



# 1 Regime Shifts in Arctic Terrestrial Hydrology 2 Manifested From Impacts of Climate Warming

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## 9 **Abstract**

10 Anthropogenic warming in the Arctic is causing hydrological cycle intensification and permafrost thaw, with implications for flows of water, carbon, and energy from terrestrial biomes to coastal zones. To better understand likely impacts of the changes, we used meteorological data from atmospheric reanalysis and two global climate models to drive simulations with a hydrology model that accounts for soil freeze-thaw across the pan-Arctic drainage basin over the period 1980–2100. The simulations point to greater changes over northernmost areas of the basin, areas underlain by permafrost, and the western Arctic. An acceleration of simulated river discharge over the recent past is commensurate with trends drawn from observations and reported in other studies. Annual total runoff increases by 17 and 25% between early (2000–2019) and late century (2080–2099), while the proportion of subsurface to total runoff is projected to increase 13 and 30%, with the largest changes noted in summer and autumn, and across areas with permafrost. Most notably, runoff contributions to river discharge shift to northern parts of the Arctic basin that contain higher amounts of soil carbon. In both simulations, each season sees an increase in subsurface runoff, spring is the only season where surface runoff dominates the rise in total runoff, and summer experiences a decline in total runoff, yet an increase in the subsurface component.



29 Greater changes where permafrost exists supports the notion that increased  
30 soil thaw is shifting hydrological contributions to more subsurface flow. The  
31 manifestations of warming, hydrological cycle intensification, and permafrost  
32 thaw will impact Arctic terrestrial and coastal environments through altered  
33 river flows and the materials they convey.

## 34 1 Introduction

35 Hydrological cycle intensification and permafrost thaw are among a myriad of  
36 environmental changes reshaping the Arctic environment (Rawlins et al., 2010; Hinz-  
37 man et al., 2013; Box et al., 2019; Overland et al., 2019). Climate forcings including  
38 increasing air temperature and precipitation are key drivers of alterations to the  
39 Arctic system (Box et al., 2019). The Arctic has warmed 2.5 to 4 times faster  
40 than the global average over the past several decades (Rantanen et al., 2022; Wang  
41 et al., 2022) and experienced substantial decreases in Arctic Ocean sea ice extent  
42 and volume (Stroeve and Notz, 2018; Serreze and Meier, 2019). Warming is lead-  
43 ing to hydrologic intensification that is projected to drive higher precipitation rates  
44 (Bintanja and Selten, 2014; McCrystall et al., 2021), with concomitant rises in river  
45 discharge (Shiklomanov and Shiklomanov, 2003; Dankers and Middelkoop, 2008).  
46 Permafrost thaw has the potential to change how water is stored and moved, and  
47 to mobilize vast stores of organic carbon sequestered in soils (Frey and Smith, 2005;  
48 Koch et al., 2022; Mohammed et al., 2022), and rising river discharge (Peterson et al.,  
49 2002; Wagner et al., 2011; Feng et al., 2021) furthermore imply associated changes  
50 in exports of water, energy, carbon, and other constituents to coastal zones (Tank  
51 et al., 2016; Behnke et al., 2021; Zhang et al., 2021). In light of these alterations, it  
52 is important to better understand how climate warming, hydrological cycle intensi-  
53 fication, and permafrost thaw will impact Arctic terrestrial hydrology and, in turn,  
54 exports of freshwater and associated materials through the Arctic drainage basin and  
55 into coastal zones.

56 The seasonal storage of precipitation in the form of snow is a defining element  
57 of Arctic hydrology, contributing to abundant surface water storages and high river  
58 flows following spring melt. The presence of permafrost is another important element  
59 influencing the region's water cycle. Climate warming is intensifying Earth's water  
60 cycle, increasing precipitation, evaporation, evapotranspiration (ET), and river dis-  
61 charge globally (Huntington, 2006, 2010), and across Arctic regions (Peterson et al.,  
62 2002; Déry et al., 2009; Rawlins et al., 2010). Intensification or "acceleration" in-  
63 volves the effects of both atmospheric moisture holding capacity and moisture avail-  
64 ability. Declining sea ice is making the Arctic Ocean and its surrounding seas a



65 more available source of moisture, with locally-driven precipitation recycling great-  
66 est in winter across the Beaufort-Chukchi, Laptev, Kara, and East Siberian Seas  
67 (Ford and Frauenfeld, 2022). Increasing late summer precipitation and a shift to-  
68 ward rainfall runoff is occurring across watersheds in northwest Alaska (Arp et al.,  
69 2020; Rawlins, 2021; Arp and Whitman, 2022). Terrestrial hydrology in the Arctic  
70 is also strongly controlled by the presence of permafrost and the seasonal thawing  
71 and freezing of soils (Tananaev et al., 2020). Permafrost underlies approximately  
72 one fifth of the global land area and influences processes involving runoff, aquatic  
73 biogeochemistry (Frey and McClelland, 2009; Spencer et al., 2015; Hu et al., 2023),  
74 and land-atmosphere greenhouse gas exchanges (Christensen et al., 2004; McKenzie  
75 et al., 2021). Permafrost acts as an impermeable hydrological barrier, helping to  
76 maintain high soil suprapermfrost moisture levels while reducing soil water storage  
77 capacity and constraining subsurface flow (Woo et al., 2008; Walvoord and Kurylyk,  
78 2016). The presence or absence of permafrost and variability in precipitation pro-  
79 cesses lead to varying amounts of surface and subsurface runoff contributions to river  
80 discharge and, in turn, land-ocean exports of freshwater and associated materials.  
81 Warming is causing long-term changes in near-surface soil freeze/thaw cycles and  
82 permafrost (Anisimov and Reneva, 2006; Koven et al., 2013; Guo et al., 2018; Peng  
83 et al., 2018; Biskaborn et al., 2019), with implications for permafrost hydrology (Woo  
84 et al., 2008; Liljedahl et al., 2016; Lafrenière and Lamoureaux, 2019; Jin et al., 2022).  
85 Subsidence due to thawing soils will likely lead to more runoff, while significantly  
86 accelerating drying of tundra landscapes in a warming climate (Painter et al., 2023).  
87 Studies suggest that permafrost degradation leads to increased moisture transport  
88 from the surface to deeper soils, potentially contributing to increased river baseflows  
89 (Walvoord and Striegl, 2007) and cold season discharge (St. Jacques and Sauchyn,  
90 2009; Shiklomanov et al., 2013; Tananaev et al., 2016; Rawlins et al., 2019; Debolskiy  
91 et al., 2021; Wang et al., 2021; Liu et al., 2022). In northwest Alaska, positive trends  
92 in air temperature and precipitation are greatest in autumn which, together with  
93 permafrost thaw, is likely leading to enhanced subsurface “suprapermfrost” runoff  
94 during that time (Rawlins, 2021).

95 Climate models are essential tools for understanding how manifestations of cli-  
96 mate warming will alter the Arctic’s terrestrial hydrology and riverine land-ocean  
97 fluxes. Model projections point to future precipitation increases over the 21<sup>st</sup> century  
98 through enhanced regional evaporation and poleward moisture transport (Bintanja  
99 et al., 2020), and sea ice declines (Bintanja and Selten, 2014). Models with the  
100 strongest warming response point to decreased snowfall across the high (70–90 °N)  
101 Arctic. The precipitation increases are firmly linked to Arctic warming and sea-ice  
102 decline (Bintanja, 2018; Arp et al., 2020), and are likely to increase river discharge



103 (Peterson et al., 2002; Zhang et al., 2013). Recent coordinated research programs  
104 have produced bias-corrected climate model data for historical and future condi-  
105 tions from consistent protocol frameworks (Warszawski et al., 2014; Lange, 2021).  
106 Simulations of permafrost dynamics and associated soil freeze-thaw processes require  
107 attention to several key processes absent in many land-surface models (Alexeev et al.,  
108 2007; Nicolsky et al., 2007; Lawrence and Slater, 2008). Slater and Lawrence (2013)  
109 concluded that, in general, permafrost is not well represented by the ensemble of  
110 CMIP5 models. Examining permafrost dynamics in global models participating in  
111 the CMIP6, Burke et al. (2020) found that simulation of active-layer thickness (ALT)  
112 and other key features often fell outside the observed range, with errors attributable  
113 to shallow and poorly resolved soil profiles and structural weaknesses in snow physics  
114 and soil hydrology within some of the models.

115 In this study we use simulations with a permafrost hydrology model to evaluate  
116 how climate alterations linked to warming, primarily hydrological cycle intensifica-  
117 tion and permafrost thaw, will influence Arctic terrestrial hydrology and, in turn,  
118 land-ocean riverine freshwater and biogeochemical fluxes. We begin by examining  
119 meteorological data from climate models to understand the atmospheric forcings  
120 and their influence on surface hydrology. Model simulations are validated against  
121 select observations for sublimation, ET, ALT, and river discharge. We then examine  
122 changes over the 21<sup>st</sup> century to gain insights into how hydrological cycle intensifica-  
123 tion and permafrost thaw are impacting key elements of Arctic terrestrial hydrology  
124 controlling river exports, and test the hypothesis that within the Arctic drainage  
125 basin, changes in subsurface runoff are greatest in permafrost areas.

