

Recent ground thermo-hydrological changes in a Southern Tibetan endorheic catchment and implications for lake level changes

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1 **Abstract**

2 Climate change modifies the water and energy fluxes between the atmosphere and the surface in
3 mountainous regions such as the Qinghai-Tibet Plateau (QTP), which has shown substantial hydrological
4 changes over the last decades, including rapid lake level variations. The ground across the QTP hosts either
5 permafrost or seasonally frozen and, in this environment, the ground thermal regime influences liquid water
6 availability, evaporation and runoff. Consequently, climate-induced changes in the ground thermal regime
7 may contribute to variations in lake levels, but the validity of this hypothesis has yet to be established.

8 This study focuses on the cryo-hydrology of the catchment of Lake Paiku (Southern Tibet) for the 1980-
9 2019 period. We use TopoSCALE and TopoSUB to downscale ERA5 data, in an effort to account for the
10 spatial variability of the climate in our forcing data. We use a distributed setup of the CryoGrid community
11 model (version 1.0) to quantify thermo-hydrological changes in the ground during this period. Forcing data
12 and simulation outputs are validated with data from a weather station, surface temperature loggers and
13 observations of lake level variations. Our lake budget reconstruction shows that the main water input to the
14 lake is direct precipitation (310 mm per year), followed by glacier runoff (280 mm per year) and land runoff
15 (180 mm per year). However, altogether these components do not offset evaporation (860 mm per year).

16 Our results show that both seasonal frozen ground and permafrost have warmed (0.17 °C per decade 2
17 m deep), increasing the availability of liquid water in the ground and the duration of seasonal thaw.
18 Correlations with annual values suggest that both phenomena promote evaporation and runoff. Yet, ground
19 warming drives a strong increase in subsurface runoff, so that the runoff/(evaporation + runoff) ratio
20 increases over time. This increase likely contributed to stabilizing the lake level decrease after 2010.

21 Summer evaporation is an important energy sink and we find active layer deepening only where
22 evaporation is limited. The presence of permafrost is found to promote evaporation at the expense of runoff,
23 consistent with recent studies suggesting that a shallow active layer maintains higher water contents close to
24 the surface. However, this relationship seems to be climate-dependent and we show that a colder and wetter
25 climate produces the opposite effect. Although the present study was performed at catchment scale, we
26 suggest that this ambivalent influence of permafrost may help to understand the contrasting lake level
27 variations observed between the South and North of the QTP, opening new perspectives for future
28 investigations.

29 **Main text**

30 **1. Introduction**

31 Climate change is amplified in mountainous environments, with major consequences for
32 ecosystems, landscapes, hydrology, human communities and infrastructure (IPCC, 2019). Station
33 observations show that global warming is elevation dependent, with the strongest warming rates being
34 observed at high elevations (Pepin et al., 2015; Wang et al., 2014). Over the Qinghai-Tibet Plateau
35 (QTP), a significant increase in surface air temperatures has been recorded since the 1980s, in particular
36 in the North of the plateau (Zhang et al., 2022). This has been accompanied by a decrease in wind speed,
37 humidification of the air, and a general increase in precipitation, although with a strong spatial
38 variability (Bibi et al., 2018). Altogether, these changes have affected the surface energy balance of the
39 plateau through a shift of the Bowen ratio towards more latent heat fluxes, limiting the sensible surface
40 warming (Yang et al., 2014a).

41 These changes in water and energy fluxes between the atmosphere and the surface have the potential
42 to alter the hydrological cycle of the QTP, which is the headwater region for major Asian rivers. As
43 such, increasing trends of evaporation over land have been measured (3.8 mm per decade since the
44 1960s) with strong spatial variability both in absolute values and increase rates (Wang et al., 2020b).
45 Changes in the seasonality of river discharge (Cao et al., 2006) and groundwater discharge (Niu et al.,
46 2011) were reported for the same period. Overall glacier shrinkage has also been observed since the
47 1960s with a persistent increase in glacier mass loss rates (Bhattacharya et al., 2021; Hugonnet et al.,
48 2021).

