processes can be much different from areas where permafrost is  
85 nearly continuous. Dissolved organic matter (DOM) transported by Arctic rivers  
86 contain geochemical signatures of the watersheds they drain, reflecting their unique  
87 characteristics (Kaiser et al., 2017). Changes in landscape characteristics and water  
88 flow paths as a result of climatic warming and associated active layer thickening  
89 have the potential to alter aquatic and riverine biogeochemical fluxes (Frey and  
90 McClelland, 2009; Wrona et al., 2016; Wickland et al., 2018). Increased flow through  
91 mineral soils has been linked to decreases in DOC export from the Yukon River  
92 over recent decades (Striegl et al., 2005). In contrast, areas with deep peat deposits  
93 that experience thaw may see increasing DOC mobilization and export as permafrost  
94 degrades (Frey and Smith, 2005).

95 This study presents baseline freshwater flux estimates and examines elements  
96 of the hydrological cycle across the North Slope over the period 1981–2010. We  
97 use measured data to assess model performance and combine with the simulated  
98 estimates to quantify freshwater export from the region. We then use the data  
99 and model simulations to investigate time changes in runoff and river discharge, the  
100 proportion of groundwater runoff, terrestrial water storage, and the timing of peak

101 daily discharge. Salient results in the context of arctic change and directions for  
102 future research are discussed.

## 103 **2 Study Area, Data and Modeling**

104 The study focuses on the North Slope of Alaska and NW Canada, partitioned by  
105 the region’s river basins that drain to the Beaufort Sea (Figure 1). Hereafter we refer  
106 to this region as the “North Slope”. The grid is based on the Northern Hemisphere  
107 EASE-Grid (Brodzik and Knowles, 2002), with a horizontal resolution of 25 km for  
108 each grid cell. The model domain contains 312 grid cells (total area = 196,060 km<sup>2</sup>)  
109 that define the North Slope drainage of northern Alaska and NW Canada. It is  
110 defined by the drainage basins of rivers (42 total, Table S1) with an outlet along the  
111 coast from just west of the Mackenzie River to Utqiakvik (formerly Barrow) to the  
112 west. Hydrologic modeling was performed for the North Slope domain encompassing  
113 the 42 watersheds. Many North Slope rivers are oriented roughly north-south, and  
114 the region is underlain by continuous permafrost, approximately 250–300 m thick in  
115 the Brooks Range and, locally, up to nearly 400 m thick near the coast (Jorgenson  
116 et al., 2008).

### 117 **2.1 Observational data**

118 Observational data used in this study include time series of daily river dis-  
119 charge, end-of-winter snow water equivalent (SWE), and seasonal maximum active-  
120 layer thickness (ALT). Historical river discharge data was retrieved from the USGS  
121 for the Kuparuk River (<http://waterdata.usgs.gov/nwis/uv?15896000>) and Colville  
122 River ([https://waterdata.usgs.gov/ak/nwis/uv/?site\\_no=15875000](https://waterdata.usgs.gov/ak/nwis/uv/?site_no=15875000)). Model simu-  
123 lated SWE is evaluated against average end-of-winter SWE from measurements  
124 across the Kuparuk River watershed. The measurements from 2000 to 2011 were  
125 taken at multiple locations distributed from the Brooks Range to the Beaufort Sea  
126 coast to better capture macro-scale SWE variability (Stuefer et al., 2013).

127 Simulated ALT from the PWBm (section 2.3) is compared with estimates from a  
128 high-resolution 1-D heat conduction model (developed by the University of Alaska’s  
129 Geophysical Institute Permafrost Laboratory, hereafter referred to as GIPL) that  
130 incorporated data on ecosystem type and was validated against measured CALM  
131 network ALTs (Nicolosky et al., 2017).

## 132 2.2 Reanalysis data

133 Gridded fields of daily surface (2 m) air temperature, precipitation (P), and wind  
134 speed are used as model forcings. Obtaining accurate temporally varying P estimates  
135 at daily resolution is particular challenging in arctic environments. Gauge undercatch  
136 of solid P is common, the gauge network is sparse and the number of stations at higher  
137 elevation is insufficient (Yang et al., 1998, 2005; Kane and Stuefer, 2015). In this  
138 study model meteorological forcings are drawn from the Modern-Era Retrospective  
139 Analysis for Research and Applications (MERRA; Rienecker et al. (2011)). In a  
140 recent intercomparison of P estimates over the Arctic Ocean and its peripheral seas,  
141 three reanalyses— ERA-Interim (Dee et al. (2011)), MERRA, and NCEP R2 (Kistler  
142 et al. (2001))— produce realistic magnitudes and temporal agreement with observed  
143 P events, while two products (MERRA, version 2 (MERRA-2), and CFSR) show  
144 large, implausible magnitudes in P events (Boisvert et al., 2018). Given a modest  
145 low bias in monthly P across the North Slope in MERRA, we derived a new bias  
146 corrected daily P time series by scaling the MERRA values by a factor defined using  
147 monthly long-term mean P (1981–2010) from MERRA, ERA-Interim, and a data set  
148 that blends simulations from ERA-Interim and the Polar WRF (Cai et al., 2018).  
149 Those three data sets exhibit a similar spatial pattern in annual P across the region.  
150 Annual P generally ranges from as low as 200 mm yr<sup>-1</sup> near the coast to over 400  
151 mm yr<sup>-1</sup> over the foothills of the Brooks Range. At each grid cell, the offset ratio  
152 was defined as average P from the 3 data sets divided by the MERRA P amount.  
153 The derived daily P (hereafter MERRA\*) was then calculated as the daily MERRA  
154 P amount multiplied by the offset ratio.

## 155 2.3 Hydrological modeling

156 The regional hydrology is characterized by water fluxes and storages expressed  
157 in simulations using a spatially-distributed numerical model. Referenced previously  
158 as the Pan-Arctic Water Balance Model (PWBM), the numerical framework encom-  
159 passes all major elements of the water cycle, including snow storage, sublimation,  
160 transpiration, and surface evaporation (Rawlins et al., 2003, 2013). Model input and  
161 output fields are resolved at a daily time step. The simulations are commonly per-  
162 formed at an implicit daily time step, typically forced with meteorological data. The  
163 PWBM has been used to investigate causes behind the record Eurasian discharge in  
164 2007 (Rawlins et al., 2009); to corroborate remote sensing estimates of surface water  
165 dynamics (Schroeder et al., 2010); and to quantify present and future water cycle  
166 changes in the area of Nome, Alaska (Clilverd et al., 2011). In a comparison against  
167 observed river discharge, PWBM-simulated SWE fields compared favorably (Rawlins

168 et al., 2007). Soil temperature are simulated dynamically are through an embedded  
169 1-D nonlinear heat conduction sub-model with phase change (Rawlins et al., 2013;  
170 Nicolsky et al., 2017). PWBM includes a multi-layer snow model that accounts for  
171 wind compaction, change in density due to fresh snowfall, and depth hoar develop-  
172 ment with time. Runoff is the sum total of surface (overland) and subsurface flow  
173 each day. Subsurface runoff occurs when the amount of water in a soil layer exceeds  
174 field capacity.

175 The model is well suited for application across the North Slope region. Active-  
176 layer thickness (ALT) simulated using the PWBM was found to be more similar to in  
177 situ observations and airborne radar retrievals in continuous permafrost areas than in  
178 lower permafrost probability areas (Yi et al., 2018). The influence of snow cover and  
179 soil thermal dynamics on the seasonal and spatial variability in soil CO<sub>2</sub> respiration  
180 has been quantified by coupling PWBM to a dynamic soil carbon model (Yi et al.,  
181 2013, 2015). A key model attribute is its ability to dynamically simulate the direct  
182 influence the snowpack exerts on soil temperature (Yi et al., 2019), with deeper  
183 snowpacks promoting warmer soils and associated effects, such as enhancement of  
184 soil decomposition and respiration from deeper ( $\geq 0.5$  m) soil layers (Yi et al., 2015).  
185 Detailed descriptions of the PWBM can be found in Rawlins et al. (2003, 2013); Yi  
186 et al. (2015, 2019) and Appendices within.

187 In this study we applied an updated version of the model, and given its detailed  
188 representation of soil freeze-thaw processes, rename it the “Permafrost Water Bal-  
189 ance Model” (hereafter PWBM v3). Recent modifications involved the incorporation  
190 of new data and parametrizations for surface fractional open water ( $f_w$ ) cover, soil  
191 carbon content, and transient ponded surface evaporation and runoff. Updates to  
192 the spatial estimates of  $f_w$  were drawn from a product derived from brightness tem-  
193 perature ( $T_b$ ) retrievals from the Advanced Microwave Scanning Radiometer for EOS  
194 (AMSR-E) (Du et al., 2017) to parameterize the grid cell fraction of open water (an-  
195 nual average) across the model domain. Properties of near surface organic-rich soils  
196 strongly control hydrological and thermal dynamics in the seasonally thawed active  
197 layer. We used soil organic carbon (SOC) estimates from version 2.2 of the Northern  
198 Circumpolar Soil Carbon Database (NCSCD), a digital soil map database linked to  
199 extensive field-based SOC storage data (Hugelius et al., 2014). The database contains  
200 SOC stocks for the upper 0–1 m and for deeper soils from 1–2 and 2–3 m depth. In  
201 the updated PWBM v3 the sum total of SOC in the upper 3 m was used to derive the  
202 organic layer thickness as described in Rawlins et al. (2013). The resulting spatially  
203 varying parameterizations of soil carbon profiles (% of volume) with depth over the  
204 domain (Figure S1a) influence soil thermal properties and hydrological storages and  
205 fluxes. Broad agreement exists in the spatial pattern of the independent soil carbon

206 and soil texture datasets (Figure S1a,b). Sandy soils and soil carbon thicknesses un-  
207 der 20 cm occur over the Brooks Range, and relatively higher soil carbon thicknesses  
208 and loam soils are present across the tundra to the north. Based on analysis of initial  
209 model simulations we increased soil carbon amounts by 10% in areas (24 grid cells) of  
210 sandy soils and reassigned the texture to loam, making the parameterizations more  
211 consistent with soil textures inferred from high-resolution ALT mapping using the  
212 GIPL model that incorporated data on ecosystem type (Nicolson et al., 2017).

213 Model calibration was performed to adapt the model and optimize its perfor-  
214 mances in simulating the water cycle across the study domain, and involved the  
215 surface transient storage pool and river flow velocity. Transient surface storage con-  
216 sists of water connected to the surface flow that is delayed in its transport to stream  
217 networks. Parameters controlling evaporation and runoff fluxes from surface stor-  
218 age were modified to better account for delays in water reaching stream channels.  
219 Defining  $E_i$ ,  $R_i$ , and  $S_i$  to represent evaporation (or evapotranspiration)( $\text{mm day}^{-1}$ ),  
220 runoff ( $\text{mm day}^{-1}$ , and storage (mm) in soil layer  $i$ , respectively, then  $E_0$ ,  $R_0$ ,  $S_0$   
221 are evaporation, runoff, and storage from the model surface layer,  $R_0 = S_0 * f$  ( $\text{mm}$   
222  $\text{day}^{-1}$ ). In the updated model  $f = 0.40$ , reduced from the prior value of 0.75. Evapo-  
223 ration from surface storage is  $E_0 = S_0 * g$ , with  $g$  now reduced to 1/3 of the potential  
224 ET rate.

225 Model estimated runoff routed through a simulated topological network (STN)  
226 (Vörösmarty et al., 2000) is expressed as river discharge (volume flux) at the coastal  
227 outlets of 42 individual watersheds draining from Point Barrow to just west of the  
228 Mackenzie River delta. A simple linear routing model is used given the relatively  
229 short travel times through the North Slope basins. Water transferred to the down-  
230 stream grid or exported off the coast is

$$Q_{\text{out}} = \frac{v}{d}S \quad (1)$$

231 where  $Q_{\text{out}}$  ( $\text{m}^3 \text{ s}^{-1}$ ) is flow downstream,  $v$  is flow velocity ( $\text{m s}^{-1}$ ),  $d$  is the distance  
232 between grid cells (m), and  $S$  is volume of river water ( $\text{m}^3$ ). Miller et al. (1994)  
233 suggested a global average of  $v = 0.35 \text{ m s}^{-1}$ . Given the relatively flat topography  
234 over much of the domain we set effective velocity at  $v = 0.175$ . Hereafter  $R$  repre-  
235 sents runoff expressed in unit depth, and  $Q$  represents river discharge volume flow  
236 estimated through the routing model.

237 The PWBM is run in a 50 year spinup over year 1980 prior to the transient time  
238 series simulation to stabilize soil temperature and water storage pools. This spinup  
239 is followed by a 30 year transient simulation over the period 1981–2010, the focus of  
240 our analysis.

241 Assessment of several model simulated quantities is made using average error  
242 and correlation. Model evaluation metrics based on squared values like the root  
243 mean square error (RMSE) are known to be biased and highly sensitive to outliers  
244 (Willmott and Matsuura, 2005; Willmott et al., 2015). Statistical significance is  
245 calculated using the Mann-Kendall test statistic (Hamed and Rao, 1998; Yue et al.,  
246 2002), with a 95% confidence level ( $p < 0.05$ ) designated as statistically significant.  
247 Time changes are estimated with a General Linear Model (GLM). We apply the  
248 modified Mann-Kendall test (Hamed and Rao, 1998) for terrestrial water storage  
249 (TWS) and its component storages of snow (water equivalent), soil liquid water  
250 and ice amounts. A one or a two-sided test is applied depending on whether the  
251 direction of change is assumed. For example, we posit null hypotheses that the  
252 region is experiencing increasing cold season discharge as a result of ALT increase.

## 253 **3 Model Validation**

### 254 **3.1 Active layer thickness**

255 Simulated maximum seasonal ALT derived from daily soil temperatures in the  
256 updated PWBM v3 model simulation with meteorological forcing from MERRA re-  
257 analysis (bias corrected MERRA\* P) is evaluated alongside ALT predicted from the  
258 GIPL model. Area averaged ALT from PWBM and GIPL is 53.5 and 55.2 cm re-  
259 spectively, a difference of  $\sim 3\%$  (Figure S2, Table 1), and smallest difference among  
260 average ALT derived from soil temperatures in simulations using alternate meteo-  
261 rological forcings. Simulated ALT exhibits the expected north-south gradient which  
262 reflects the gradient in summer (and annual) air temperature (Figure S3). Agree-  
263 ment in ALT between PWBM (MERRA\*) and GIPL is strongest in coastal areas.  
264 The estimates differ most near the center of the domain where the PWBM produces  
265 relatively smaller ALT compared to GIPL. The differences increase toward the ex-  
266 tremes of each field, pointing to higher spatial variability in the PWBM simulations  
267 (Figure S2). ALT from simulations with the default MERRA P forcing are shallower  
268 and less in agreement with the GIPL data.