## 126 2 Methods

### 127 2.1 Study area and spatial grid

128 The pan-Arctic drainage basin used in this study encompasses approximately  
129 22.45 million square kilometers. It has a wide range of land cover types, from grass-  
130 lands in southern Canada and central Eurasia to boreal forests to tundra in far  
131 northern areas. This domain includes basins of rivers draining to the Arctic Ocean,  
132 Hudson Bay, and the Bering Strait, with the large Yukon River draining into the lat-  
133 ter. The region's four largest rivers—the Ob, Yenesej, Lena, and Mackenzie—flow  
134 primarily in a south-to-north direction, and account for roughly half (49%) of the  
135 pan-Arctic basin area. Model forcing data, simulations, and outputs were produced  
136 on the 25×25 km EASE-Grid (Brodzik and Knowles, 2002), with 35,693 gridcells  
137 covering the study spatial domain.



## 138 2.2 Modeling approach

139 The modeling approach leverages simulations with the Permafrost Water Balance  
140 Model (PWBM v4) to investigate the impacts of warming, hydrological cycle intensi-  
141 fication, and permafrost thaw on terrestrial hydrological fluxes within and through  
142 the pan-Arctic drainage basin. The PWBM simulates all major elements of the water  
143 cycle, including transpiration and soil and surface-water evaporation, snow storage,  
144 sublimation (Rawlins et al., 2003, 2013), runoff (Rawlins et al., 2021), and soil freeze-  
145 thaw. Past applications include assessment of causes behind record Eurasian dis-  
146 charge (Rawlins et al., 2010); estimation of surface water dynamics (Schroeder et al.,  
147 2010); analysis of present and future water budgets (Clilverd et al., 2011); quanti-  
148 fication of freshwater and dissolved organic carbon fluxes (Rawlins et al., 2021); inves-  
149 tigation of trends in those fluxes to a coastal lagoon in northwest Alaska (Rawlins,  
150 2021); and exploration of the links between surface organic soil properties and mois-  
151 ture dynamics across the Alaska North Slope (Yi et al., 2022). PWBM operates at an  
152 implicit daily time step, with meteorological forcings (air temperature, precipitation,  
153 wind speed) typically drawn from reanalysis data for regional-scale simulations or,  
154 when applied to smaller watersheds, meteorological station data. Daily simulated ET  
155 depends on atmospheric demand and surface and soil conditions. The model includes  
156 a surface water pool that is typically transient and most often occurs after snowmelt.  
157 The sum of surface and subsurface runoff from one or more soil layers within a grid-  
158 cell constitutes daily total runoff. We use the term “subsurface runoff” for the water  
159 flux often called “baseflow”, which is water seeping into the stream from groundwa-  
160 ter. Subgrid fraction of inundated area (lakes and ponds) are parameterized based  
161 on observed data (Du et al., 2016), with total runoff across each gridcell calculated  
162 as a weighted total from the inundated and non-inundated areas. The snow model  
163 simulates the effects of seasonal changes in snow density and, in turn, snow ther-  
164 mal conductivity (Liston et al., 2007; Sturm et al., 1995). Soil freeze-thaw process  
165 representations include a multi-layer soil module with algorithms for unfrozen water  
166 dynamics and phase change, as well as specification of the thermal and hydrological  
167 properties of organic soils (Sazonova and Romanovsky, 2003; Nicolsky et al., 2007).  
168 The PWBM has a 60 meter soil column, includes parameterizations for thermal and  
169 hydraulic properties of organic soils, and simulates the effect of depth hoar and wind  
170 compaction on snow density. Rawlins et al. (2013) describe the soil freeze-thaw and  
171 snow algorithms, and calibration procedures, which involve factors controlling ET,  
172 snow sublimation, and subsurface runoff that differ between forest and tundra land-  
173 scapes. In this study each transient simulation was preceded by a 50-year spinup on  
174 year 1980 to stabilize soil temperature, moisture, and soil dissolved organic carbon  
175 (DOC) pools. While parameterizations for fields such as soil texture and vegetation



176 cover are fundamental elements of land surface and hydrological model simulations,  
177 simulated runoff in Arctic regions is most sensitive to the time-varying meteorological  
178 forcings (Rawlins et al., 2003).

179 Permafrost extent is based on end of season soil temperatures. If the soil column  
180 down to the maximum 60 meter depth is frozen, beneath a thawed upper zone (i.e.  
181 active layer), the gridcell is deemed to have permafrost that year. The impact of  
182 subsidence on permafrost thaw is not accounted for in the simulations, though the  
183 effect may be relatively small (Painter et al., 2023). In models operating at continen-  
184 tal scales, estimates of permafrost extent across transition zones between continuous  
185 permafrost and the non-permafrost areas are more uncertain due to limitations re-  
186 solving spatial variations.

### 187 2.3 Meteorological forcings

188 This study focuses on numerical model simulations that were forced with grid-  
189 ded meteorological data (Table 1). We begin by examining simulations forced with  
190 reanalysis data to characterize dynamics over the recent past. Changes over the 21<sup>th</sup>  
191 century are assessed using simulations forced with meteorological data from coupled  
192 climate models, rather than the hydrology (eg. runoff) from them, as outputs from  
193 individual models can vary widely, and often imply unrealistic long-term systematic  
194 changes in water storage and level within entire basins (Bring et al., 2015).

Table 1: Simulations conducted in the study, time period for the transient simulation, and origin of forcing data. Each transient simulation was preceded by a 50 year spinup. For the climate model forcing, the 1980–2100 period includes two different experiments.

Model simulations		
Name	Period	Forcing
PWBM-W5E5	1980–2019	Bias-adjusted ECMWF Reanalysis v5 (ERA5)
PWBM-ERA5	1980–2019	ERA5 Reanalysis
PWBM-MERRA	1980–2013	Modern-Era Retrospective Analysis for Research and Applications
PWBM-IPSL	1980–2100	IPSL-CM6A-LR (Historical: 1980–2014, SSP3-7.0: 2015–2100)
PWBM-MPI	1980–2100	MPI-ESM1-2-HR (Historical: 1980–2014, SSP3-7.0: 2015–2100)



195 Simulations were made using forcings from three reanalysis datasets (W5E5,  
196 ERA5, MERRA) and two global climate models from the Coupled Model Inter-  
197 comparison Project Phase 6 (CMIP6). The WFDE5 data—WATCH Forcing Data  
198 methodology applied to ERA5 reanalysis—is bias-adjusted ERA5 data at  $0.5^\circ \times 0.5^\circ$   
199 spatial and sub-daily resolutions, generated specifically to be used as climate data in-  
200 puts for impacts studies (Cucchi et al., 2020). The WFDE5 over land is merged with  
201 ERA5 over the ocean to produce W5E5 data (Lange, 2019), compiled as part of phase  
202 3b of the Inter-Sectoral Impact Model Intercomparison Project (ISIMIP3b) (Lange,  
203 2019, 2021). We downloaded and analyzed W5E5 version 2 data for use as meteoro-  
204 logical forcings for simulations over the historical period. We use bias-adjusted data  
205 (W5E5 v2 and climate models) prepared as part of the ISIMIP framework (Cucchi  
206 et al., 2020; Lange et al., 2021). We also applied data from ERA5 and MERRA  
207 reanalysis to gauge the accuracy of the air temperature (2 m), precipitation, and  
208 wind speed forcings and for model validation. Precipitation amounts in the W5E5  
209 data are lowest among the three reanalysis datasets. To ameliorate this bias in the  
210 simulation forced with W5E5 we increased each precipitation value by 20%. The  
211 ISIMIP3b climate model forcing data are bias adjusted and statistically downscaled,  
212 and available for five CMIP6 models (GFDL-ESM4, IPSL-CM6A-LR, MPI-ESM1-  
213 2-HR, MRI-ESM2-0, UKESM1-0-LL) forced with three Shared Socioeconomic Path-  
214 ways (SSP) scenarios (SSP1-2.6, SSP3-7.0, SSP5-8.5). In our two simulations over  
215 years 1980–2100 we used data from two models (MPI-ESM1-2-HR, IPSL-CM6A-LR)  
216 forced with SSP3-7.0, which is a high emissions scenario, and suitable to investigate  
217 the response of Arctic hydrology to a strong climate forcing. These two climate  
218 models generally bracket the range of climate projections for the pan-Arctic region  
219 across the five CMIP6 models (Fig. S1). The selection of these two models—hereafter  
220 IPSL and MPI—is aimed at capturing a wide range of temperature and precipita-  
221 tion projections, but not necessarily the full range. Air temperature and precipita-  
222 tion changes expressed by the models are described in Sect. 4.1 and 4.2 respectively. In  
223 a study examining which CMIP3 models performed best at capturing meteorological  
224 quantities across parts of the Arctic, a predecessor of the MPI-ESM ranked highest  
225 (Walsh et al., 2008).