49 The QTP also features more than 1,000 lakes larger than 1 km² (Zhang et al., 2017), most of them
50 located in endorheic catchments. Lake volume changes are therefore attributable to climatic and
51 hydrological changes occurring within the lake catchment, such as glacier melt, ground ice melt,
52 precipitation, evaporation or runoff patterns. The majority of these lakes have experienced a pronounced
53 increase in water levels since the 1990s (Lei et al., 2013, 2014), a trend that was suggested to be mainly
54 driven by changes in precipitation and evaporation patterns (Yao et al., 2018) rather than by an increase

55 in glacier mass loss and runoff (Brun et al., 2020; Zhang et al., 2021a). Nevertheless, lake level
56 variations are not uniform across the QTP and exhibit important spatial variability. Whereas the
57 northern and central QTP have recorded lake expansion, the southern parts of the plateau have
58 experienced lake shrinkage (Qiao et al., 2019; Zhang et al., 2021a, 2020a). Shrinking lakes have
59 received less attention in the literature than rising lakes because they are fewer. For this reason the
60 drivers of this shrinkage are still unclear. Qiao et al. (2019) reported that recent lake shrinkage over the
61 QTP could be driven by local precipitation decrease and/or evaporation increase (in relation to air
62 temperature increase). Zhang et al. (2020a) suggests that the divergent trends in lake level variations
63 across the QTP could be linked to the contrasting evolution of moisture transport between the north and
64 south of the plateau. On longer timescales, lake shrinkage over the QTP during the Holocene seems to
65 be related to variations in the intensity of the Asian monsoon (Chen et al., 2013). Overall, such a
66 complex pattern of rising and shrinking lakes challenges our understanding of the hydrological changes
67 occurring in these high Asian watersheds.

68 In this regard, new insights on hydroclimatic changes over the QTP can emerge from the
69 investigation of the coupled energy and water fluxes between the ground surface/subsurface and the
70 atmospheric boundary layer. These fluxes are driven by the climate and have a major impact on cold-
71 region hydrology (Pomeroy et al., 2007; Gao et al., 2021; Bring et al., 2016). Indeed, hydrological
72 variables (precipitation, evaporation, runoff) affect the soil water content, which changes its thermal
73 properties, the distribution between latent and sensible fluxes and thus substantially influences the
74 ground thermal regime (Bring et al., 2016; Koren et al., 1999; Martin et al., 2019). In turn, the ground
75 thermal regime modifies the relative proportion of frozen and liquid subsurface water, influencing
76 infiltration possibilities and the amount of water available for evaporation and surface/subsurface runoff
77 (Yi et al., 2006; Carey and Woo, 2001).

78 So far, climate induced thermo-hydrological changes over the QTP have received limited attention.
79 Large-scale modeling studies reported changes in the seasonal ground freezing cycles characterized by
80 a reduction of the frost depth and duration of the frozen period since the 1960s (Qin et al., 2018; Wang
81 et al., 2020a) and notable ground warming trends in summer and winter (Qin et al., 2021). Similar
82 ground warming trends were reported in the regional modeling study from Qin et al. (2017), along with

83 an increasing trend in evaporation and a decrease of the runoff coefficient over time. Plateau-scale
84 surface energy balance modeling from Wang et al. (2020b) reported that increasing trends in
85 evapotranspiration could be mainly explained by variations in air temperature and net radiation at the
86 surface.