### 269 **3.2 Snow water equivalent**

270 In the Kuparuk River basin maximum end of season SWE typically occurs near  
271 the end of April. Simulated end of season SWE each year is calculated as the average  
272 of daily values from April 24 to May 7, also averaged across all basin grid cells.  
273 Average simulated SWE largely tracks the interannual variations in measured end



274 of season SWE over the period 2000–2010, with an average difference of 5.3 mm  
275 or 4.8% of the average (109.7 mm) from the field measurements (Figure S4). The  
276 Pearson correlation coefficient is  $r = 0.78$ , with the relationship significant at  $p <$   
277  $0.01$  (Figure S5).

## 278 **3.3 Runoff and river discharge**

### 279 **3.3.1 Spring freshet**

280 Modeled runoff (R) from the simulation forced with MERRA\* is evaluated against  
281 observed R for the Colville and Kupaaruk River watersheds. USGS measurements for  
282 the Kupaaruk River at Deadhorse over the period 1981–2010 show that an average  
283 of 98.3 mm of runoff (R) is exported as discharge during the spring freshet, which  
284 we calculate as R occurring from day of year (DOY) 100 to 180 (Figure 2, 3b).  
285 Simulated R over the freshet period totals 98.0 mm. Simulated May R exceeds  
286 observed R by  $29 \text{ mm month}^{-1}$ , while simulated June R is  $29 \text{ mm month}^{-1}$  lower  
287 than observed R, resulting in the relatively small error (percent difference  $+0.3\%$ )  
288 for total R over the freshet period. Simulated R closely tracks observed R in other  
289 months of the year with flow (Figure 2). For the Colville River, the available data  
290 beginning in late May show that the total volume simulated over the spring freshet is  
291 well captured, with average error of 10% (Figure 3a). Simulated R is underestimated  
292 in summer. The timing of simulated maximum daily Q closely matches the timing  
293 based on the measured data (Figure 3a). For the Kupaaruk River simulated discharge  
294 leads observed discharge by approximately one week ( $-7.8$  days, Figure 3b). For  
295 this region the flow routing sub-model is relatively insensitive to the specified flow  
296 velocity. Two sensitivity simulations using a velocity 33% lower and 33% higher  
297 than the default velocity ( $v = 0.175 \text{ m}^3 \text{ s}^{-1}$ ) resulted in errors of  $-5.4$  and  $-9.0$   
298 days respectively. Many of the rivers in this region are shorter than the Kupaaruk, so  
299 travel times are relatively brief.

### 300 **3.3.2 Annual runoff**

301 For the Kupaaruk River annual total R as the long-term (30 yr) average from USGS  
302 observations and from the model simulation is  $144$  and  $134 \text{ mm yr}^{-1}$ , respectively  
303 (percent difference =  $-6.8\%$ ) (Figure 4). Annual R from the simulation is correlated  
304 with observed annual R (Pearson correlation  $r = 0.74$ ,  $p < 0.001$ ), with average  
305 error of  $+3.1 \text{ mm yr}^{-1}$  (Figure S6). Observed R varies from  $75$ – $238 \text{ mm yr}^{-1}$ , while  
306 simulated R is more conservative, extending over a range from  $90$ – $200 \text{ mm yr}^{-1}$ . In  
307 other words, the model tends to overestimate R in years when observations are high

308 and underestimate R in years with low observed flow. For measured R partitioned  
309 at:  $R < 100 \text{ mm yr}^{-1}$ ,  $100 \leq R \leq 200 \text{ mm yr}^{-1}$ , and  $R > 200 \text{ mm yr}^{-1}$ , average  
310 errors are +24.5, -1.8, and -52.2  $\text{mm yr}^{-1}$ , respectively. It is notable that in both  
311 1996 and 2003 annual R is higher in the year following a peak (within a several year  
312 span) in annual P. This lag highlights the role that antecedent storage plays in the  
313 region’s river discharge regimes, and is consistent with previous research (Bowling  
314 et al., 2003; Stuefer et al., 2017).

## 315 4 Baseline Hydrology and Assessment of Changes

### 316 4.1 Annual precipitation and river discharge

317 For the period 1981–2010 annual total P averaged across the North Slope drainage  
318 basin ranged from 195  $\text{mm yr}^{-1}$  (1990) to 383  $\text{mm yr}^{-1}$  (2003) based on the adjusted  
319 MERRA\* P data. Annual total P over the Kuparuk Basin varied from 182  $\text{mm yr}^{-1}$   
320 (2007) to 433  $\text{mm yr}^{-1}$  (2003) (Figure 4). There is no significant trend in observed  
321 or simulated annual P or R for the Kuparuk (Figure 4) or any other river over the  
322 30 yr period. Much higher annual runoff has been documented for the Kuparuk  
323 River in 2013, 2014, and 2015 (Stuefer et al., 2017). The spatial pattern in annual R  
324 (Figure 5a) reflects a similar gradient expressed in annual P from the coast southward  
325 into the Brooks Range, as R in this region is largely controlled by snow accumulation  
326 variations. Annual R averages over 250  $\text{mm yr}^{-1}$  across parts of the Brooks Range,  
327 while coastal areas average under 100  $\text{mm yr}^{-1}$ .

328 Simulated R is routed through the STN and expressed as a volume flux of river  
329 discharge (Q) at the Beaufort Sea coast. There is a notable absence of routine moni-  
330 toring of Q at river outlets near the coast. The Colville, Kuparuk, and Sagavanirktok  
331 Rivers are the three largest gauged North Slope rivers and occupy 46.2% of the study  
332 domain. Measurements for the Kuparuk River at Deadhorse are year round since  
333 the 1970s and capture flow from most of the basin. Data for the Colville at Umiat  
334 are available from late May until early October since 2002, but Q from just 56%  
335 of the full basin area flows past the gauge location. Data for the Sagavanirktok at  
336 Pump Station 3 are available from June through September since 1995. This gauge  
337 site is located far from the coast and captures Q from only 30% of the basin. Given  
338 these constraints we estimate baseline Q exports using the observed data for the  
339 Kuparuk River, a composite of measured data and model simulation for subbasins  
340 of the Colville, and simulated Q for the remainder of the study domain.

341 Annual Q (1981–2010) for the Kuparuk River based on the USGS observations is  
342 1.4  $\text{km}^3 \text{ yr}^{-1}$  (144  $\text{mm yr}^{-1}$ ) (Table 2). The model simulated Q of 1.3  $\text{km}^3 \text{ yr}^{-1}$  closely

343 aligns with the observations and matches the  $1.3 \text{ km}^3 \text{ yr}^{-1}$  for 2000–2007 reported by  
344 McClelland et al. (2014) based on model simulations using Catchment Based Land  
345 Surface Model (CLSM). We leverage the measured data for the Colville River at  
346 Umiat ( $36,447 \text{ km}^2$ ) to estimate total Q for the entire ( $60,095 \text{ km}^2$ ) Colville River  
347 basin. A data-model composite for the subbasin defined by the gauge at Umiat (area  
348 =  $36,447 \text{ km}^2$ ) is calculated from the daily averages using measured Q when available  
349 (DOY 147 to 275) and simulated Q for the remainder of the year (Figure 3a). This  
350 gives a total Q of  $9.2 \text{ km}^3 \text{ yr}^{-1}$  ( $251 \text{ mm yr}^{-1}$ ). For the ungauged section of the  
351 basin ( $27,648 \text{ km}^2$ ) we bias adjust simulated monthly 2002–2010 R in months July,  
352 August and September assuming the ratio of simulated to observed at Umiat applies  
353 to the lower subbasin. This scaling for the ungaged subbasin produces  $4.8 \text{ km}^3 \text{ yr}^{-1}$ ,  
354 and combined with the discharge volume for the Umiat subbasin of  $9.2 \text{ km}^3 \text{ yr}^{-1}$   
355 gives  $14.0 \text{ km}^3 \text{ yr}^{-1}$  for the full basin (Table 2). This estimate compares favorably  
356 to the  $16 \text{ km}^3 \text{ yr}^{-1}$  described by Arnborg et al. (1966) based on measurements in  
357 1962, and is lower than the  $19.7 \text{ km}^3 \text{ yr}^{-1}$  (2000–2007) from McClelland et al. (2014).  
358 PWBM simulated Q (1981–2010) for the Sagavanirktok of  $3.0 \text{ km}^3 \text{ yr}^{-1}$  is bracketed  
359 by the  $1.6 \text{ km}^3 \text{ yr}^{-1}$  for 2000–2007 estimated by McClelland et al. (2014) and the  
360  $6.5 \text{ km}^3 \text{ yr}^{-1}$  for 1971–2001 estimated by Rember and Trefry (2004) using USGS  
361 data. Our composite estimate for the Colville ( $14.0 \text{ km}^3 \text{ yr}^{-1}$ ), measured Q for the  
362 Kuparuk ( $1.4 \text{ km}^3 \text{ yr}^{-1}$ ) and modeled Q for the Sagavanirktok ( $3.0 \text{ km}^3 \text{ yr}^{-1}$ ) totals  
363  $18.4 \text{ km}^3 \text{ yr}^{-1}$  for the three rivers combined, which is 57.7% of North Slope domain  
364 total annual Q of  $31.9 \text{ km}^3 \text{ yr}^{-1}$  (Table 2).

## 365 4.2 Cold season discharge (CSD)

366 Cold season (Nov–Apr) discharge (CSD) from the region simulated over the pe-  
367 riod 1981–2010 ( $0.116 \text{ km}^3 \text{ season}^{-1}$ ) is 0.4% of annual total Q, and between 0.2–0.3%  
368 for each of the Colville, Kuparuk, and Sagavanirktok rivers. In this region nearly all  
369 of the CSD occurs during the first half of winter, namely November and December.  
370 CSD for the entire North Slope basin, and both the Colville and Kuparuk rivers,  
371 increased significantly (Mann-Kendall test,  $p < 0.05$ , Table 2, Figure 6). The CSD  
372 increase from the Colville is 215% of the long-term average. For the North Slope  
373 basin as a whole CSD increased 134% of the long-term average. Increasing CSD is  
374 noted for 9.0% of the North Slope domain, and 28.4% of the Colville basin, primarily  
375 in headwater catchments of the foothills of the Brooks Range (Figure 5b). In total  
376 the affected terrain covers  $88,601 \text{ km}^2$  or 45% of the North Slope drainage.

### 377 4.3 Fraction of subsurface runoff

378 We examine variations in modeled surface and subsurface R through the year to  
379 better understand how warming is altering the hydrological flows. For the region as  
380 a whole the fraction of subsurface runoff to total runoff (hereafter ( $F_{sub}$ ) increased  
381 4.4% ( $p < 0.01$ ), a 31% change relative to the 30 yr average of 14%. Both the  
382 Colville and Sagavanirktok rivers show statistically significant ( $p < 0.05$ ) increases  
383 in  $F_{sub}$ , as do 20 of the 40 remaining basins. Significant increases are noted during  
384 several months, most widespread in September (58 of 312 grid cells, 18.6% of domain)  
385 (Figure 7). Conversely, July shows a decrease in  $F_{sub}$ , although over less total area  
386 (5.4% of domain). For June and September the  $F_{sub}$  increases average 34.8 and  
387 40.2% respectively for the total change over the period. For July the average is  
388  $-38.3\%$ , with 17 grids showing a decrease and two an increase. At the annual time  
389 scale the increase in  $F_{sub}$  is significant for 24.7% of the study domain, most notably  
390 across the northern foothills of the Brooks Range from the western part of the region  
391 (Colville basin) eastward and toward the coast (Figure 8).  $F_{sub}$  is consistently 100%  
392 of total runoff after October. Areas with increasing  $F_{sub}$  are co-located with the areas  
393 experiencing increasing CSD.

394 Increasing  $F_{sub}$  is noted in areas with a significant increase in active-layer thick-  
395 ness (ALT), primarily across parts of the northern foothills of the Brooks Range  
396 and the smaller basins near  $140^\circ\text{W}$  longitude (Figure 9). Statistically significant in-  
397 creases in ALT have been widespread, noted across two thirds (66.7%) of the region.  
398 The simulation shows that one fifth (20.2%) of the region experienced a significant  
399 increase in both  $F_{sub}$  and ALT ( $p < 0.05$ , Table 3). A fraction of the foothills region  
400 (5.1% of domain) is characterized by a positive trend in  $F_{sub}$  only. The ALT trend  
401 average for grid cells with a significant increase in  $F_{sub}$  only, a significant increase  
402 in ALT only, and a significant increase in both are 0.17, 0.75, and 1.00  $\text{cm yr}^{-1}$ ,  
403 respectively (Figure 10, Table 3). These relatively large ALT increases in areas of  
404 significant  $F_{sub}$  increase indicate a connection between enhanced permafrost thaw  
405 and subsurface water flow in those areas.

### 406 4.4 Terrestrial water storage

407 Terrestrial water storage (TWS) over a given time interval is defined by the total  
408 amount of water stored in snow, soil liquid water, and soil ice as estimated by the  
409 model simulation. Over the 1981–2010 period annual average TWS (all 312 domain  
410 grids) exhibits a negative trend of approximately  $-2 \text{ mm yr}^{-1}$  ( $p < 0.001$ , Figure 11).  
411 Declines in annual minimum ( $-1.7 \text{ mm yr}^{-1}$ ) and maximum TWS ( $-2.3 \text{ mm yr}^{-1}$ )  
412 are also significant. Among the component storages there is no significant change in

413 SWE over the 30 year period (Figure S7). Increases in regionally averaged maximum  
414 and minimum soil liquid water, and decreases in soil ice amounts, are significant ( $p$   
415  $< 0.01$ , modified Mann-Kendall test). The  $-2 \text{ mm yr}^{-1}$  decrease in TWS reflects a  
416 decrease in soil ice storage of  $-2.5 \text{ mm yr}^{-1}$ , a decline in SWE of  $-0.16 \text{ mm yr}^{-1}$ ,  
417 and an increase in soil water storage of  $0.61 \text{ mm yr}^{-1}$ .