## 226 2.4 Statistical analysis

227 Our analysis of changes closely connected to Arctic rivers centers on differences  
228 between 20-yr intervals representing early (2000–2019) and late (2080–2099) cen-  
229 tury conditions. Specifically we mapped climatological averages over these periods  
230 and examined the differences for each domain gridcell. Domain-wide averages were



231 computed from all 35,693 gridcells covering the domain. The statistical significance  
232 of differences between the two periods were calculated for select quantities. Before  
233 applying the statistical significance test we used graphical analysis and the Shapiro–  
234 Wilk test (Shapiro and Wilk, 1965) to determine if the data series of interest is  
235 approximately normally distributed. The paired t test was then applied to test the  
236 null hypothesis that the mean difference between two variables is zero. Relative (per-  
237 centage) difference is calculated based on the standard formula: Relative difference  
238 (%) =  $(Z_2 - Z_1) / Z_1 \times 100$ , where  $Z_1$  and  $Z_2$  are values for early and late periods  
239 respectively. In this study we leverage these simulations forced by the two climate  
240 models to investigate the sensitivity of thermal and hydrological responses to dif-  
241 ferent climate forcings, not to provide robust quantitative projections, which would  
242 require a multi-model, multi-scenario ensemble.

### 243 3 Model Validation

244 We first compared key components of the simulated water budget–active-layer  
245 thickness, sublimation, evapotranspiration, and discharge–with different observa-  
246 tional datasets to assess the credibility of the PWBM simulations. Simulated active-  
247 layer thickness (ALT) and model-estimated permafrost extent is compared to ALT  
248 data from the National Tibetan Plateau/Third Pole Environment Data Center (TPDC)  
249 (Fig. 1a–c) and permafrost area from International Association of Permafrost (IPA)  
250 data. Simulated ALT in the model simulations spans a greater range compared with  
251 the TPDC data (Fig. 1d). However, the TPDC ALT estimates are known to have a  
252 reduced distribution range owing to the machine learning approach used (Ni et al.,  
253 2021). Permafrost extent is generally well captured, with differences from total area  
254 of continuous and discontinuous permafrost in the IPA dataset of less than 10%  
255 (Table 2).

256 We used the simulation forced with W5E5 data (PWBM-W5E5) to evaluate the  
257 magnitude of vertical fluxes of water from sublimation and ET over the recent past  
258 (Fig. 2). Overestimates in simulated sublimation are noted (domain-wide average  
259 sublimation of  $40 \text{ mm yr}^{-1}$  for GLEAM and  $57 \text{ mm yr}^{-1}$  for PWBM-W5E5), though  
260 the discrepancy is small relative to the magnitudes of annual total runoff and ET.  
261 Simulated ET ( $260 \text{ mm yr}^{-1}$ ) falls between the estimates from GLEAM ( $304 \text{ mm}$   
262  $\text{yr}^{-1}$ ) and remote sensing-based data ( $230 \text{ mm yr}^{-1}$ ), differences of 14% and 12%  
263 respectively. The model generally captures the spatial pattern in sublimation and  
264 ET.

265 We compared simulated discharge volume to a new dataset, the Remotely-sensed  
266 Arctic Discharge Reanalysis (RADR), that was generated through assimilation of ap-



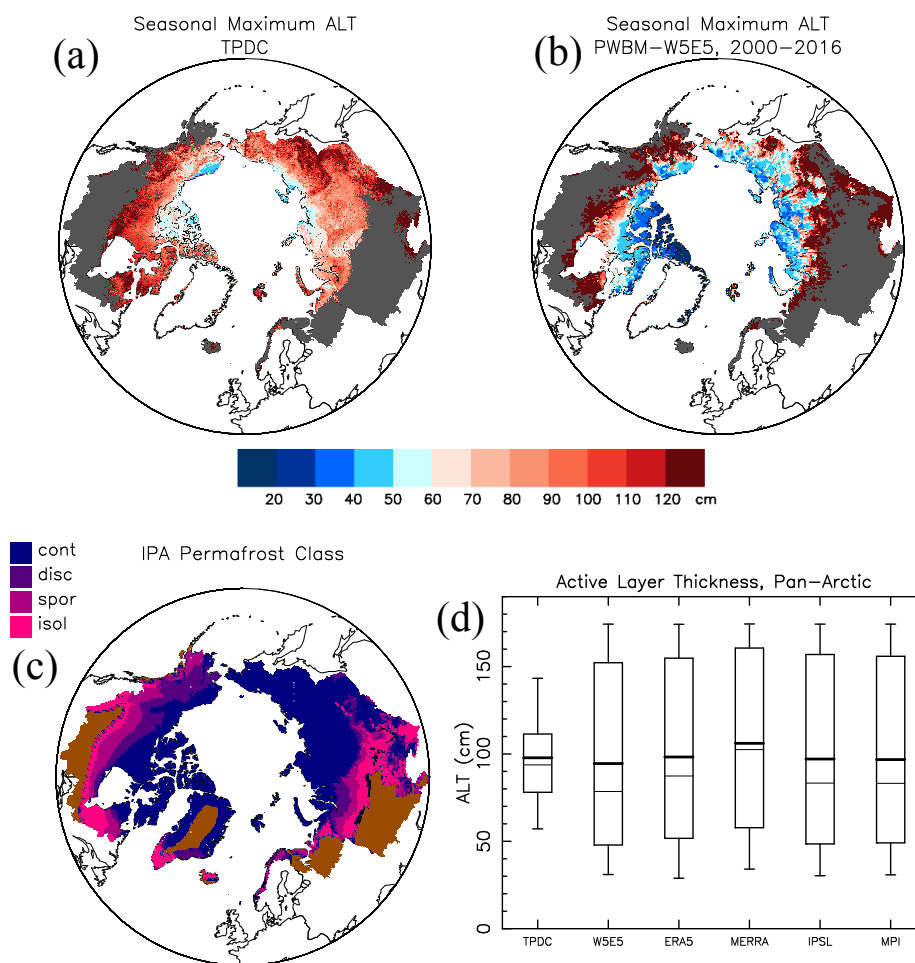


Figure 1: (a) Active-layer thickness (ALT, cm) from the TPDC database (Ran et al., 2022) for the period 2000–2016, and (b) from the PWBM simulation forced with W5E5 data over same period. Grey shading indicates non-permafrost areas. (c) Permafrost classification from International Association of Permafrost (IPA) data. (d) Distributions of seasonal maximum ALT (cm) for all grids with permafrost. ALT is the average for each year over the period 2000–2016. TPDC is used as validation for the ALT estimated by simulations forced with data from W5E5, ERA5, MERRA (2000–2013), IPSL, and MPI. Boxplot rectangles bracket the 25<sup>th</sup> and 75<sup>th</sup> percentiles. Whiskers extend to the 5<sup>th</sup> and 95<sup>th</sup> percentiles. Thick and thin horizontal lines mark the distribution mean and median respectively



Table 2: Permafrost areal extent and difference from observed extent across the study domain. Area in million km<sup>2</sup> from the International Permafrost Association (IPA) classification (Brown et al., 2001), the National Tibetan Plateau Data Center (TPDC) dataset (Ran et al., 2022), and PWBM simulations. Areas of continuous and discontinuous permafrost were added for the IPA estimate. Difference is defined based on observations from the IPA-based extent. For the simulated estimates, a gridcell is deemed to have permafrost under the standard definition of ground (model soil layer) that remains at or below 0°C for at least 2 consecutive years.

Data	Area (10 <sup>6</sup> km <sup>2</sup> )	Difference (%)
IPA	13.2	–
TPDC	12.5	–5.5
PWBM-W5E5	12.7	–4.2
PWBM-ERA5	13.1	–0.8
PWBM-MERRA	10.5	–20.4
PWBM-IPSL	12.4	–6.2
PWBM-MPI	11.8	–10.9