87 Complementary to seasonally frozen ground, permafrost is also a distinctive feature of climate-
88 surface interactions in cold regions. Large-scale permafrost modeling suggests that it covers a
89 significant part of the QTP, mainly as continuous permafrost in the north of the plateau and as
90 discontinuous or sporadic in the south (Obu et al., 2019). Permafrost on the QTP usually has a low ice
91 content due to limited precipitation and strong evaporation (Wu et al., 2005; Yang et al., 2010).
92 Borehole temperature measurements show that it is a relatively warm type of permafrost (Biskaborn et
93 al., 2019; Wu and Zhang, 2008) and its exposure to high solar radiations makes it sensitive to changes
94 in surface conditions and climate change (Yang et al., 2010). Since the 1960s, climate change has driven
95 permafrost warming across the plateau (Ran et al., 2018; Shaoling et al., 2000). Ran et al. (2018) reports
96 that most of the plateau exhibits a warming trend of the ground comprised between 0.26 and 0.74 °C
97 per decade and half of the plateau warms at a rate higher than 0.5 °C per decade. This warming is
98 accompanied by upward migration (of around 100 m between the 1960s and 2000s) and shrinkage of
99 permafrost covered areas (24% of the permafrost extent lost between the 1960s and the 2000s, Ran et
100 al., 2018).

101 Permafrost grounds are characterized by a strong interplay between the ground thermal regime and
102 the land hydrology. Seasonal thawing and freezing of the active layer are driven by the surface energy
103 balance which, in return, influences surface and subsurface runoff (Kurylyk et al., 2014; Walvoord and
104 Kurylyk, 2016; Sjöberg et al., 2021) and evaporation (Gao et al., 2021). In this regard, both large-scale
105 and regional modeling indicate that thawing permafrost enhances evapotranspiration (Qin et al., 2017;
106 Wang et al., 2020b). Qin et al. (2017) also report that the increase in evaporation is logically
107 concomitant with a decrease in the runoff coefficient. Additionally, permafrost stores water as ground
108 ice and its thawing can trigger the release of liquid water in the watershed, contributing up to 15% of
109 the annual river streamflow (Cheng and Jin, 2013; Yang et al., 2019).

110 These hydrological changes are tied to various interdependent climate-driven physical processes
111 happening at the ground surface and subsurface (e.g. surface energy balance, infiltration, water phase
112 change, heat conduction...). Because these processes exhibit a strong spatial variability in high mountain
113 environments, it is challenging to represent them accurately together on large spatial scales. Therefore,
114 a deeper understanding of the impact of ground thermo-hydrological changes on the High Asia water
115 cycle can be gained through small-scale physical modeling of these processes. Yet, for now, physics-
116 based approaches at the catchment scale aiming to connect the ground thermo-hydrological regime and
117 the observed hydrological changes on the QTP (such as lake level changes) remain scarce. They are
118 however a powerful approach to tackle the question: how much might climate-driven ground thermal
119 changes affect the water cycle in high mountain headwater regions? In this study, we use physical land
120 surface modeling to quantify the ground thermo-hydrological changes in an endorheic Tibetan
121 catchment over the last 40 years as a response to climate change. We show the interplay in the water
122 and energy fluxes occurring between the atmosphere, the surface and the subsurface and discuss their
123 impact on the hydrology of the catchment and their implication regarding lake level variations.