## 418 4.5 Timing of maximum daily discharge

419 Warming and associated changes in snowmelt have the potential to cause shifts in  
420 the timing of peak discharge ( $Q$ ) during the spring freshet period. Maximum spring  
421 discharge is determined from the daily model simulated and routed  $Q$  for each of the  
422 42 North Slope domain rivers. In the simulation only one of the 42 basins exhibits  
423 a significant shift to earlier maximum daily  $Q$ . None show a significant shift to later  
424 maximum  $Q$ . While many rivers show simulated peak discharge shifting nearly one  
425 week earlier over the 30 yr period, high interannual variability in annual  $Q$  renders  
426 the changes insignificant at the 95% level. The average date of maximum daily  $Q$   
427 across the 42 basin advanced by approximately 4.5 days (Figure S8), though the  
428 change is only marginally significant ( $p = 0.1$ ). Maximum daily  $Q$  from the region in  
429 recent years occurs near DOY 150 (end of May), though this estimate is potentially  
430 biased based on the comparison of simulated and observed runoff for the Kuparuk  
431 River (subsection 3.3).

## 432 5 Summary and Discussion

433 Recent studies have investigated how hydrological cycle intensification and per-  
434 mafrost thaw may alter terrestrial hydrological fluxes and, in turn, materials export  
435 to coastal zones (Walvoord and Striegl, 2007; Frey and McClelland, 2009; Rawl-  
436 ins et al., 2010; Spencer et al., 2015; Vonk et al., 2015). Changes unfolding across  
437 high latitude watersheds have the potential to significantly alter water, carbon, and  
438 other constituent fluxes, with implications for nearshore arctic biogeochemical and  
439 ecological processes.

440 Our synthesis of measured data and model simulations reveals that approximately  
441  $32 \text{ km}^3 \text{ yr}^{-1}$  of freshwater is exported by the region's rivers, with 57.7% of the total  
442 originating from the Colville, Kuparuk, and Sagavanirktok Rivers. Simulated runoff  
443 for the Kuparuk River shows maximum daily spring discharge that exhibits a sys-  
444 tematic bias of approximately 8 days early relative to gauge data. Timing is well  
445 estimated for the Colville River. The timing bias for the Kuparuk is unrelated to the

446 specification of river flow velocity in the routing scheme, and likely due to a combina-  
447 tion of errors in air temperature forcing or modeled snowmelt processes (warm bias)  
448 that lead to early snowpack thaw, and/or insufficient surface storages in the model  
449 which serve to delay the transfer of water to stream networks. Simulated R timing  
450 may improve by better accounting for these lags in snowmelt runoff. Future studies  
451 should investigate how dynamic surface inundation data obtained from microwave  
452 and radar remote sensing (Schroeder et al., 2010; Du et al., 2016) can be used to  
453 constrain surface water storage, its partitioning to runoff and evaporation, and flow  
454 direction in areas of low topographic relief. The lag in annual runoff for the Kuparuk  
455 River in 1996 and 2003 highlight how precipitation and antecedent storage conditions  
456 can influence the following year’s runoff (Bowling et al., 2003; Stuefer et al., 2017).

457 The quantity and quality of freshwater export is expected to change significantly  
458 as the Arctic hydrological cycle intensifies and the system transitions toward increas-  
459 ing groundwater water flows (Frey et al., 2003; Frey and McClelland, 2009). In this  
460 study evidence of change is evident in cold season discharge from the North Slope  
461 region over the 30 year (1981–2010) period examined. There is no significant trend  
462 in annual total discharge for the region or its rivers. However, we note that the Ku-  
463 paruk and nearby Putuligayuk River experienced high annual runoff in 2013, 2014,  
464 and 2015 (Stuefer et al., 2017), consistent with expectations under an intensifying  
465 arctic hydrological cycle (Wu et al., 2005; Rawlins et al., 2010). Climate models  
466 project a future increase in Arctic precipitation that is generally greatest in autumn  
467 and winter and smallest in summer, and greatest over the higher latitudes of Eura-  
468 sia and North America (ACIA, 2005; Kattsov et al., 2007). Higher winter snowfall  
469 across the North Slope would likely lead to increased freshwater discharges. The  
470 model simulation shows increases in cold season discharge of 134% and 215% of the  
471 long-term average for the North Slope (domain total) and Colville River, respectively.  
472 Basins showing a significant increase in cold season discharge cover 45% of the re-  
473 gion. Within the Colville basin the changes are greatest in headwater catchments of  
474 the northern foothills and mountains of the Brooks Range (Figure 5b). Landscape  
475 conditions in those areas strongly influence the quality of water exported during the  
476 first half of winter, including the solubility, chemical character, and biodegradability  
477 of carbon, nitrogen and other nutrients (Wickland et al., 2018). Effects of permafrost  
478 thaw on soil infiltration, flowpath length, and subsurface water movement has been  
479 identified in the observed rise in low flows in parts of the Arctic (St. Jacques and  
480 Sauchyn, 2009; Smith et al., 2007; Walvoord and Striegl, 2007). The controls per-  
481 mafrost exerts have been implicated in the observed increase in the ratio of maximum  
482 to minimum monthly discharge in the continuous permafrost regions of the middle  
483 and lower Lena River basin (Gautier et al., 2018), linked with increased CSD from

484 1935–1999 (Yang et al., 2002). More broadly, cold-season low-flow is increasing over  
485 most of the pan-arctic (Rennermalm et al., 2010).

486 Our results also show changes in the proportion of groundwater runoff for the  
487 region as a whole, and individually the Colville, Sagavanirktok, and 22 of the other  
488 40 river basins. Increases are noted across the foothills and higher elevations of the  
489 northern Brooks Range. The growing subsurface flows are contributing to the in-  
490 creasing cold season discharge amounts, with the most significant changes in both  
491 quantities found across headwaters of several of the larger basins (Colville and Saga-  
492 vanirktok), as well as areas near the coast east of approximately 140°W. Increases in  
493 both subsurface runoff and cold season discharge are likely manifestations of climate  
494 warming, as active layer thaw depths are highly responsive to warming air temper-  
495 atures (Hinkel and Nelson, 2003). Approximately 20% of the region, the Brooks  
496 Range foothills and smaller watersheds near 140°W, shows significant increases in  
497 both the fraction of subsurface runoff and active layer thickness. The active layer  
498 increase is greatest in those areas experiencing growing subsurface runoff contribu-  
499 tions, suggesting a direct connection between thawing soils and changing subsurface  
500 flows.

501 A deepening active layer associated with climate warming will likely lead to a  
502 longer unfrozen period in deeper soils (Yi et al., 2019), enhancing subsurface runoff  
503 flow. A deeper active layer delays the soil freeze up and increases the amount of liquid  
504 pore water. A larger thawed zone permits additional water storage that supports  
505 runoff in late autumn, before soils freeze completely. The changes captured in the  
506 modeling are consistent with the notion that permafrost thaw enhances hydrogeologic  
507 connectivity and increases low flows in permafrost regions (Bense et al., 2009, 2012;  
508 Bring et al., 2016; Lamontagne-Hallé et al., 2018). Observational and modeling  
509 studies suggest that permafrost thaw can lead to increased subsurface runoff and  
510 cold season discharge, as increasing thickness of the thawed zone and shallow aquifer  
511 provide a conduit for flow to rivers (Walvoord and Striegl, 2007; Bense et al., 2009;  
512 Walvoord and Kurylyk, 2016; Lamontagne-Hallé et al., 2018). Alternatively, these  
513 change in continuous permafrost zones can also arise where permafrost is locally  
514 discontinuous, or through flow from unfrozen surface water bodies.

515 Evidence of permafrost thaw and increasing groundwater flow has been reported  
516 in studies using measurements from arctic rivers. Recent increases in nitrate concen-  
517 trations and export from the Kuparuk River are consistent with permafrost degrada-  
518 tion and deepening flow paths (McClelland et al., 2007). 'Old' carbon measured in  
519 Arctic rivers indicates mobilization of pre-industrial organic matter and subsequent  
520 transfer to rivers (Schuur et al., 2009; Mann et al., 2015; Dean et al., 2018). St.  
521 Jacques and Sauchyn (2009) concluded that increases in winter baseflow and mean

522 annual streamflow in the NWT were caused predominately by climate warming via  
523 permafrost thawing that enhances infiltration and deeper flowpaths and hydrological  
524 cycle intensification (Frey and McClelland, 2009; Bring et al., 2016). The magni-  
525 tude of subsurface runoff change in the present study should be viewed with caution  
526 given the intrinsic resolution of model parameterizations for soil texture, organic  
527 layer thickness, and other landscape properties. Our results, however, do point to a  
528 close correspondence between active layer thickness and subsurface runoff increases  
529 across the foothills of the Brooks Range. The enhanced changes there suggest that  
530 the relatively thin surface organic layer and sandy soils in the foothills areas may  
531 be seeing a relative larger impact on soil warming and thaw. Our results thus lend  
532 additional support to findings in other recent studies pointing to bigger impacts of  
533 warming on permafrost thaw in areas with relatively low vegetation and low soil  
534 organic content (Yi et al., 2019; Jones et al., 2019). For example, Yi et al. (2019),  
535 using the PWBm in a modeling framework driven with data from remote sensing  
536 observations, found that ALT deepening across much of the Brooks Range has been  
537 greater than in the tundra to the north (Yi et al., 2018).

538 Consistent with recent warming and associated ALT increases, our results sug-  
539 gest an overall decline ( $-2 \text{ mm yr}^2$ ) in terrestrial water storage across the North  
540 Slope drainage basin over the 1981–2010 period. This decrease is driven by losses in  
541 soil ice, with an increase in liquid water storage which does not fully offset the ice  
542 losses. With continued warming it is likely that the timing of snowmelt will advance,  
543 with impacts to the timing of peak (maximum daily) spring discharge. Averaged  
544 across all 42 basins, the date of daily maximum discharge advanced 4.5 days over the  
545 1981–2010 period, though the change is only marginally significant ( $p = 0.1$ ) at the  
546 95% confidence level. Individual river basins show larger shifts to earlier maximum  
547 daily discharge. Future changes toward earlier peak discharge can be expected given  
548 projections of future warming.

549 Modeling studies of the impacts of climate warming on permafrost thaw and  
550 groundwater discharge are key to our understanding of lateral hydrological flows and  
551 associated constituent exports. The underestimate in summer runoff for the Colville  
552 River is likely attributable to errors in the meteorological forcings and the model  
553 simulation of fluxes including snow sublimation and evapotranspiration. Solid pre-  
554 cipitation observations in this region are highly uncertain (Scaff et al., 2015), and  
555 this lack of information hinders verification of reanalysis precipitation products and  
556 associated studies of changes in seasonal precipitation, which may be playing a role  
557 in the hydrological alterations. Results of this study should be corroborated through  
558 evaluation of simulations produced with alternate forcings and through parameter  
559 sensitivity analysis. The good agreement for the Kuparuk River and the underesti-



560 mate in simulated discharge for the Umiat subbasin of the Colville point to the need  
561 for improved estimates of precipitation across higher elevations of the Brooks Range.  
562 A fuller understanding of the extent of water cycle alterations in this region will  
563 require new observations of river discharge, precipitation, snow storage, soil mois-  
564 ture and other key variables needed to parameterize and validate numerical models,  
565 including those which capture the important role ground ice plays in runoff gen-  
566 erating processes. Data being gathered within the region’s watersheds and coastal  
567 environments can provide important information for model parametrization and ver-  
568 ification. Measurements of river discharge and dissolved organic carbon at multiple  
569 locations along the coast are critical to an improved understanding of land-ocean  
570 carbon exports. Regarding linkages with biogeochemical fluxes, water samples from  
571 the mouths of major Arctic river show that dissolved organic carbon in those rivers is  
572 sourced primarily from fresh vegetation during the two month of spring freshet and  
573 from older, soil-, peat-, and wetland-derived DOC during groundwater dominated  
574 low flow conditions (Amon et al., 2012). Stable isotope data obtained from river  
575 water samples can be used to guide partitioning of surface and groundwater water  
576 flows to better understand how soil drainage and soil moisture redistribution will  
577 change with future permafrost thaw and ALT deepening (Walvoord and Kurylyk,  
578 2016).

579 High performance computing is helping to provide insights into hydrological flows  
580 and biogeochemical cycling in arctic environments (Lamontagne-Hallé et al., 2018;  
581 Neilson et al., 2018). Improvements in numerical model simulations of groundwater  
582 flow regimes in permafrost areas have provided insights on the important roles that  
583 microtopography and soil properties play in groundwater runoff regimes. Model cal-  
584 ibration and validation for simulations at finer spatial scales is dependent on new  
585 field measurements of parameters such as water table height, active layer thickness,  
586 and soil organic carbon content with depth. Simulations for future conditions in  
587 the region should take into account processes directly influenced by permafrost thaw  
588 (Bense et al., 2012; Lamontagne-Hallé et al., 2018). To overcome challenges in de-  
589 riving parameterization from multiple disparate data sets, high-resolution ecosystem  
590 maps of the Alaska North Slope can provide a convenient upscaling mechanism to  
591 parameterize ground soil properties across the region (Nicolosky et al., 2017). Given  
592 its considerable effect on soil thermal and hydraulic properties, modeling efforts will  
593 benefit from improved mapping of soil organic matter.

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## 603 **7 Author Contributions**

604 M.A.R designed the study, executed the model simulations, and performed the  
605 analysis. L.C, S.L.S., and D.N. contributed data. M.A.R drafted the initial manuscript  
606 and all authors contributed to its development and publication.

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Table 1: Distribution statistics (cm) for spatial fields of active layer thickness (ALT) from the GIPL and PWBM simulation with MERRA\* forcing shown in Figure S3. Also shown are statistics for a simulation using original (non-adjusted) MERRA precipitation (P) data.

<b>Active Layer Thick Distribution Statistics (cm)</b>					
<b>Data</b>	<b>5<sup>th</sup></b>	<b>25<sup>th</sup></b>	<b>mean</b>	<b>75<sup>th</sup></b>	<b>95<sup>th</sup></b>
GIPL	37.3	49.9	55.2	61.4	69.4
PWBM (MERRA)	30.5	40.3	50.4	58.6	75.2
PWBM (MERRA*)	32.0	43.7	53.5	61.3	79.0

Table 2: River basin area, annual discharge (Q), and cold season discharge (CSD) for the Colville, Kuparuk, and Sagavanirktok rivers and the full North Slope domain. River basins with a significant increase in CSD are indicated with a superscript \*. Basin areas are based on their specification in the simulated topological river network.