267 proximately 9.18 million discharge observations derived from 227 million river width  
268 measurements from Landsat images (Feng et al., 2021). Simulated discharge vol-  
269 ume is the sum total of runoff over the contributing river basin. This evaluation  
270 was performed for total discharge from the pan-Arctic drainage basin and five large  
271 Arctic rivers: the Ob, Yenesej, Lena, Mackenzie, and Yukon (Fig. S2). The model  
272 tends to overestimate discharge across western Eurasia and underestimate it across  
273 eastern Eurasia. Differences are modest for the two North American rivers. Yet the  
274 magnitude of pan-Arctic discharge is well constrained. Average freshwater export  
275 to the Arctic Ocean from the study domain over the period 1984–2018 is 5,169 km<sup>3</sup>  
276 yr<sup>-1</sup> based on RADR. Over the same period, annual total discharge is 5752, 5822,  
277 and 5811 km<sup>3</sup> yr<sup>-1</sup> in the simulations forced by W5E5, IPSL, and MPI respectively  
278 (Fig. S3), giving differences from RADR discharge of less than 13%. The simulation  
279 forced with W5E5 captures the acceleration in Arctic discharge reported in other  
280 studies (Peterson et al., 2002; Feng et al., 2021). The linear trend of 8.3 km<sup>3</sup> yr<sup>-2</sup>  
281 (0.15% yr<sup>-1</sup>) closely aligns with the acceleration (11.6 km<sup>3</sup> yr<sup>-2</sup>, 0.22% yr<sup>-1</sup>) from  
282 RADR discharge (Feng et al., 2021), and is in the upper range of estimates (3.5–  
283 10 km<sup>3</sup> yr<sup>-2</sup>) from prior studies (Shiklomanov et al., 2000; McClelland et al., 2006;  
284 Rawlins et al., 2010). For comparison, an analysis for the four largest Arctic-draining  
285 rivers (Mackenzie, Ob, Yenisei, and Lena) indicates that the combined annual dis-  
286 charge increased by 89 km<sup>3</sup> decade<sup>-1</sup> over the period 1980–2009, amounting to an  
287 approximate 14% increase over the 30-year period (Ahmed et al., 2020). Hydrologi-  
288 cal cycle intensification is connected with warming, and also manifested by increases

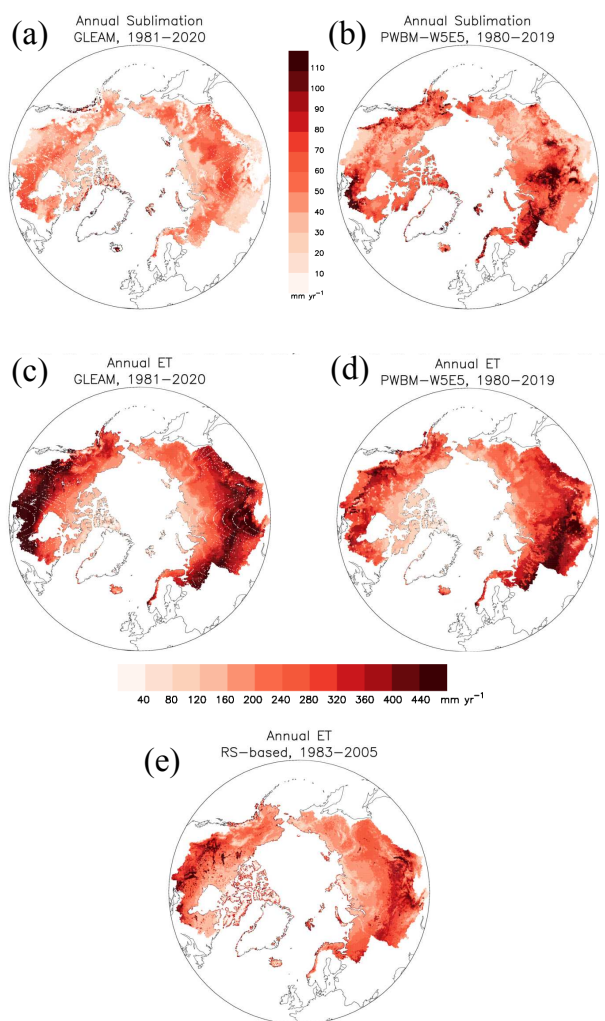


Figure 2: (a) Annual total sublimation (mm yr<sup>-1</sup>) and (c) evapotranspiration (ET, mm yr<sup>-1</sup>) from GLEAM (Miralles et al., 2011; Martens et al., 2017) and PWBM-W5E5 (b,d). Bottom panel (e) shows ET from a dataset derived from remote sensing data (Zhang et al., 2009).

289 in vertical fluxes of precipitation and ET. The differences of less than 15% between  
290 model simulated ET and discharge, and the estimates from the validation datasets,



291 suggests that the water budget components are sufficiently well constrained to enable  
292 evaluation of the impact of climate warming on runoff and river discharge in Arctic  
293 rivers. In general, the comparisons with observations support the model's ability to  
294 reliably simulate key hydrological variables of interest.

## 295 4 Alterations connected to hydrological cycle in- 296 tensification and permafrost thaw

### 297 4.1 Air temperature

298 In this analysis we use the simulations forced by the two climate models to bracket  
299 changes likely to occur this century, focusing primarily on twenty-year periods repre-  
300 senting early (2000–2019) and late (2080–2099) century conditions. The IPSL model  
301 projects stronger warming compared to MPI, with warming between early and late  
302 century of 7.2 °C (domain-wide mean value) and 6.2 °C, respectively (Table 3). Both  
303 show the strongest warming over the highest latitudes of the pan-Arctic basin, with  
304 warming of over 10 °C across far northern Canada projected by IPSL. More modest  
305 warming of 3–4 °C is noted over southwestern Canada and central Eurasia in the  
MPI data.

Table 3: Climatological averages for early (2000–2019) and late (2080–2099) century periods from the simulations forced with IPSL and MPI meteorological data. <sup>a</sup>Relative (percentage) difference shown for each except air temperature, which is shown in degrees C. Differences are statistically significant for all quantities listed based on the paired T test (Sect. 2.4).

Variable	PWBM-IPSL			PWBM-MPI		
	early	late	% diff <sup>a</sup>	early	late	% diff <sup>a</sup>
air temp (C)	−5.3	1.9	7.2	−5.3	−0.9	6.2
precipitation (mm yr <sup>−1</sup> )	578	697	21	573	643	12
net precipitation (mm yr <sup>−1</sup> )	258	315	22	259	300	16
rainfall (mm yr <sup>−1</sup> )	334	437	31	354	413	17
snowfall (mm yr <sup>−1</sup> )	244	260	7	219	230	5
rainfall fraction (%)	56	62	11	43	63	47
runoff (mm yr <sup>−1</sup> )	264	329	25	266	310	17
F <sub>sub</sub> (%)	27	35	30	30	34	13

306



307 In the results that follow, unless otherwise noted, statements reporting two statis-  
308 tics will be written in order for PWBM-MPI and PWBM-IPSL respectively. In nearly  
309 every instance, changes are greater with the latter simulation due to the influence of  
310 forcing from the more strongly warming (and wetter) IPSL climate model.

## 311 4.2 Precipitation

312 Rainfall rates have also been increasing across much of the pan-Arctic. Rainfall  
313 will continue to increase this century, particularly along favored storm track regions  
314 over northwestern Eurasia and western Alaska (Fig. 3a, S4a), where the majority of  
315 water vapor transport into the Arctic occurs (Nash et al., 2018). Climatological aver-  
316 age rainfall (domain average) is higher by late century, with relative differences of 17  
317 and 31% for the MPI and IPSL models, respectively (Table 3). Snowfall is projected  
318 to increase over a smaller geographic extent, mainly the higher latitudes and across  
319 the colder parts of eastern Eurasia, and decrease over most of the pan-Arctic, most  
320 prominently western Eurasia and southern Canada (Fig. 3b, S4b). The domain-wide  
321 change averages 5 and 7%. The sizable rainfall increases drive the projected rise in  
322 the fraction of rainfall to total precipitation (Fig. 3c, S4c) averaging 11 and 47%  
323 for the two simulations. Net precipitation—the difference between precipitation and  
324 the sum of evapotranspiration and snow sublimation—is projected to increase across  
325 most (> 75%) of the pan-Arctic basin. Decreases will occur across southern Canada  
326 and Eurasia. For areas with and without permafrost, mean changes (increases) are 31  
327 and 42%, and 5 and 6%, respectively. The simulations thus reveal bigger impacts—a  
328 net wetting—over permafrost regions, and a strong latitudinal south-north gradi-  
329 ent in future precipitation changes that will influence river discharge quantity and  
330 quality.

## 331 4.3 Permafrost extent and active layer thickness

332 Research studies have documented hydrological cycle intensification and per-  
333 mafrost thaw across the terrestrial Arctic. To better understand changes in per-  
334 mafrost hydrology attributable to warming and increasing soil thaw we calculated  
335 ALT averages from the two climate-model-forced simulations (Fig. 4, S5). For  
336 PWBM-IPSL, permafrost area decreases by 7.8 million km<sup>2</sup> (12.3 to 4.5 million km<sup>2</sup>)  
337 from the early to late century periods, a decline of 63% of present day permafrost  
338 area. For PWBM-MPI, some 4.9 million km<sup>2</sup> or 42% of present area loses permafrost  
339 (11.7 to 6.8 million km<sup>2</sup>). Predictions of soil temperature from CMIP5 models point  
340 to permafrost fractional losses by end of century of 15% to 87% for RCP4.5, and 30%

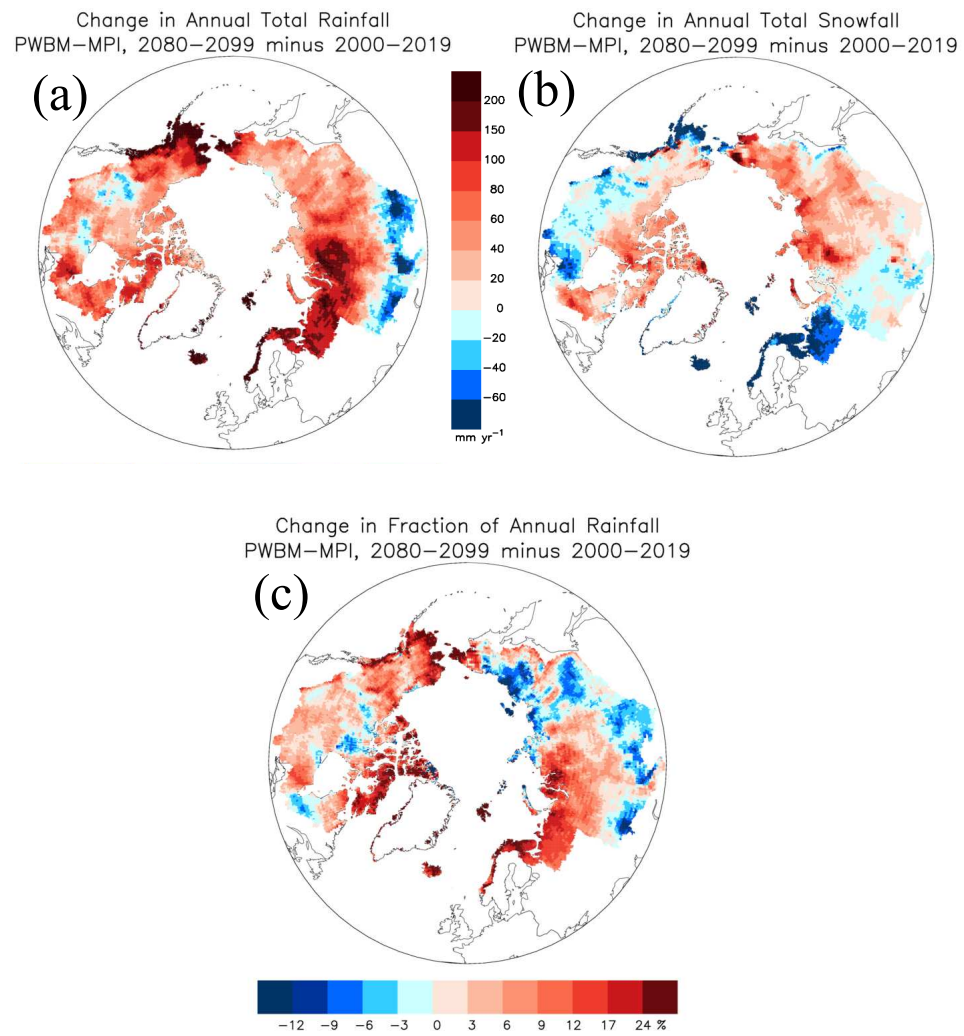


Figure 3: Change in (a) annual rainfall ( $\text{mm yr}^{-1}$ ), (b) snowfall ( $\text{mm yr}^{-1}$ ), and (c) the fraction of rainfall to total precipitation from PWBM-MPI simulation.

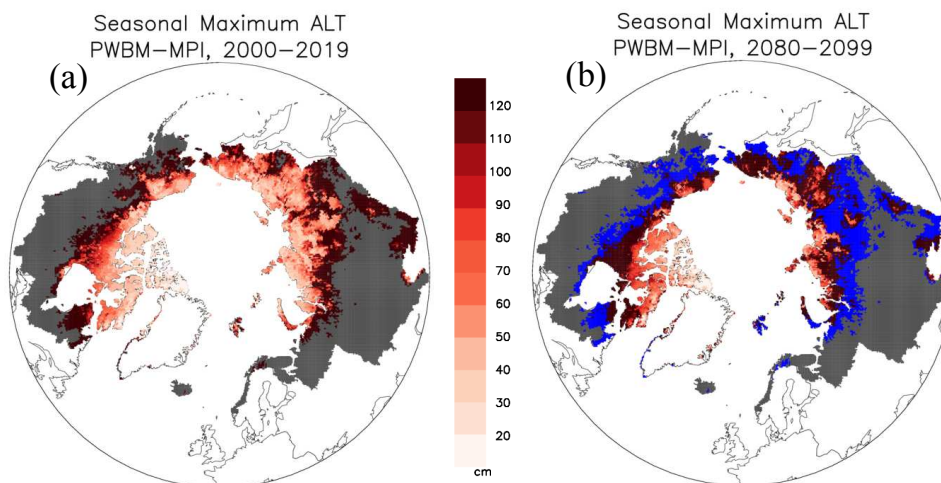


Figure 4: Simulated active-layer thickness (ALT, cm) for (a) early (2000–2019) and (b) late century (2080–2099) periods from PWBM-MPI. Blue shading highlights areas that are no longer characterized as permafrost in the future period. Gray areas are non-permafrost areas of the Arctic basin.

341 to 99% for RCP8.5 (Koven et al., 2013). Across areas that maintain permafrost, the  
342 ALT increases between the two periods average 56 and 91 cm. For comparison, esti-  
343 mates over permafrost areas obtained from an air temperature-based thawing index  
344 applied to 16 CMIP5 models (2006–2100) forced under RCP8.5 averaged a similar  
345 6.5 cm decade<sup>-1</sup>.

#### 346 4.4 Runoff and river discharge

347 Annual runoff within the pan-Arctic basin is typically highest across eastern  
348 Canada, western Eurasia, and coastal regions of western Canada and western Alaska.  
349 Runoff changes between the early and late century periods were calculated here to  
350 assess future alterations to river discharge (Fig. 5a, S6a). In Eurasia the change in  
351 annual total runoff, as a percent of the early period, is greater over northeast parts  
352 of the continent. Across North America the increases are also greater in the colder  
353 northern parts of the Canadian archipelago and over northern Alaska. Averaged  
354 across all gridcells, annual runoff increases by 19% (45 mm yr<sup>-1</sup>) and 31% (65 mm  
355 yr<sup>-1</sup>) from PWBM-MPI and PWBM-IPSL, respectively. Not surprisingly, the spatial  
356 pattern in runoff change closely aligns with the pattern in net precipitation. There  
357 is also a significant difference in the mean change in annual runoff between gridcells  
358 with permafrost (67 and 99 mm yr<sup>-1</sup> increase) and those without permafrost (21 and  
359 25 mm yr<sup>-1</sup>). This divergence is driven by changes in net precipitation (64 and 89  
360 mm yr<sup>-1</sup> vs. 18 and 19 mm yr<sup>-1</sup>), as well as differing influences from deepening ALT

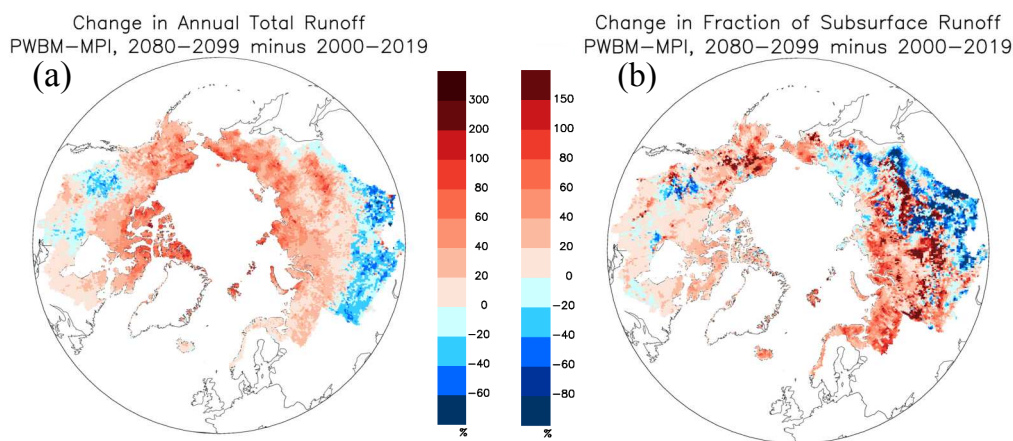


Figure 5: Change in (a) annual total runoff (%) and (b) fraction of subsurface to total runoff ( $F_{sub}$ , %) from the simulations.

361 and longer thawed periods in areas with and without permafrost. Across permafrost  
362 areas, the difference between net precipitation and runoff—in a water budget, an  
363 approximation for change in storage—is 3–10 mm yr<sup>-1</sup>, a small amount relative to  
364 the runoff increase. Over the early century period, river discharge volume is 5839,  
365 5955, 5917 km<sup>3</sup> yr<sup>-1</sup> for the PWBM-W5E5, PWBM-MPI, PWBM-IPSL simulations  
366 respectively (Fig. S3). By late century, discharge volume increases to 6955 and  
367 7374 km<sup>3</sup> yr<sup>-1</sup>, relative increases of 17 and 25% for PWBM-MPI, and PWBM-IPSL  
368 respectively (runoff equivalents in Table 3). The trend is statistically significant ( $p$   
369 < 0.01) for both time series.