<b>River Basin and Domain-Wide Discharge</b>			
<b>Basin</b>	<b>Area (km<sup>2</sup>)</b>	<b>Annual Q (km<sup>3</sup> yr<sup>-1</sup>)</b>	<b>CSD (km<sup>3</sup> season<sup>-1</sup>)</b>
Colville	64 095	14.0	0.023*
Kuparuk	10 054	1.4	0.004*
Sagavanirktok	16 338	3.0	0.006
3 River Total	90 487	18.4	0.032
North Slope	196 061	31.9	0.116*

Table 3: Number of grid cells, associated area fraction of domain, and average ALT and  $F_{sub}$  for each category shown. Study domain consists of 312 grid cells spanning an area of 196,060 km<sup>2</sup> (Figure 1).

<b>Number of grids, area, and ALT and <math>F_{sub}</math> averages for each subregion.</b>				
	<b>N</b>	<b>area (%)</b>	<b><math>F_{sub}</math> (%<sup>3</sup> yr<sup>-1</sup>)</b>	<b>ALT (cm yr<sup>-1</sup>)</b>
$F_{sub}$ increase only	16	5.1	0.43	0.17
ALT increase only	211	67.6	0.05	0.75
both	63	20.2	0.35	1.00
neither	22	7.1	0.22	0.22

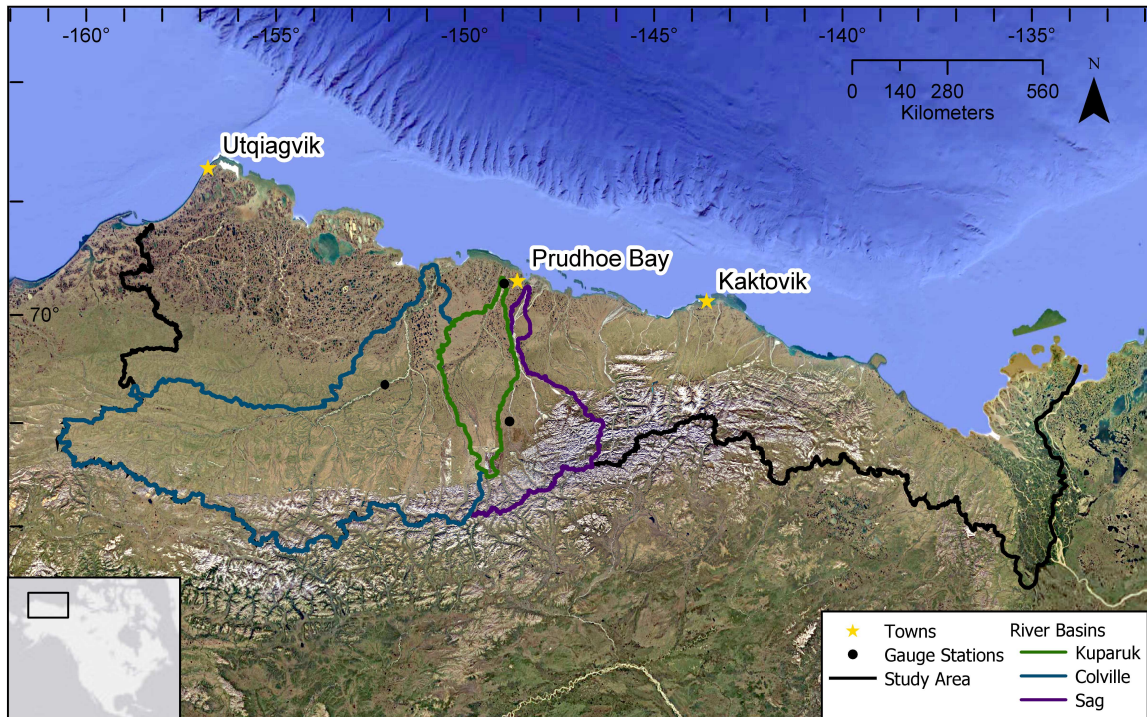


Figure 1: Study domain of North Slope of Alaska. Black line delineates the full North Slope drainage basin. This domain includes all land (196,060 km<sup>2</sup>) which drains to the Beaufort Sea coast. Blue, green, and purple lines mark boundaries for the drainage basins of the Colville, Kuparuk, and Sagavanirktok rivers, respectively. The three dots mark locations where USGS discharge measurements are obtained for each river at, respectively, Umiat, Deadhorse, and Pump Station #3. The 42 individual basins defined by the simulated topological network (STN) are listed in Table S1. Locations shown for population centers Utqiagvik, Prudhoe Bay, and Kaktovik.

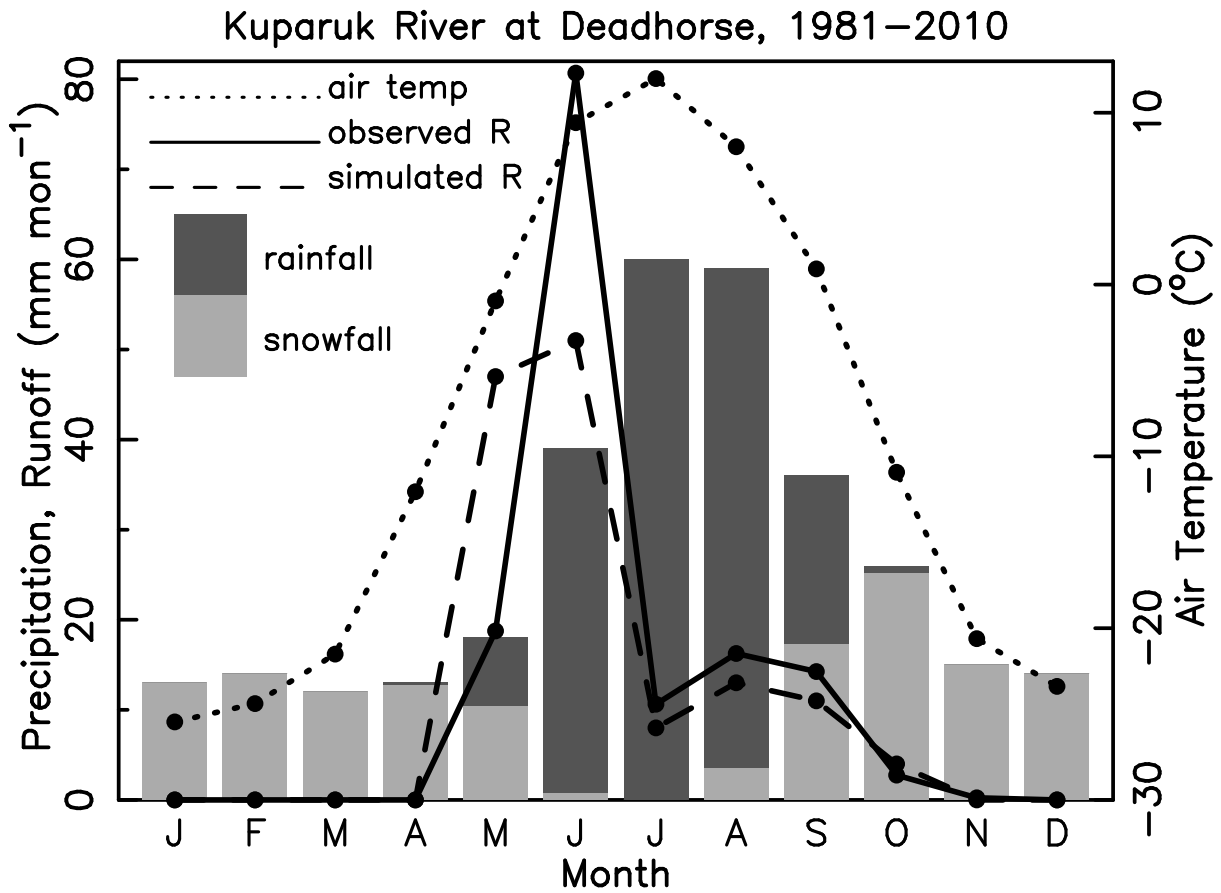


Figure 2: Simulated and observed runoff ( $R$ ,  $\text{mm month}^{-1}$ ) for the Kuparuk River basin 1981–2010. Simulated  $R$  expressed in unit depth was calculated from the routed river discharge ( $Q$ ) volume. Observed  $R$  was drawn from the USGS database (section 2.1). The PWBm simulation was forced with meteorological data from the MERRA reanalysis, with precipitation adjustment (MERRA\*) as described in section 2.2. Monthly air temperature is the average over the Kuparuk basin from the MERRA data used in the model simulation. Monthly climatological precipitation ( $P$ ) shown in totals ( $\text{mm month}^{-1}$ ) for rainfall and snowfall.



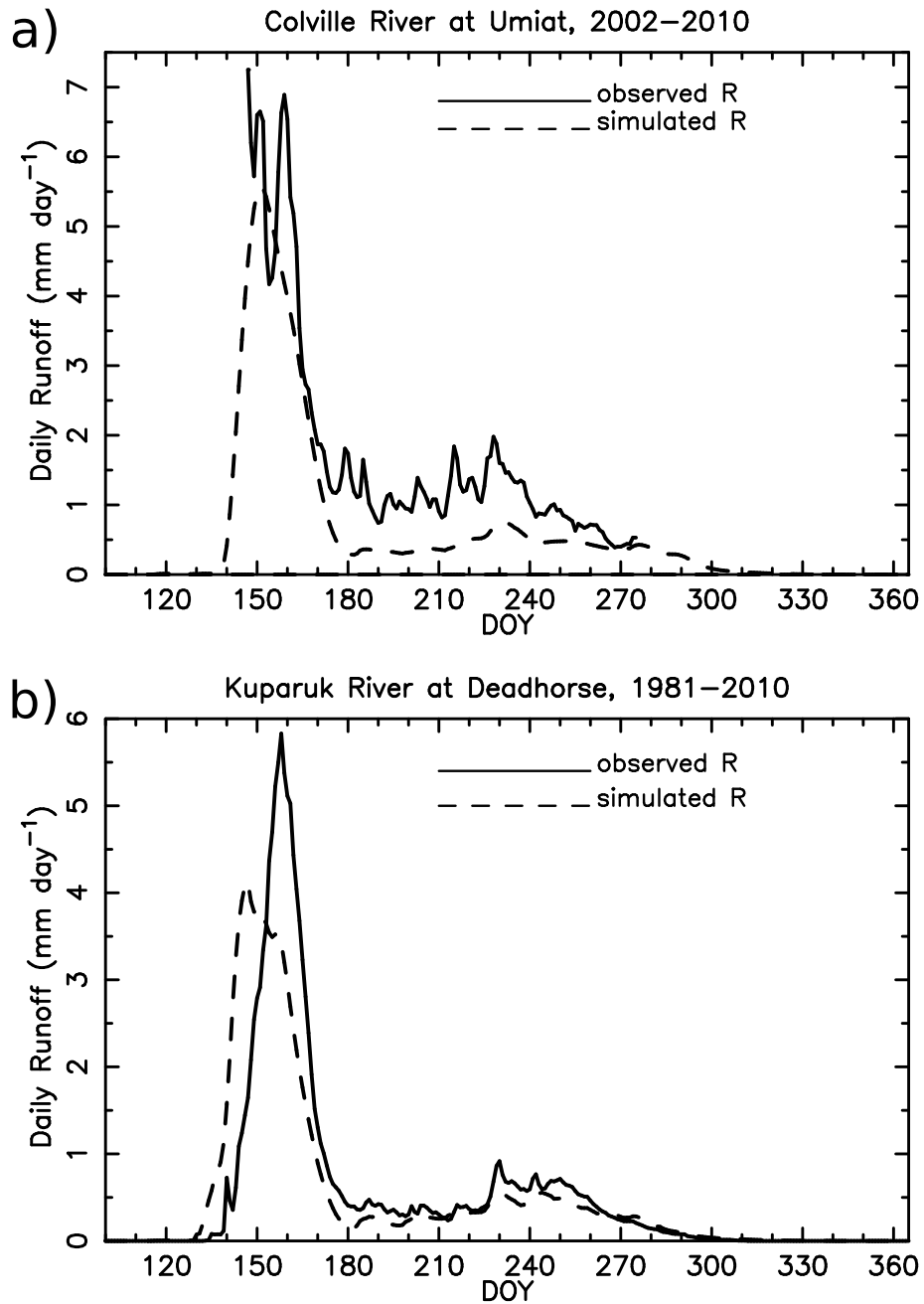


Figure 3: Simulated and observed runoff ( $R$ ,  $\text{mm day}^{-1}$ ) for the (a) Colville River at Umiat, AK and (b) Kuparuk River at Deadhorse AK. Discharge data for the Colville River published by the USGS are generally available each year from the end of May until early October. Runoff calculated as unit depth as in Figure 2.

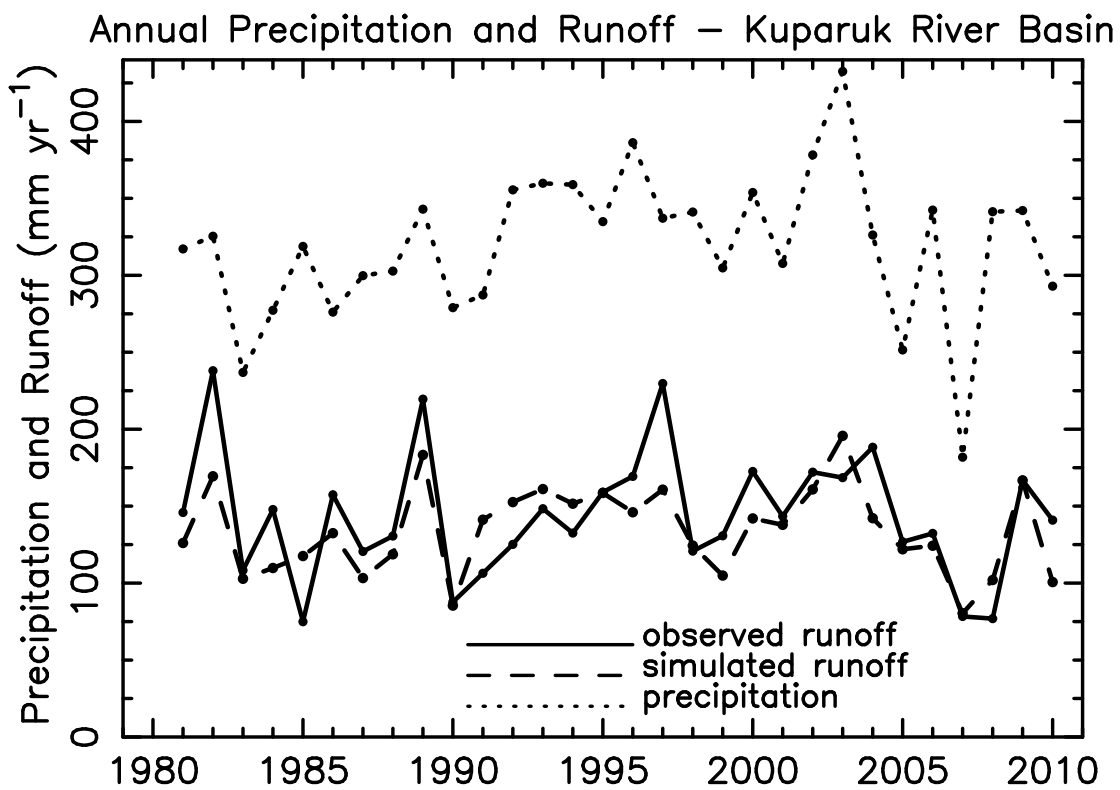


Figure 4: Annual total P from the adjusted MERRA (MERRA\*, section 2.2) and simulated and observed R (mm yr<sup>-1</sup>) for the Kuparuk River basin for the simulation period 1981–2010.

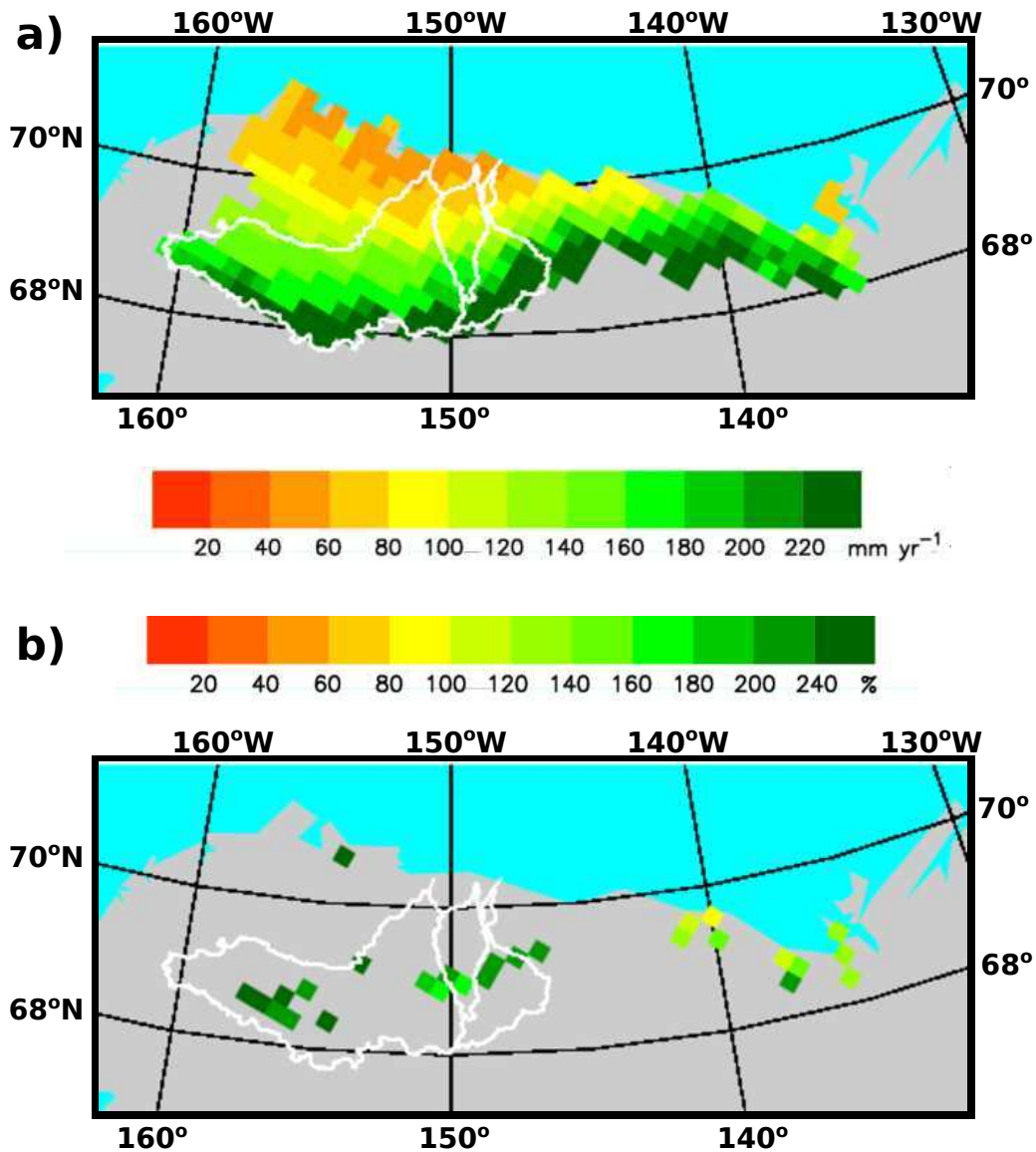


Figure 5: a) Annual total R 1981–2010 ( $\text{mm yr}^{-1}$ ) from the model simulation and b) grid cells with a statistically significant ( $p < 0.05$ ) change in simulated cold season (Nov–Apr) Q over the period 1981–2010. The change is shaded as a percentage of the 30 yr average for cold season R for that grid. White outlines are basin boundaries for the (west to east) Colville, Kuparuk, and Sagavanirktok rivers.

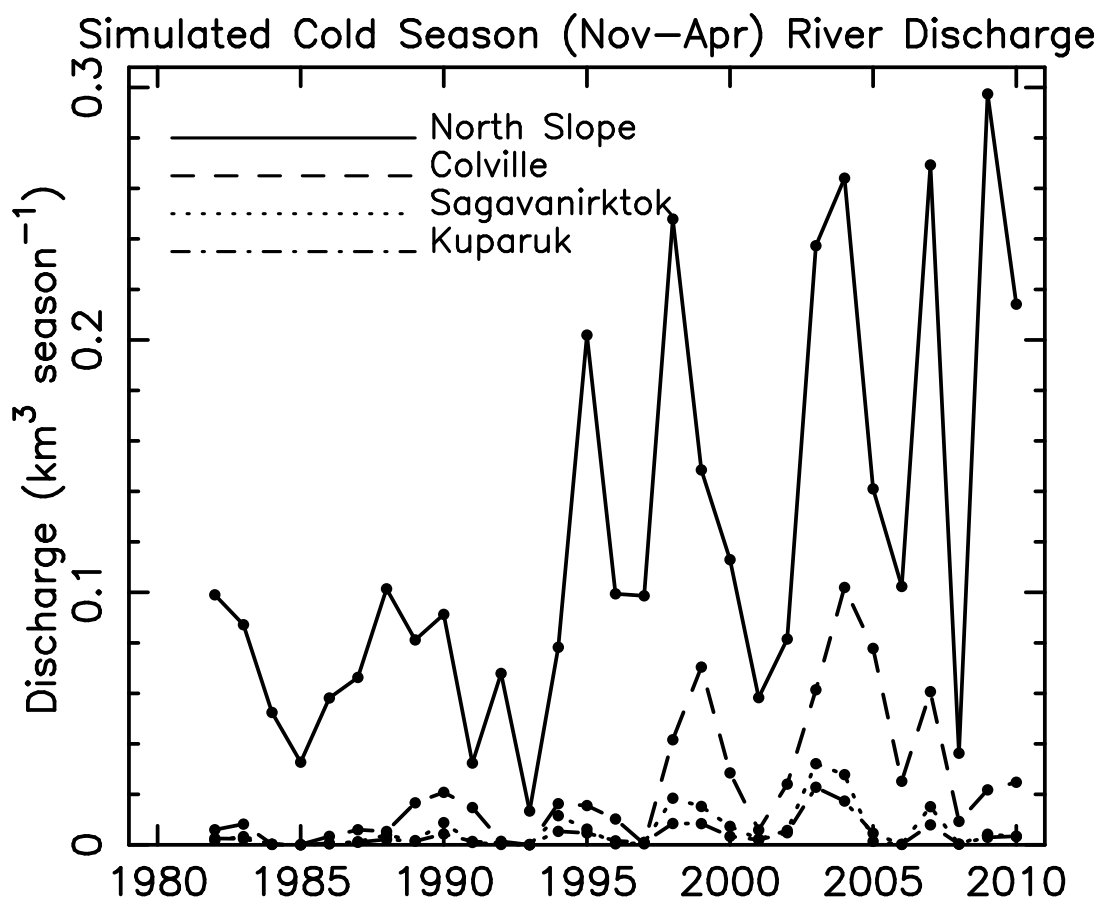


Figure 6: Simulated cold season  $Q$  ( $\text{km}^3 \text{ season}^{-1}$ ) for the full North Slope region and for separately the Colville, Sagavanirktok, and Kuparuk rivers.

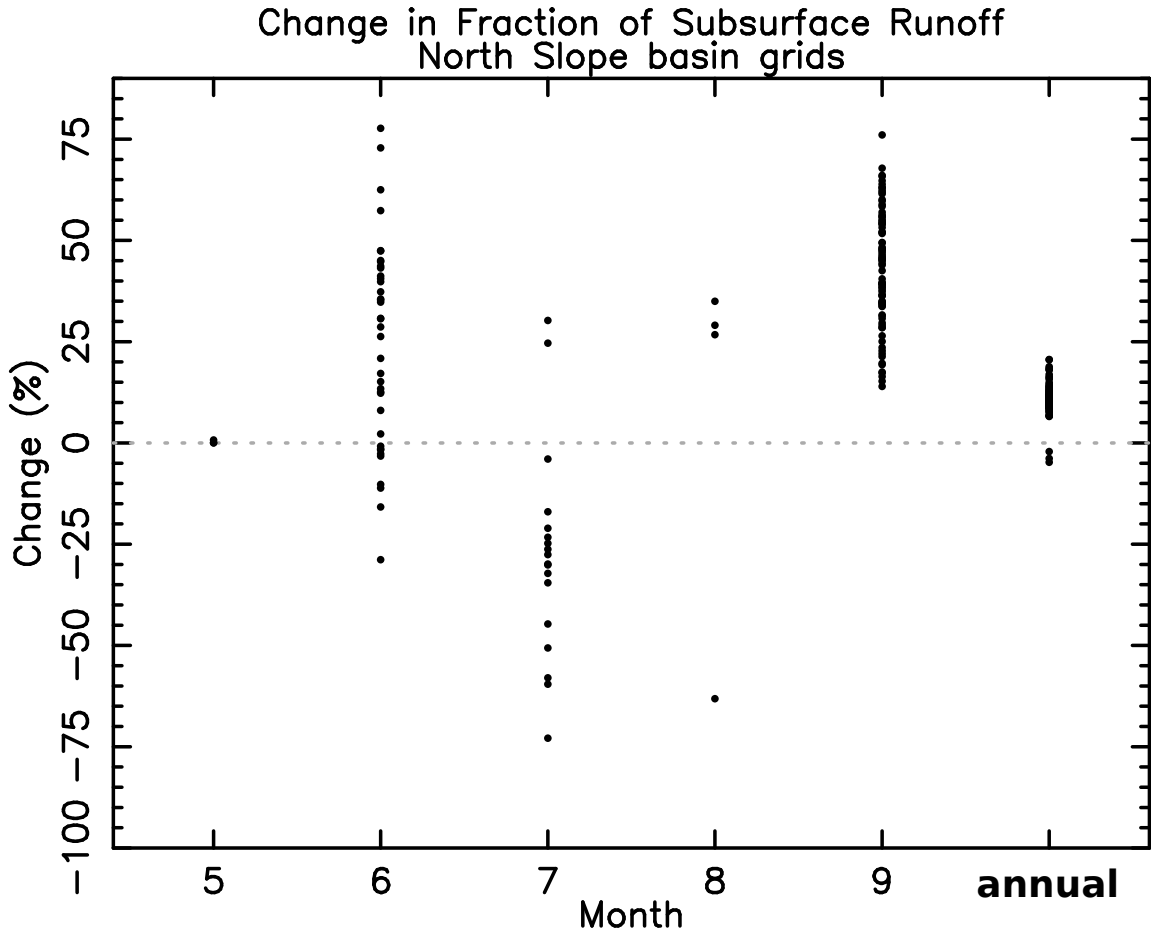


Figure 7: a) Grid cell change in fraction of subsurface R ( $F_{sub}$ ) for warm season months May–September and for annual total  $F_{sub}$  and R.  $F_{sub}$  changes are not defined for other months due to  $F_{sub}$  consistently at 100%, or the grid cell having no runoff for that month in more than 50% (15 of 30) of the data years. Change is expressed with respect to the long-term average. Dots represent grid cells that show a significant change at  $p < 0.05$ . Average for grids with a significant change at the annual scale is +11.0%

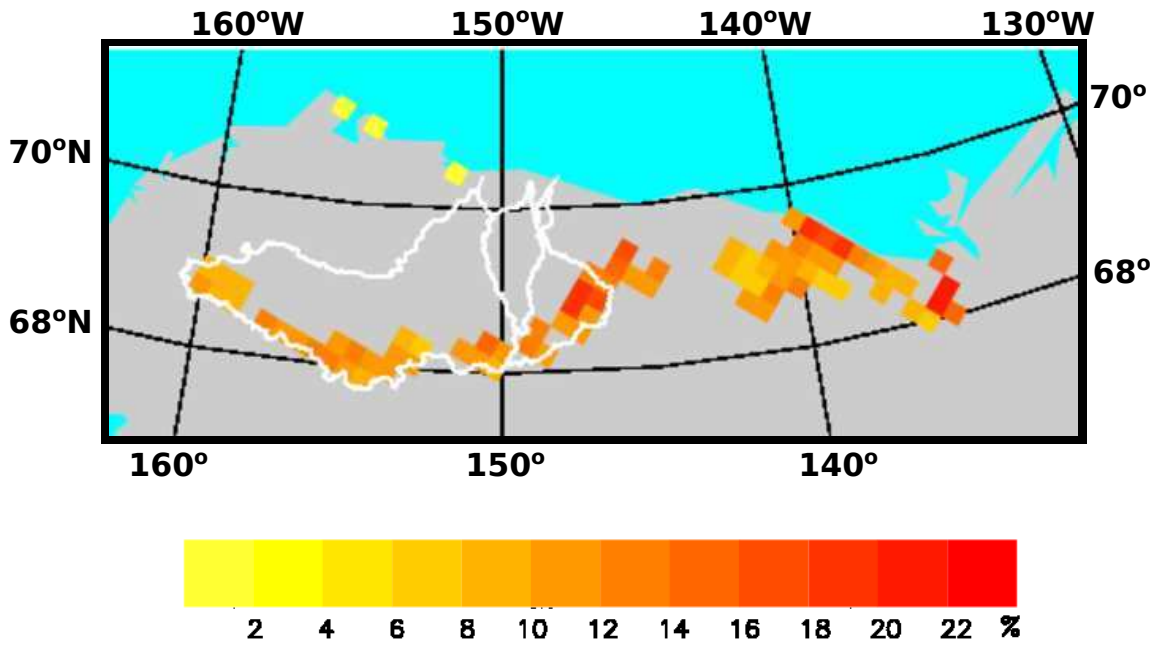


Figure 8: Change in  $F_{sub}$  (%) over the period 1981–2010. Mapped grids show a significant change at  $p < 0.05$  based on a two-sided test.