370 A transition from runoff dominated by surface water contributions toward in-  
371 creasing amounts of subsurface flow is expected as the climate warms (Frey and  
372 McClelland, 2009). Compared to change in total runoff, the change in the fraction  
373 of subsurface to total runoff ( $F_{sub}$ ) is more spatially variable across the pan-Arctic  
374 (Fig. 5b, S6b). During the early century period,  $F_{sub}$  averages 30% and 27% in the  
375 PWBM-MPI and PWBM-IPSL simulations respectively (Fig. 6). The fractions in-  
376 crease to 34% and 35% by end of century, giving relative (percent) increase in domain  
377 mean  $F_{sub}$  of 13 and 30% for PWBM-MPI and PWBM-IPSL respectively. Based on  
378 the modest warming PWBM-MPI run, approximately 72% of permafrost areas will  
379 have higher subsurface runoff fractions by end of century. This spatial extent in-



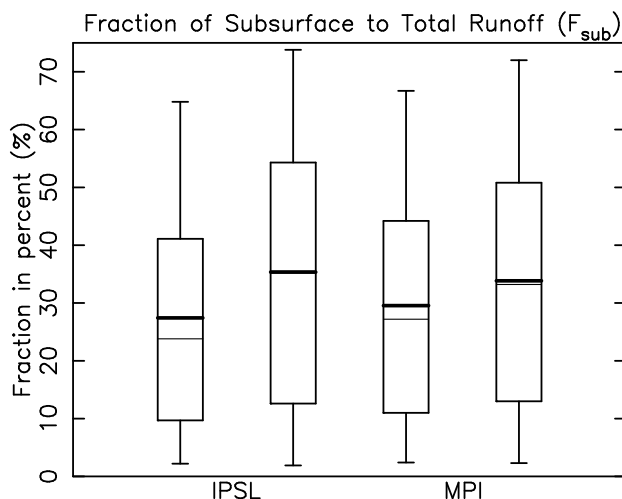


Figure 6: Fraction of subsurface to total runoff ( $F_{sub}$ ) for early and late century periods for all pan-Arctic grids from PWBM-IPSL and PWBM-MPI simulations.

380 creases to 88% of the permafrost region under the more aggressive warming depicted  
381 under PWBM-IPSL (Fig. S6b). The shift in  $F_{sub}$  is larger in permafrost areas, with  
382 significant differences in spatial mean  $F_{sub}$  in areas with and without permafrost  
383 (relative differences 15.7 and 13.5% respectively for PWBM-MPI; 31.1 and 24.4%  
384 for PWBM-IPSL). The PWBM-MPI simulation reveals a significant relationship ( $p$   
385  $< 0.01$ ) between change in ALT and  $F_{sub}$ , with a 6.4% increase in  $F_{sub}$  per 0.1 m  
386 increase in ALT. While the positive correlation does not exist under PWBM-IPSL,  
387 the more pervasive growth in  $F_{sub}$  in PWBM-MPI suggests a connection between soil  
388 thaw and increasing contributions from subsurface runoff to river discharge during  
389 this century, particularly in regions underlain by permafrost.

390 The runoff changes in both simulations exhibit a significant positive relationship  
391 with latitude (Fig. 7a, S7a). The linear fit suggests an additional 2.9 and 4.2% runoff  
392 (PWBM-MPI and PWBM-IPSL) for each degree northward in latitude. Under this  
393 pattern river discharge shifts over time to being sourced more from the northerly  
394 parts of the four largest river basins (Ob, Yenesej, Lena, Mackenzie; Fig. 8a, S8a,  
395 Table 4). Decreases are projected for the southerly half of the Ob, Yenesej, and  
396 Mackenzie Rivers. For the Ob basin, less runoff across the southern half of the river  
397 basin will be offset by higher flow in the north, so that annual total discharge exported  
398 at the coast is relatively unchanged. The Yenesej shows a similar pattern, with

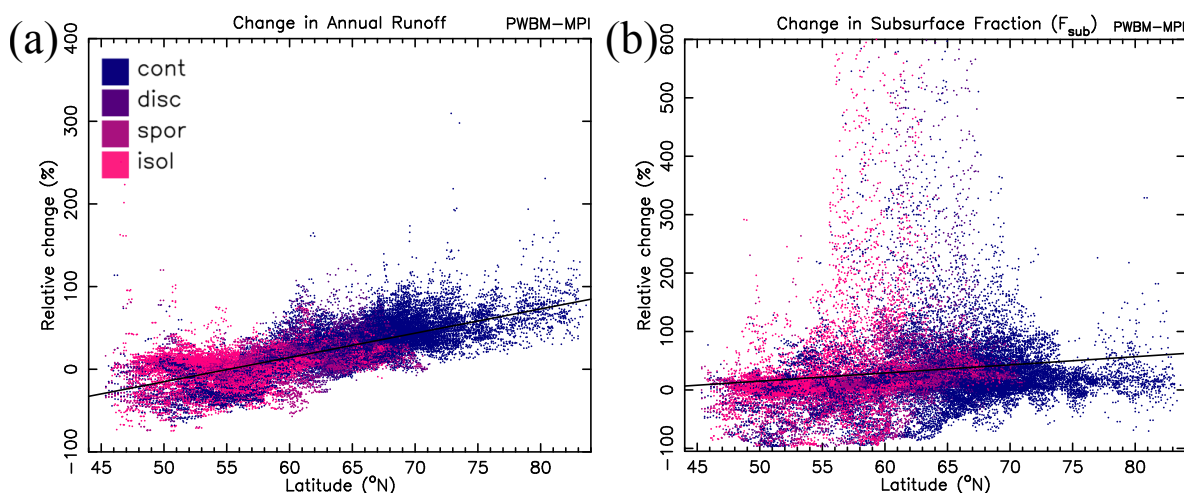


Figure 7: Change in (a) annual total runoff (%) and (b)  $F_{sub}$  with gridcell latitude from PWBM-MPI simulation for all pan-Arctic domain gridcells. Colors indicate permafrost classification for the cell from IPA (Figure 1c).

399 accumulated discharge at the coast higher by late century. The Lena and Mackenzie  
400 Rivers will receive substantial additional discharge from their northern areas, with  
401 the Lena projected to export  $66$  and  $128 \text{ km}^3 \text{ yr}^{-1}$  (16 and 31%) more freshwater  
402 discharge by late century. The sharp increase in export from the Yenesei and Lena  
403 arising from their northern watersheds is driven primarily by higher snowfall rates  
404 (Fig. 3b, S4b). Averaged across the four, the downstream half of the rivers will  
405 receive approximately 20–30% more accumulated discharge from the northern half  
406 of their contributing area. A south-north gradient also exists in soil carbon storage  
407 in these basins, with the highest amounts in the far north (Fig. 8b, S8b). Subsurface  
408 runoff increases are also greater to the north (Fig. 7b, S7b), though the scatter is  
409 substantial compared to the change in annual total runoff.

410 Runoff is projected to increase during most months in both simulations (Fig. 9, S9),  
411 with monthly changes remarkably similar between the two runs. Averaged over sea-  
412 sons, runoff increases (depth in mm) are greatest in spring (MAM). The increase in  
413 spring, particularly during May, is attributable to additional snowmelt runoff and a  
414 shift to earlier snowpack melting. As a consequence, less snowmelt and runoff occur in  
415 June. We applied a simple river flow accumulation and linear routing model (Rawlins  
416 et al., 2019) to estimate the timing shift in discharge export at the coast. Averaged  
417 across the six largest rivers (Ob, Yenesei, Lena, Mackenzie, Yukon, Kolyma), peak

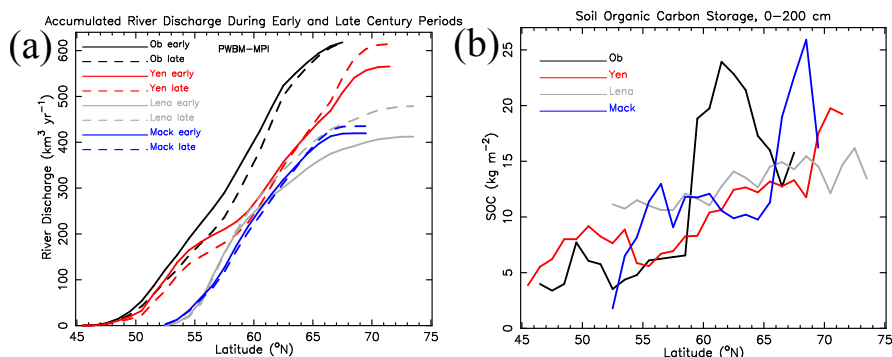


Figure 8: (a) Accumulated annual total river discharge (km<sup>3</sup> yr<sup>-1</sup>) for the Ob, Yeneseey, Lena, and Mackenzie Rivers for 1° latitude bands as averages over early (solid line) and late (dashed) century periods from PWB-MPI. (b) Soil carbon storage (kg m<sup>-2</sup>) in soil 0–200 cm zone from the Northern Circumpolar Soil Carbon Database (Hugelius et al., 2013).