### Regions With Significant Increase in $F_{sub}$ and ALT

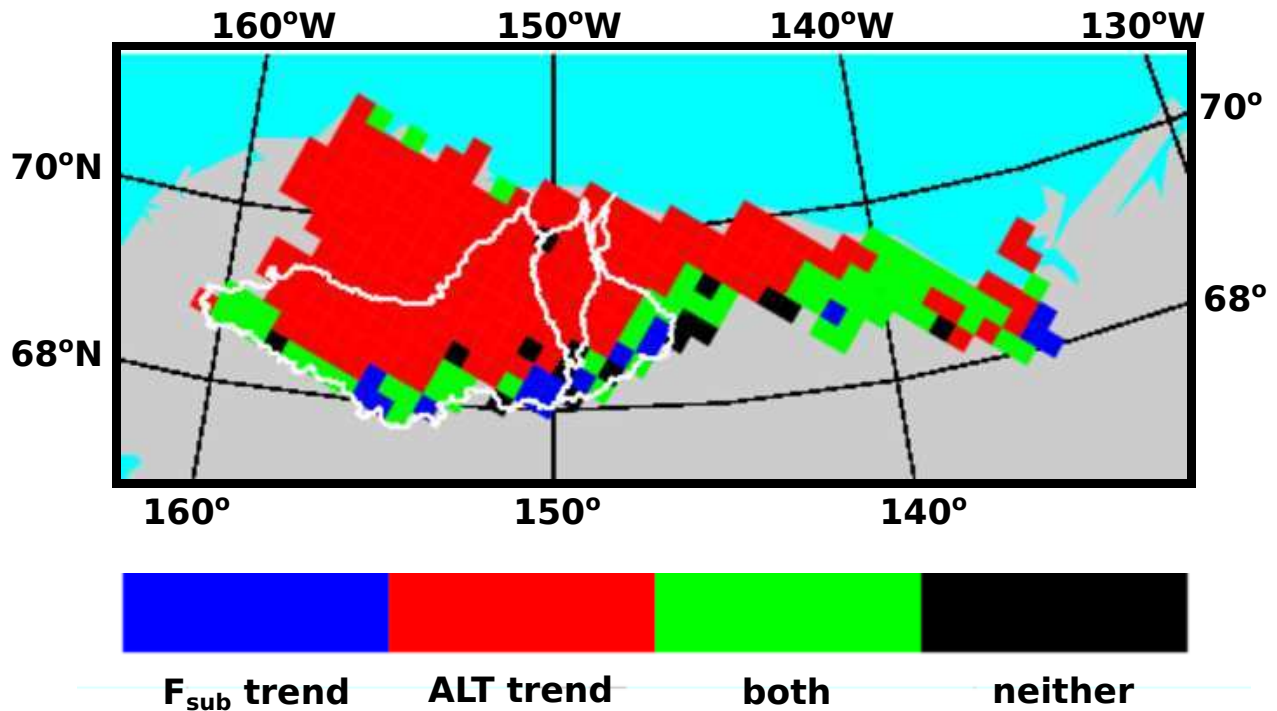


Figure 9: Spatial extent of regions showing a significant increase in annual  $F_{sub}$  only (blue), a significant increase in active layer thickness (ALT) only (red), significant increases in both (green), and neither (black). The number of grid cells, area fraction impacted, and average  $F_{sub}$  and ALT increase for each category are shown in Table 3.

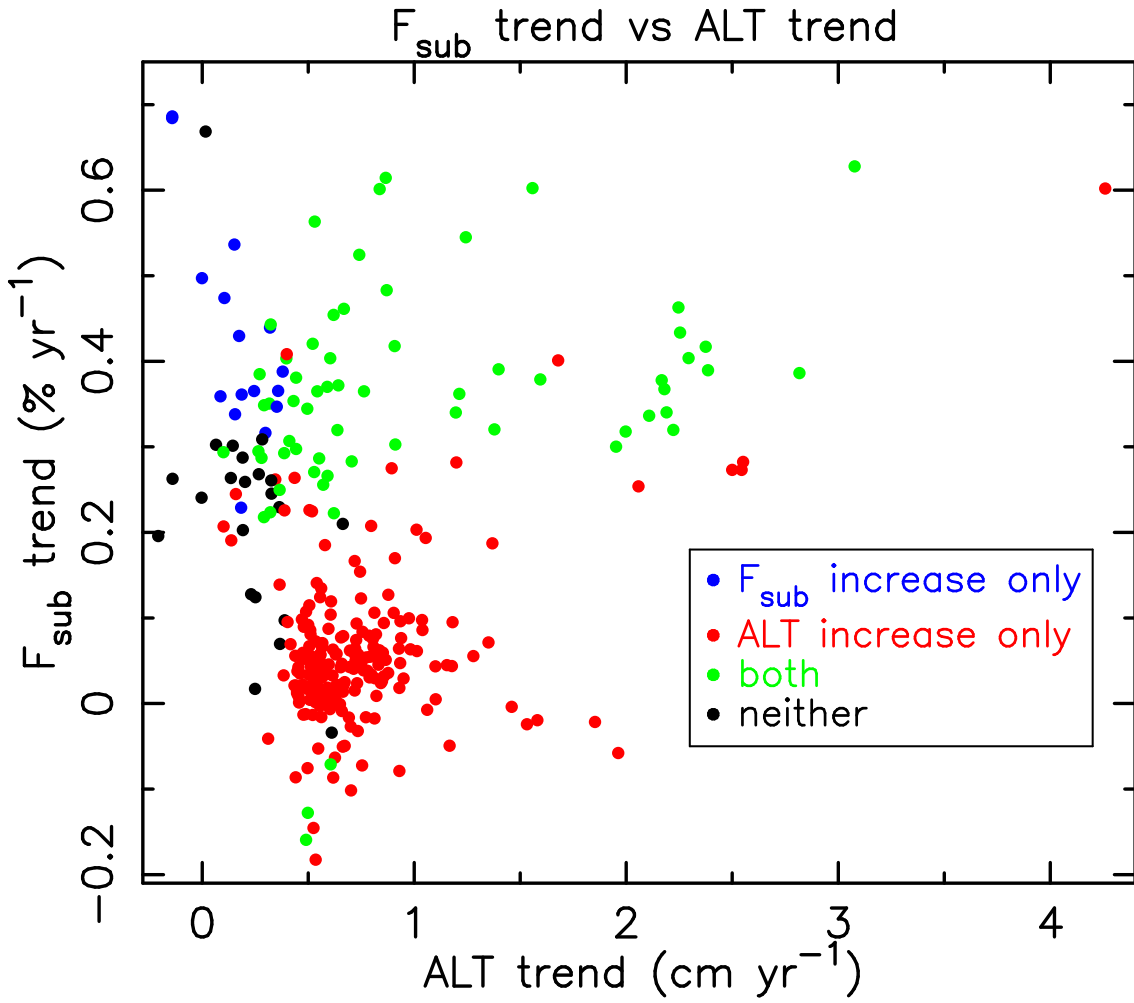


Figure 10: Increase in annual  $F_{sub}$  ( $\% \text{ yr}^{-1}$ ) vs increase in seasonal maximum ALT ( $\text{cm yr}^{-1}$ ) for all 312 domain grid cells. Relevant statistics are listed in Table 3.



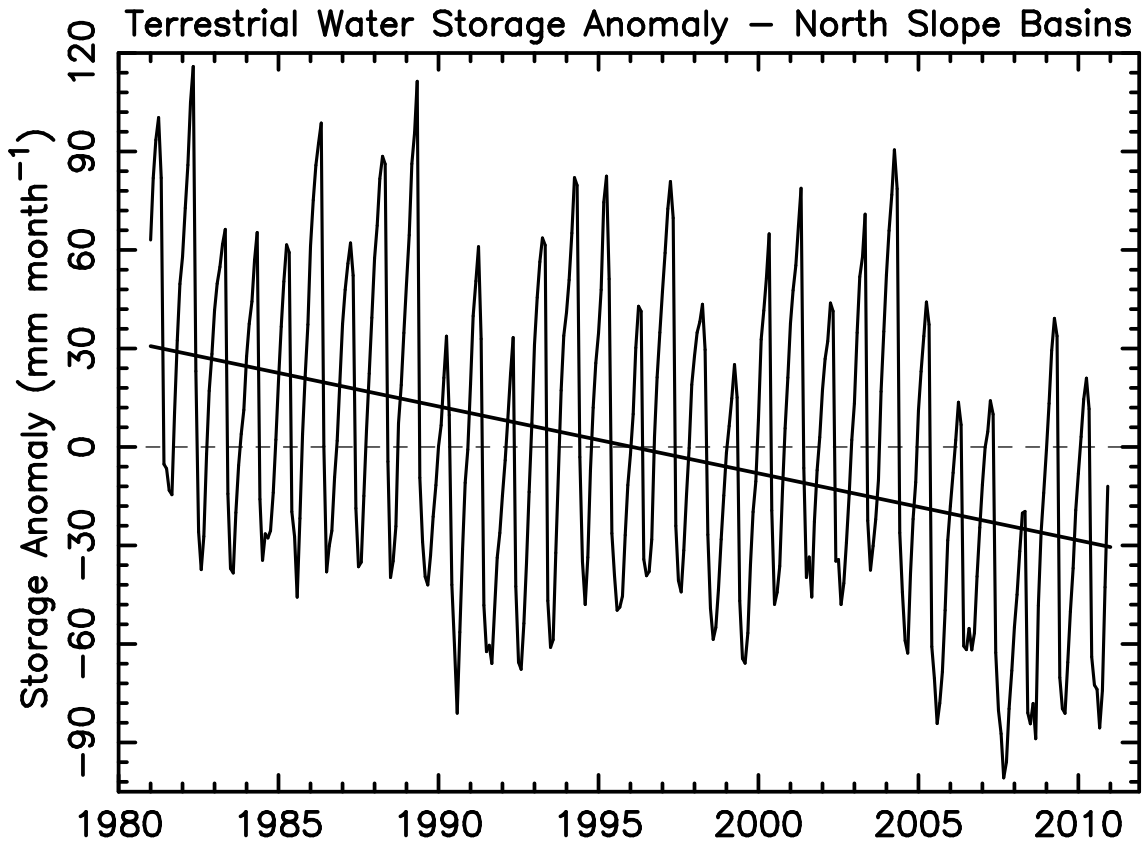


Figure 11: Terrestrial water storage (TWS) anomaly (mm month<sup>-1</sup>) as an average across the North Slope drainage basin. Anomaly is with respect to the long-term average (1981–2010). In the PWBM, TWS includes soil liquid water, ice, and snow storage. It does not include water stored in permanent water bodies such as ponds and lakes.

Table S1: River basins ordered by size for the North Slope drainage region. Basins in the simulated topological network (STN) were defined on the  $25 \times 25$  km<sup>2</sup> EASE-Grid (Brodzik and Knowles, 2002). Areas in km<sup>2</sup> based on extent in the STN of the full drainage basin expressed to the respective river mouth at the coast. Names listed for rivers with areas greater than 4000 km<sup>2</sup>. Unnamed rivers are numbered by size among all river basins in the pan-Arctic STN.

<b>Latitude</b>	<b>Longitude</b>	<b>Basin area</b>	<b>Name</b>
70.3288	-151.0736	64095	Colville
70.6501	-154.3348	18851	Ikpikpuk
70.2604	-148.1340	16338	Sagavanirktok
70.9372	-156.1757	12568	Meade
70.3802	-148.6959	10054	Kuparuk
69.4239	-139.4672	6284	Firth
70.0799	-146.1292	5655	Canning
69.8753	-144.1624	5027	Hulahula
70.0150	-147.0306	4399	Shaviovik
68.5119	-135.8551	4399	Unnamed
70.8438	-155.5560	3770	Basin 1659
69.5061	-141.7360	3142	Basin 1882
68.6613	-137.1530	3142	Basin 1896
69.9243	-143.2594	2514	Basin 1949
69.7866	-142.7447	2514	Basin 1966
69.1231	-138.5215	2514	Basin 2012
68.6711	-136.2922	2514	Basin 2041
69.6471	-142.2369	2514	Basin 2104
68.8289	-136.7357	1885	Basin 2279
68.9706	-138.0587	1885	Basin 2354
70.1386	-147.5789	1885	Basin 2463
69.5720	-139.9503	1885	Basin 2464
68.6760	-135.4308	1885	Basin 2466
71.2383	-156.5290	1257	Basin 3496
70.9549	-154.6538	1257	Basin 3497
70.3011	-149.6013	1257	Basin 3498
69.9515	-145.5915	1257	Basin 3500
69.8212	-145.0607	1257	Basin 3501
69.2742	-138.9909	1257	Basin 3503
69.3244	-135.4441	1257	Basin 3504
70.8546	-152.5256	628	Basin 4393
70.4159	-150.1729	628	Basin 4394
69.5415	-140.8446	628	Basin 4398
69.0003	-135.4374	628	Basin 4409
68.8388	-135.0000	628	Basin 4410
69.3244	-134.5559	628	Basin 4416
69.4845	-134.1048	628	Basin 4419
71.1461	-155.8978	628	Basin 6501
70.4384	-151.6543	628	Basin 6502
70.0604	-143.7812	628	Basin 6507
68.8167	-137.6026	628	Basin 6511
69.1605	-135.8814	628	Basin 6513

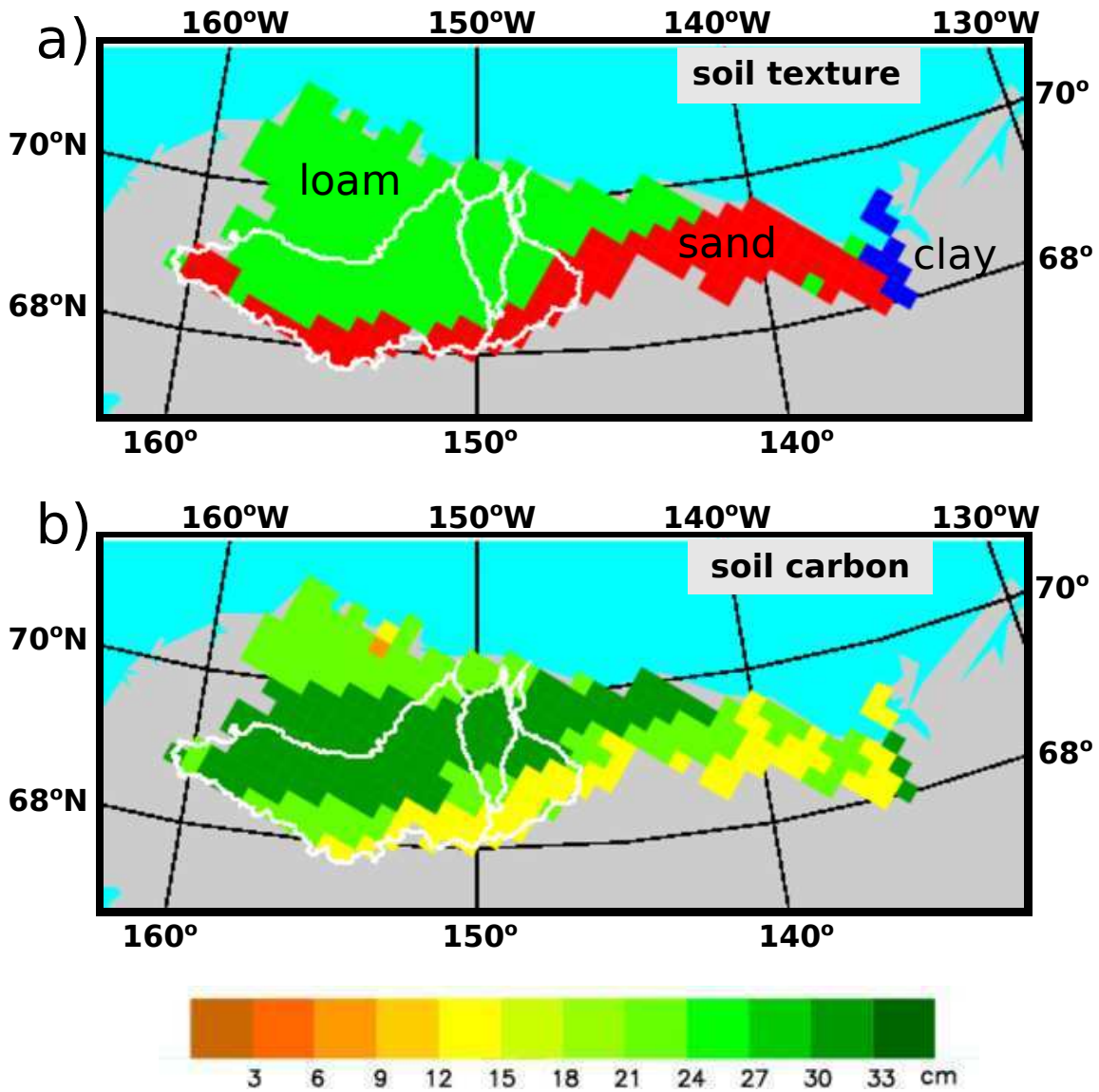


Figure S1: a) Soil texture classes and b) thickness of surface soil carbon layer used in model parameterizations. Soil textures are drawn from the UNESCO Food and Agriculture Organization’s Digital Soil Map of the World (Food and Agriculture Organization/UNESCO, 1995). Soil carbon is taken from the Northern Circumpolar Soil Carbon Database (NCSCD) (Hugelius et al., 2014). Soil carbon thickness derived from the NCSCD data and used in the PWBM includes all soil layers for which some amount of carbon is present. Primarily mineral soil exists downward over the remainder of the soil column.

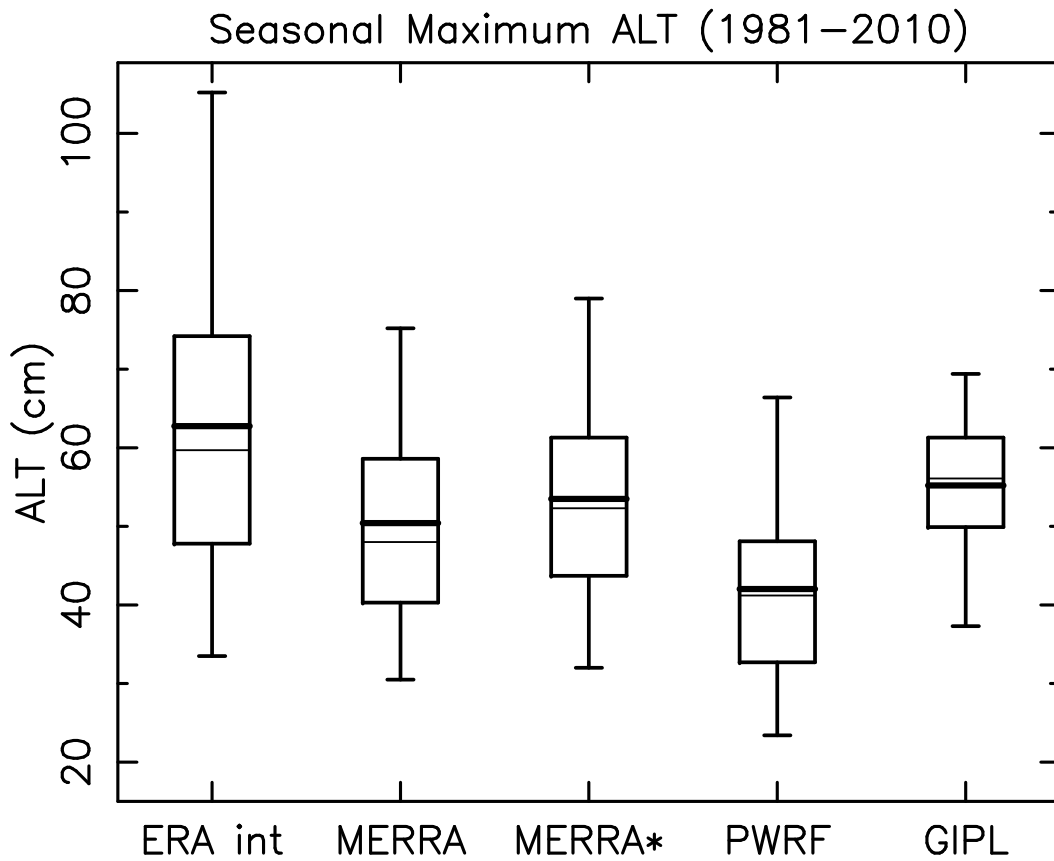


Figure S2: a) Seasonal maximum ALT (cm) as an average over the period 1981–2010 from PWBM simulations and the GIPL model. Boxplots represent the 217 (of 312) PWBM domain grid cells for which GIPL ALT data are available. Boxplots were drawn from PWBM simulation using climate forcings from ERA interim, MERRA, MERRA with precipitation adjustment (MERRA\*), and Polar WRF. Heavy line in each box is the distribution mean. Thin line is the distribution median. Boxes bracket the 25<sup>th</sup> and 75<sup>th</sup> percentiles. Whiskers show the 5<sup>th</sup> and 95<sup>th</sup> percentiles. From PWBM soil temperatures the seasonal maximum ALT is calculated as the depth to which the 0 °C penetrates each summer. Nicolsky et al. (2017) provide details on the GIPL ALT.

# Seasonal Maximum ALT

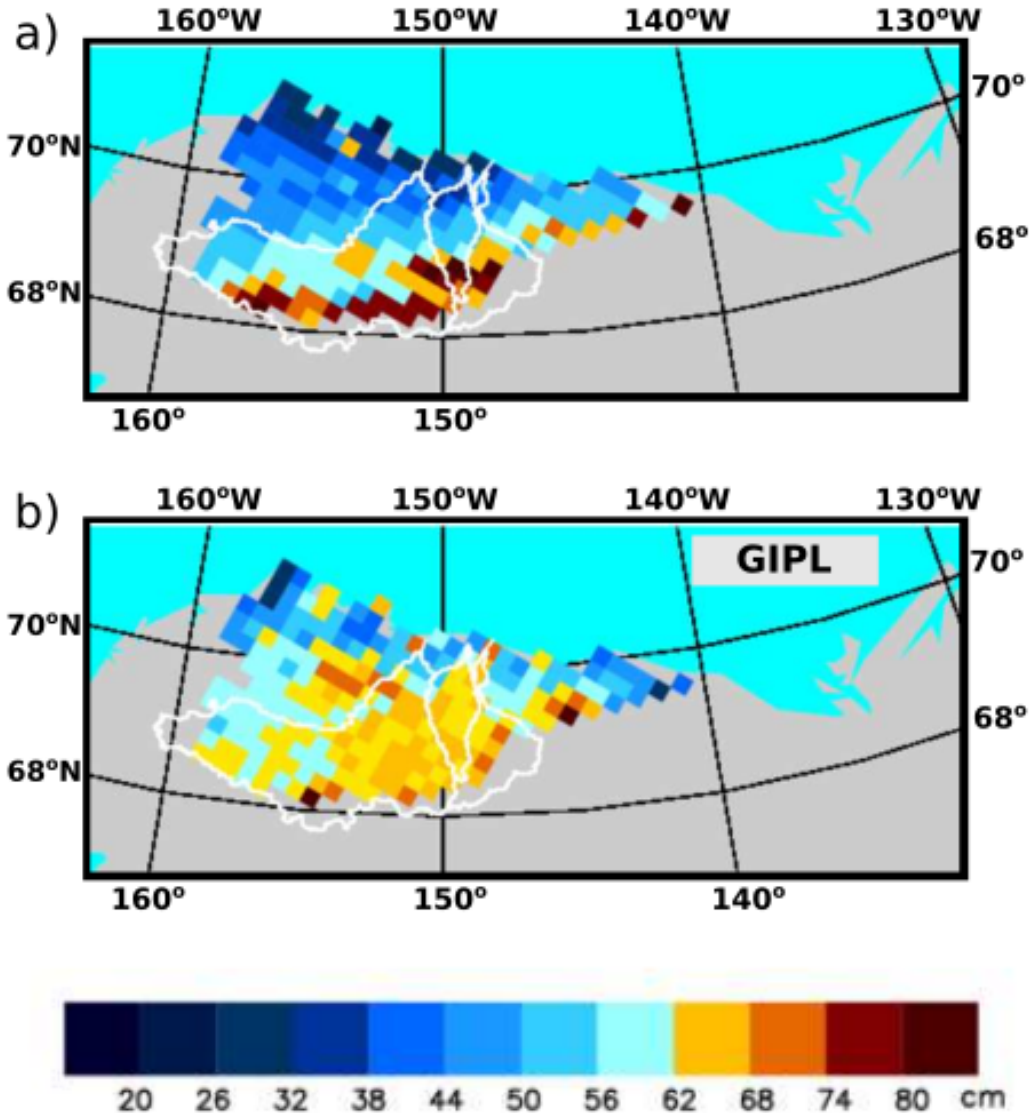


Figure S3: a) Seasonal maximum ALT (cm) as an average over the period 1981–2010 from a) PWBM with MERRA\* forcing and b) GIPL.

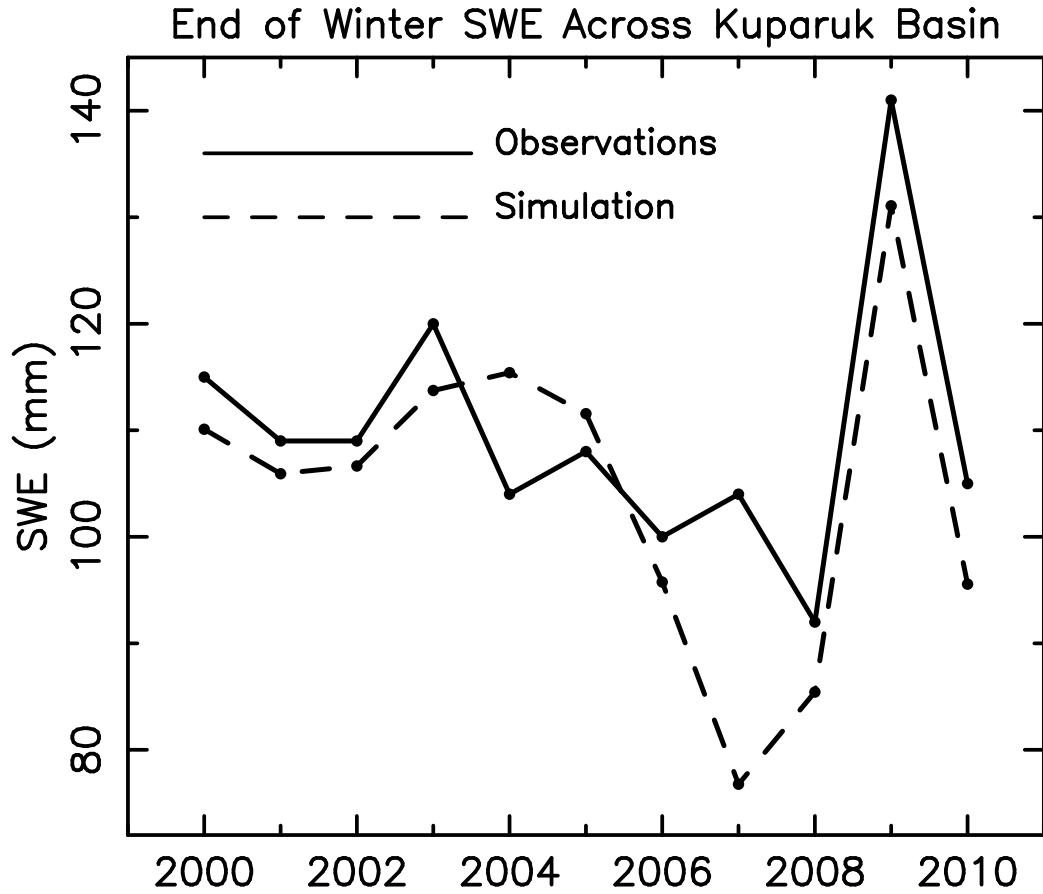


Figure S4: Observed and model simulated end of winter snow water equivalent (SWE, mm) for the Kuparuk River basin 2000–2010. Observed values represent an average of measurements made across the basin as described by Stuefer et al. (2013). Simulated end of season SWE is calculated as the average between 24 April and 7 May each year.