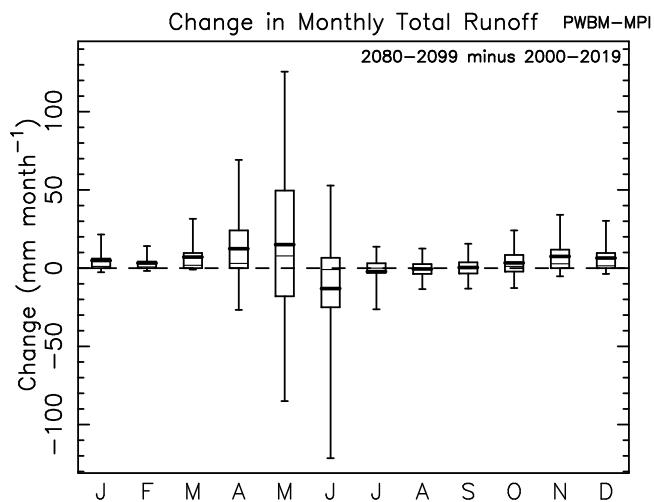


Figure 9: Distribution in change in monthly total runoff (mm month<sup>-1</sup>) between early and late century periods for all pan-Arctic gridcells from PWB-MPI.



Table 4: Relative (percentage) change in accumulated river discharge for the upstream (southern) half and downstream (northern) half of each of the four largest Arctic rivers. Averages are calculated from the totals shown in Fig.s 8, S6. Total row represents the average from the four.

River	PWBM-IPSL		PWBM-MPI	
	up (%)	down (%)	up (%)	down (%)
Ob	-9.8	7.4	-19.4	13.6
Yenesey	-1.5	27.9	-14.2	22.2
Lena	26.4	43.8	12.5	25.9
Mackenzie	-0.2	35.3	-5.3	17.3
Total	3.7	28.6	-6.6	19.7

418 daily discharge at each coastal outlet shifts earlier by end of century by approxi-  
419 mately 11 days in both simulations (DOY 180 to 169 in PWBM-IPSL and DOY 176  
420 to 165 in PWBM-MPI). Runoff is largely unchanged in July, August and September,  
421 and the changes are not statistically significant in June and July. Seasonally, the  
422 relative change (percentage change) is greatest in winter, with runoff by late century  
423 a factor of 5–10 greater compared to the early century period averages. Significant  
424 percentage increases are noted in autumn and spring as well. Interestingly, snow  
425 storage (snow water equivalent, SWE) increases in both simulations are significant  
426 in February, March, and April only. Notably, no increase in SWE is projected during  
427 autumn.

428 The intensifying hydrological cycle and thawing permafrost will manifest in chang-  
429 ing amounts of surface and subsurface runoff contributions to river discharge (Fig. 10).  
430 The shifts vary strongly with season, and spatially across the terrestrial Arctic, with  
431 remarkably similar change magnitudes in the two simulations, due largely to similar-  
432 ities in patterns in net precipitation and its change this century. At the pan-Arctic  
433 scale, modest increases are projected in both surface and subsurface runoff for the  
434 annual total and in winter, spring, and autumn. The acceleration during winter and  
435 autumn will come predominantly from additional subsurface runoff. Spring increases  
436 are mainly attributable to increased surface runoff. Runoff is projected to decrease  
437 slightly in summer due to less surface runoff, despite a small increase in subsur-  
438 face runoff. The autumn change is particularly noteworthy over northern Alaska.  
439 Also there, summer shows a strong shift from surface to subsurface runoff. Runoff  
440 decreases are projected to occur in most seasons over southwest Canada, owing to  
441 relatively large precipitation declines (Fig. 3, S4).

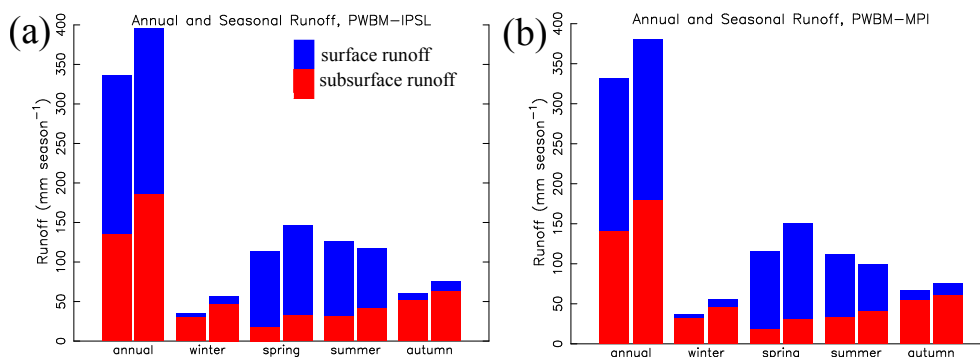


Figure 10: Annual and seasonal total runoff for the early (left bar) and late century (right bar) periods, expressed as surface (blue) and subsurface (red) amounts for (a) PWBM-IPSL and (b) PWBM-MPI simulations.

## 442 5 Discussion

443 The Arctic basin is drained by several rivers that receive runoff contributions  
444 over great distances, from grasslands and forests in the south to tundra in the north.  
445 Surface runoff has typically been a substantial component of river discharge, with  
446 subsurface flows characterizing low flows in summer and early fall. These character-  
447 istic patterns and dynamics are shifting due to influences from warming, primarily  
448 hydrological cycle intensification and permafrost thaw. The shifts are altering the  
449 water cycle from processes manifesting both horizontally, via primarily atmospheric  
450 effects, and vertically, from soil thaw, and seasonally, through a combination of both  
451 impacts. Recent research suggests that a warming Arctic will experience changes in  
452 moisture sources that will influence freshwater exports from rivers. The two coupled  
453 climate models from which outputs were used in this study capture substantial  
454 precipitation increases in regions adjacent to the Arctic Ocean. This is a robust  
455 feature of climate models that is linked to a more open Arctic Ocean later this cen-  
456 tury (Barnhart et al., 2016; McCrystall et al., 2021). River basins near the western  
457 Arctic Ocean, particularly far northeast Eurasia, northwest Canada, and northern  
458 Alaska, will experience relatively large increases in river discharge, driven partly by  
459 higher snowfall rates and spring SWE amounts. These are cold areas that will warm  
460 significantly and, in turn, increasingly be fed by additional moisture, including from  
461 more frequent atmospheric rivers (Zhang et al., 2023). In contrast, southern parts



462 of the pan-Arctic basin are projected to experience a decline in net precipitation  
463 and runoff contributions to rivers. In general, rivers in central Eurasia and southern  
464 Canada will receive less runoff, particularly during summer. Our results suggest that  
465 nearly 90% of the increase in river discharge from permafrost regions will arise from  
466 an increase in net precipitation (Cubasch et al., 2001), rather than a “de-watering”  
467 of permafrost from thawing of soil ice. This connection to net precipitation is con-  
468 sistent with attribution studies for the river discharge trends observed during the  
469 recent past (McClelland et al., 2004, 2006). Our results point to significant shifts  
470 in sources of freshwater entering Arctic rivers, with less runoff to river networks in  
471 the south and more in the north. The headwaters of the large Arctic rivers like the  
472 Lena, Ob, Yenisey, Mackenzie, originate well south of what is typically considered  
473 Arctic lands. The simulations suggest that by end of century, some 20–30% more  
474 freshwater discharge will enter, accumulate in, and be export from the northern half  
475 of the four large rivers.

476 In addition to geographic shifts involving atmospheric influences, ongoing soil  
477 thaw and permafrost losses will also influence runoff and materials contributions to  
478 rivers. Our results support a growing body of evidence that deepening active layers  
479 and losses in permafrost extent will increase subsurface runoff contributions to rivers.  
480 Permafrost extent declines by 42 and 63% (PWBM-MPI and PWBM-IPSL respec-  
481 tively) between early (2000–2019) and late (2080–2099) century periods, indicative  
482 of recent and future permafrost degradation. Recent observations in northern Alaska  
483 suggest that increased precipitation and deepening ALT play increasingly important  
484 roles in sustaining low flows and enhancing subsurface hydrologic processes (Arp  
485 et al., 2020; Cooper et al., 2023). Projected changes in subsurface runoff are more  
486 spatially variable compared to total runoff, though a similar south-north gradient  
487 exists. Increased subsurface runoff can lead to decreases in summer stream temper-  
488 atures in headwater catchments (Sjöberg et al., 2021). Pronounced seasonal shifts  
489 in runoff contributions will also occur. Increased runoff in late spring will likely be  
490 driven by higher snow storage and earlier melt that will shift peak spring freshet  
491 runoff earlier by approximately 11 days this century. Increased autumn discharge in  
492 the simulations is not attributable to higher SWE, forced instead by thawing per-  
493 mafrost that is lengthening the period when flow occurs, and creating deeper active  
494 layers that store and release water later in the season. More runoff during November  
495 and December, an approximate 5-fold increase in the modest warming simulation,  
496 highlights the physical connection between warming, permafrost degradation, and  
497 increasing subsurface flows to streams and rivers (St. Jacques and Sauchyn, 2009;  
498 Rawlins et al., 2019). The relatively large changes in November–April runoff de-  
499 scribed here are congruent with a recent study that documented a 10% per decade



500 increase in cold season discharge from nine rivers in Alaska with long data records  
501 (Blaskey et al., 2023). Warming, prominent in this region during autumn and early  
502 winter, can promote increased soil water storage, delaying the release of water into  
503 the streams, and thus contribute to increases in winter flow (Streletskiy et al., 2015).  
504 Results of this study support the hypothesis that across the Arctic basin subsurface  
505 runoff increases will be greatest in permafrost areas.

506 **Taken together**, the spatial shifts suggest alterations in materials exported to  
507 coastal waters. Warming and higher rainfall rates will enhance thaw and increase  
508 coastal erosion. Higher runoff rates will drive additional subsurface contributions  
509 of freshwater and DOC to coastal seas and lagoons (Connolly et al., 2020). More  
510 cold season river discharge has the potential to affect sea ice dynamics and other  
511 near-shore processes involving quantities such as salinity and biogeochemistry. The  
512 impacts extend to water quality and materials exports by rivers. For example, DOC  
513 input to the Arctic Ocean has a very high temporal and geographical variability  
514 with a strong bias towards the large Eurasian Rivers and the freshet period (Amon  
515 et al., 2012). Our results suggest impacts to carbon of differing quality, as Amon et al.  
516 (2012) reported that lignin phenol and p-hydroxybenzene composition of Arctic river  
517 DOC point to the abundance of young, boreal-vegetation-derived leachates during  
518 spring flood and older, soil-, peat-, and wetland-derived DOC during groundwater  
519 dominated low flow conditions. In northern tundra areas where soil carbon amounts  
520 are greater, warmer temperatures and increased runoff will likely lead to increased  
521 riverine DOC exports. Indeed, Frey and Smith (2005) concluded that, assuming no  
522 change in either river discharge or in-channel processes, warming would produce a  
523 2.7–4.4 Tg yr<sup>-1</sup> increase in terrestrial DOC flux from West Siberia to the Arctic  
524 Ocean by 2100, with even larger increases likely should river discharge from the  
525 region continue to increase, as depicted in the simulations examined here. Warming  
526 and shifting snowmelt dynamics could increase transport and mobilization of DOC  
527 as subsurface pathways become active earlier in the year (Croghan et al., 2023). In  
528 contrast, some areas may experience a decrease in DOC export over time due to  
529 longer flow paths and residence times, along with increased microbial mineralization  
530 of DOC in the soil column (Striegl et al., 2005). Increasing soil thaw is expected to  
531 accelerate the release of old carbon (Dean et al., 2018; Schwab et al., 2020), which in  
532 turn will be entrained into, processed by, and exported from Arctic rivers. Moreover,  
533 DOC from deep sediments (> 3 m) could also become a significant contribution  
534 of carbon to Arctic rivers as the climate continues to warm (Mohammed et al.,  
535 2022). Nitrate concentrations are greater at lower latitudes as compared with higher  
536 latitudes where permafrost is more prominent (Frey and McClelland, 2009). Changes  
537 expressed predominantly across northern parts of the Arctic basin will have a direct



538 influence on coastal zone processes. On balance, our results point to continued  
539 increases in DOC export by Arctic rivers, and the mobilization and transport of  
540 ancient carbon in subsurface runoff from permafrost areas.

541 The use of two climate model forcing sets increases confidence in elements of  
542 the model outputs and associated analysis. It is noteworthy that results involving  
543 runoff, in particular the spatial patterns, are similar between the two simulations.  
544 Magnitudes of air temperature and precipitation increases are greater in the sim-  
545 ulation forced with IPSL (PWBM-IPSL). Under those warmer temperatures, the  
546 Hamon potential evapotranspiration function captures the temperature dependence  
547 on actual and potential evapotranspiration. Higher precipitation rates in a warmer  
548 forcing scenario, like IPSL, are offset by higher simulated ET, resulting in relatively  
549 similar magnitudes of annual net precipitation and annual total runoff. This plausi-  
550 ble modeling result suggests less uncertainty with the magnitudes of runoff changes  
551 compared with the changes in meteorological forcings projected by the climate mod-  
552 els. The model validation analysis suggests that the magnitude of simulated annual  
553 total runoff and discharge are comparable to independent observational datasets,  
554 with time trends similar in magnitude to those reported in other studies.

555 Salient conclusions from this study come with caveats related to the limits of the  
556 analysis. Foremost is the large degree of uncertainty in meteorological data across  
557 Arctic regions, attributable to a sparse observation network, as well as uncertain-  
558 ties in the magnitude of meteorological changes projected by the two coupled climate  
559 models. This uncertainty is ameliorated somewhat through the use of reanalysis data  
560 and model calibration. Results are implicitly linked to the connection between land-  
561 scape runoff and river discharge export. Results are also influenced by the choice of  
562 climate model forced under the SSP3-7.0 scenario. In light of this, one might expect  
563 lower magnitudes of change should atmospheric greenhouse gas concentrations not  
564 rise to levels depicted in SSP3-7.0. The broad spatial extent and moderate model res-  
565 olution (25×25 km gridcells) employed in this study limit our ability to incorporate  
566 influences such as thermokarst and talik formation on runoff contributions to streams  
567 and rivers. However, it is not clear that these local processes are a major component  
568 of riverine materials exports by Arctic rivers (Dean et al., 2018). The model simula-  
569 tions do not include interactions between lakes and the river networks, so, impacts  
570 from lake thaw drainage events (Smith et al., 2005; Andresen and Lougheed, 2015;  
571 Jones et al., 2022) are not simulated. The influence of land subsidence on soil temper-  
572 ature, moisture, and water storage is also not simulated. While subsidence is unlikely  
573 to lead to abrupt thaw over large areas, it can have significant effects on the hydrology  
574 of polygonal tundra, generally increasing landscape runoff (Painter et al., 2023).  
575 Our results underscore the importance in better understanding the myriad transfor-





576 mations reshaping Arctic environments. Large changes in the far north emphasize  
577 the need for more frequent and spatially extensive sampling of small and medium-  
578 sized rivers that ring the Arctic Ocean. Increased confidence in the magnitude of  
579 likely responses will require a multi-model, multi-scenario ensemble of simulations  
580 to obtain a range of projections consistent with known uncertainties. Incorporating  
581 small-scale effects such as thermokarst and lake drainage on river discharge will re-  
582 quire higher-resolution simulations. New model parameterization obtained from high  
583 resolution remote sensing observations will improve model capabilities in simulating  
584 permafrost hydrology in data sparse regions of the Arctic.

## 585 **6 Code and data availability**

586 This study is based on publicly available data for observa-  
587 tions used in model validation. The W5E5 data are available at  
588 <https://dataservices.gfz-potsdam.de/pik/showshort.php?id=escidoc:4855898>  
589 (last access: 15 October 2022). The MERRA reanalysis data are avail-  
590 able at <https://gmao.gsfc.nasa.gov/reanalysis/MERRA/> (last access: 23 Janu-  
591 ary 2023). The ECMWF Reanalysis v5 (ERA5) data are available at  
592 <https://www.ecmwf.int/en/forecasts/dataset/ecmwf-reanalysis-v5> (last access:  
593 19 March 2023). The TPCDC data are available at <http://data.tpcdc.ac.cn/en>  
594 (last access: 3 February 2023). The IPA permafrost data in the Circum-  
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596 available at <https://nsidc.org/data/ggd318/versions/2> (last access 1 Au-  
597 gust 2022). The Global Land Evaporation Amsterdam Model (GLEAM)  
598 data are available at <https://www.gleam.eu/> (last access: 17 April 2023).  
599 The pan-Arctic ET data derived from remote sensing are available at  
600 [http://files.ntsg.umt.edu/data/PA\\_Monthly\\_ET/](http://files.ntsg.umt.edu/data/PA_Monthly_ET/) (last access: 16 April 2023).  
601 Climate model data used as forcings are available in the ISIMIP Repository  
602 located at <https://data.isimip.org/>. The PWBM source code is available at  
603 <https://blogs.umass.edu/csdc/pwbm/>. The climate model forcings and PWBM  
604 simulation outputs are available upon request from the authors of this study.

## 605 **7 Author contributions**

606 MAR ran all the simulations, analyzed the results and wrote the paper. AVK  
607 prepared the climate modeling forcing data and contributed to writing of the paper.



## 608 **8 Competing interests**

609 The authors declare that they have no conflict of interest.

## 610 **9 Acknowledgements**

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620 OPP-1656026).

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