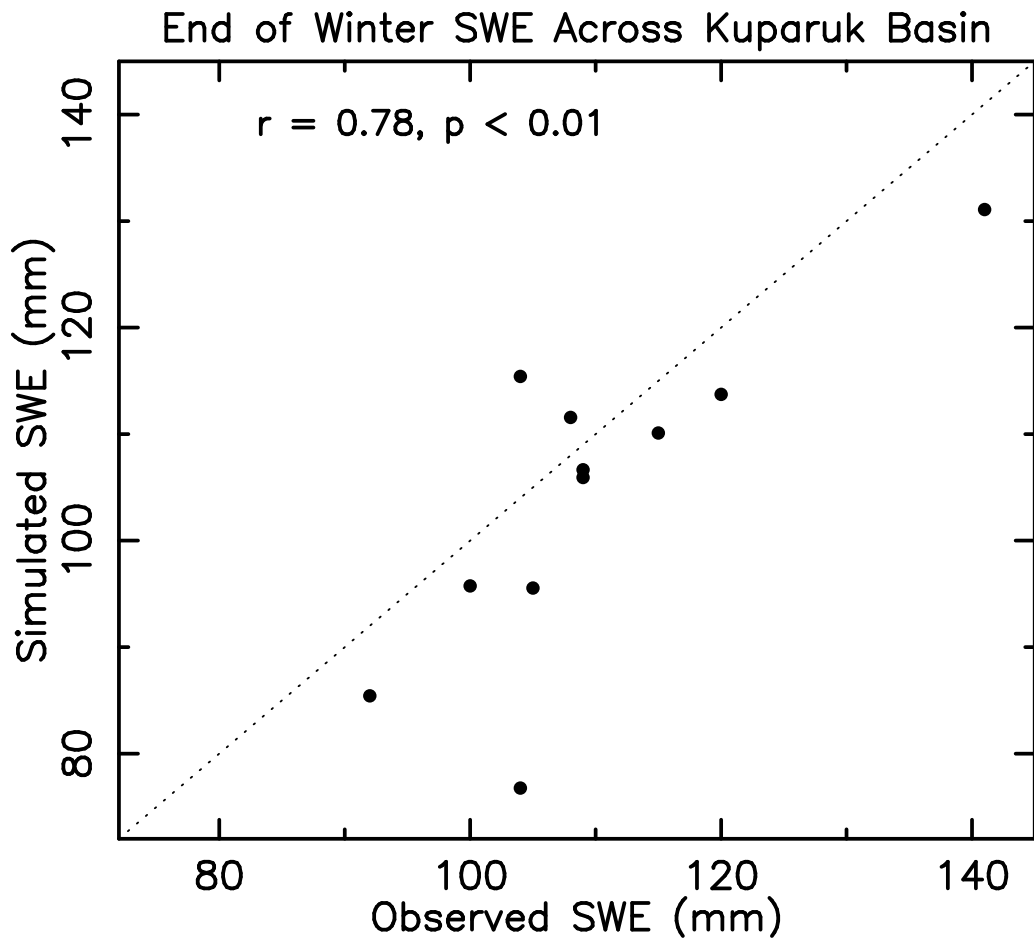


Figure S5: Observed and model simulated end of winter SWE (mm) for the Kuparuk Basin 2000–2010.

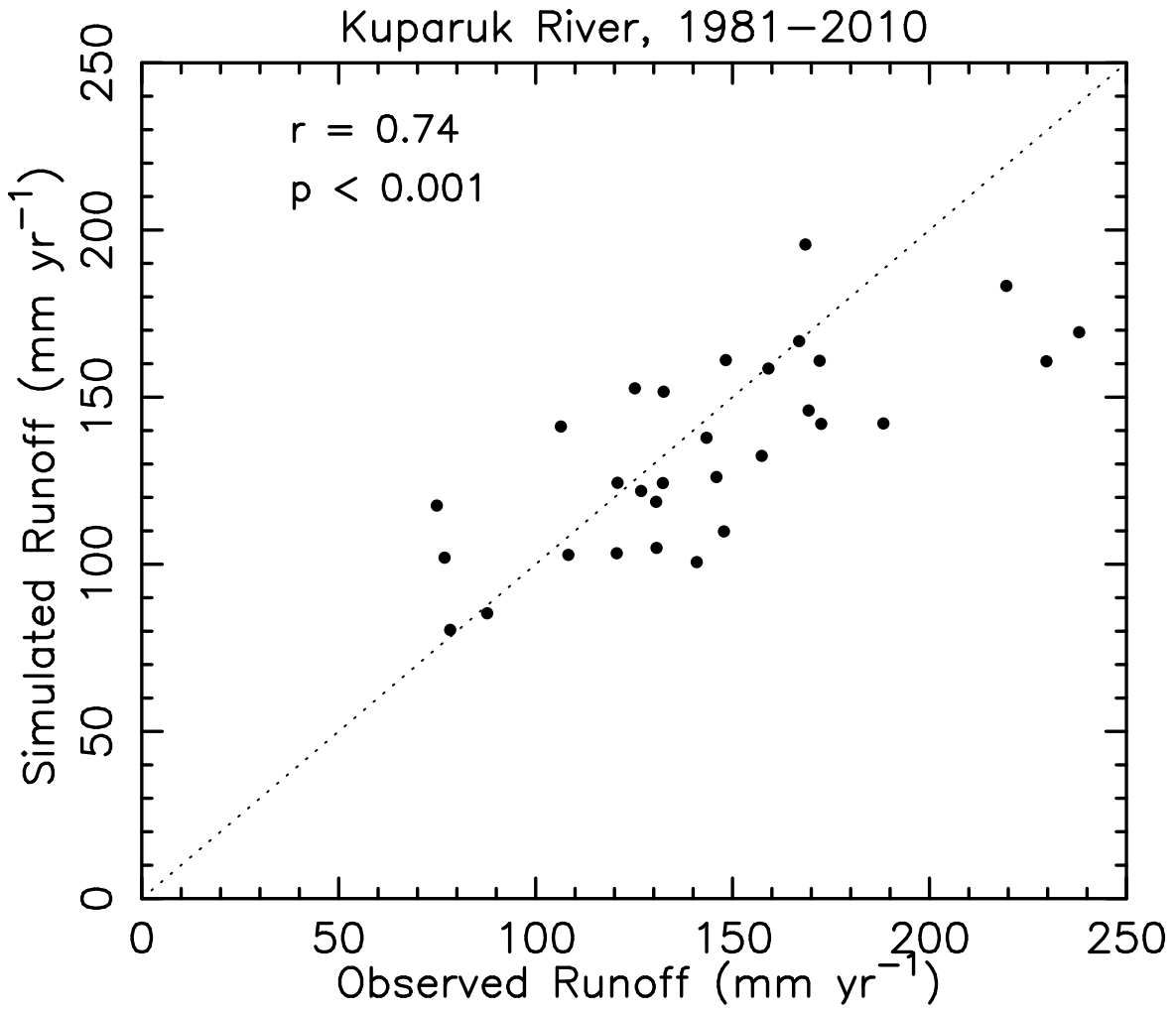


Figure S6: Simulated vs. observed annual total R ( $\text{mm yr}^{-1}$ ) for the Kuparuk basin. Correlation coefficient (LLS) is  $r = 0.73$  ( $p < 0.001$ ).



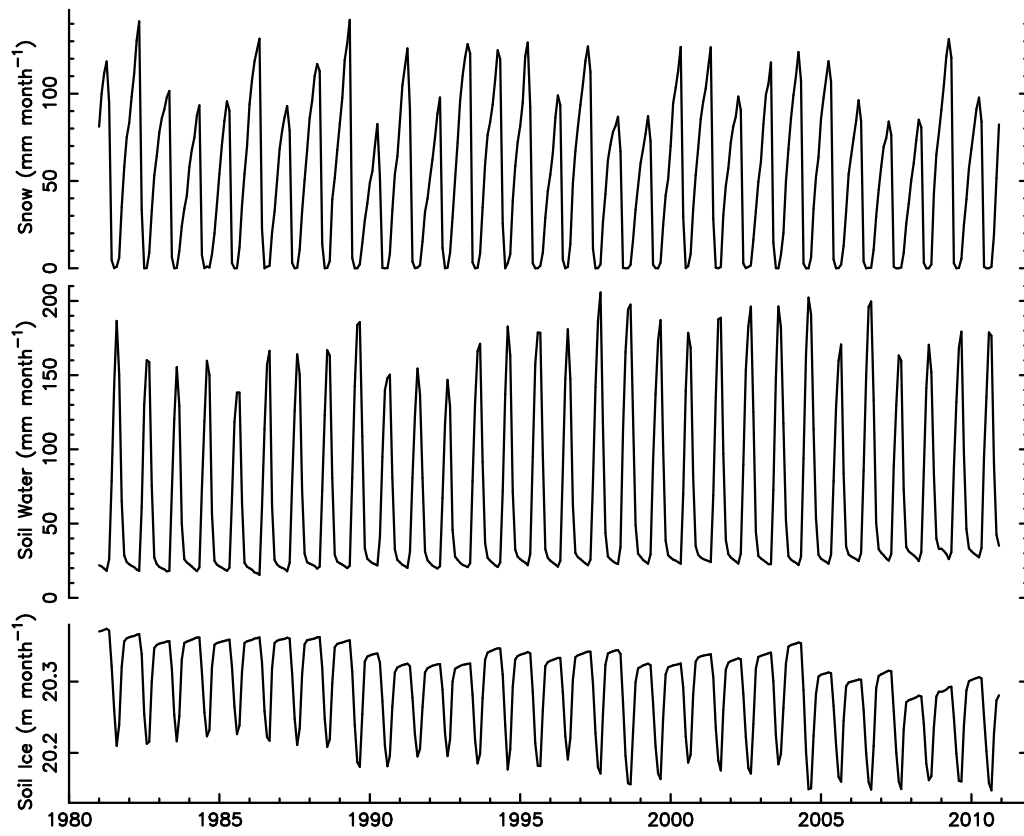


Figure S7: Monthly water storage for snow (solid and liquid portions, mm month<sup>-1</sup>), soil water (mm month<sup>-1</sup>), and soil ice (m month<sup>-1</sup>) as an average across the North Slope drainage basin. Amounts are totaled over the full 60 m model soil column

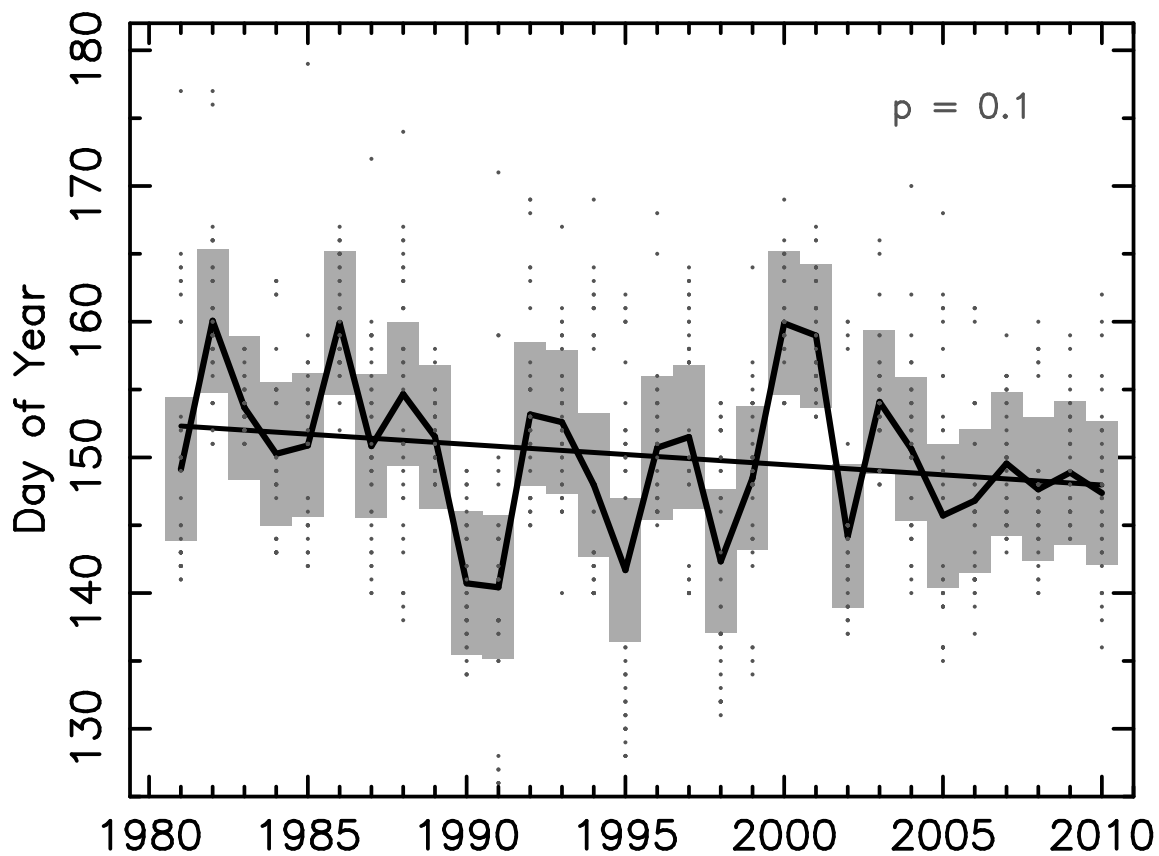


Figure S8: Date of maximum daily Q 1981–2010 for all 42 North Slope rivers. Gray bar shows the 1- $\sigma$  range around the average date (solid line). Dots indicate the date for each river. Linear least squares trend shown. Significance of linear trend (GLM) is approximately  $p = 0.1$