124 **2. Study area: the Paiku catchment**

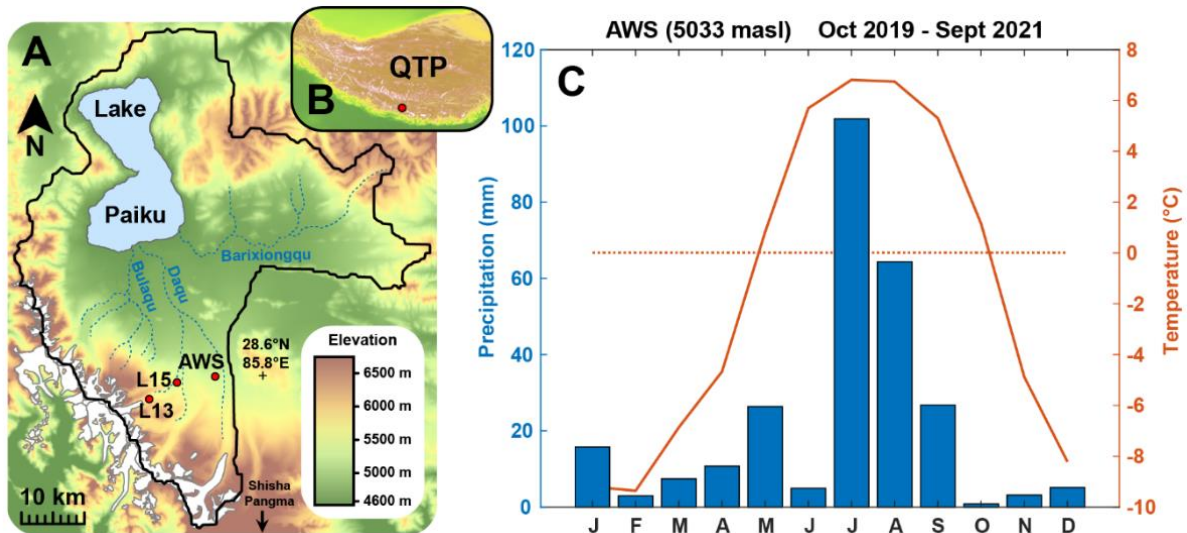
125 The Paiku catchment is located in south-western Tibet, China, close to the border with Nepal
126 (28.8°N - 85.6°E, Fig. 1). Its southern edge lies 7 km from the Shishapangma peak (8027 masl). The
127 catchment is endorheic and spans over 78 km from North to South, 66 km from East to West and covers
128 2 400 km². The median elevation of the catchment is 4872 masl, ranging from 7272 masl to its lowest
129 point, lake Paiku at 4580 masl. Geologically, the catchment is mainly located in the Tethys Himalayan,
130 and thus, an important part of the formations underlying the catchment are metamorphized sedimentary
131 series (Appendix B, Fig. B1). The southern part of the catchment crosses the Southern Tibetan
132 Detachment, and thus, the southern ridges of the massif belong to the High Himalayan metamorphic
133 formations in the west and to the High Himalayan leucogranites of the Shishapangma massif on the
134 east. The north and north-east ridges are formed by granite intrusions surrounded by metamorphic
135 domes. The inner part of the catchment presents Plio-Quaternary formations such as alluvial fans close
136 to the ridges and inclined alluvial plains in its inner parts (Aoya et al., 2005; Searle et al., 1997;
137 Wünnemann et al., 2015).

138 Automatic Weather Station (AWS) observations (5033 masl, Oct 2019 – Sept 2021, Fig. 1) show
139 that the climate in the catchment is characterized by a relatively small temperature amplitude during the
140 year (around 20 °C, JJA being the warmest months and DJF the coldest) and significant daily amplitude
141 (up to 10 °C during the warm season). The mean annual temperature is -1.5 °C at the AWS, where night
142 freezing can occur until the beginning of June and resume at the beginning of October. The catchment
143 is dry (200-300 mm year⁻¹) and precipitation mostly falls as rain during the monsoon (JJAS).

144 Around 5% of the catchment is covered by glaciers (RGI Consortium, 2017), which are
145 concentrated in its southwestern part. They feed several proglacial lakes that can reach up to 6 km in
146 length. Geodetic glacier mass budgets show that, similar to other glaciers in the region, glaciers of the
147 Paiku catchment have undergone sustained mass loss at least since the 1970s, with an average mass
148 balance of -0.3 m w.e.a⁻¹ until the beginning of the 2000s and around -0.4 m w.e.a⁻¹ thereafter
149 (Bhattacharya et al., 2021). There are more than 10 rivers that drain the catchment towards the lake and
150 most of them only exhibit a seasonal activity during the monsoon months. The three main ones are

151 (Fig. 1), Daqu (glacier-fed, 450 km²), Bulaqu (glacier-fed, 325 km²) and Barixiongqu (non-glacier-fed,
152 703 km², Lei et al., 2018).

153 In the north-west of the catchment, Lake Paiku covers approx. 280 km² (11.5% of the catchment
154 surface area) and spans over 27 km from North to South. It has a mean water depth of 41 m, with a
155 maximum water depth of 73 m (Lei et al., 2018). It receives water from direct precipitation and from
156 land and glacier runoff which can be routed at the surface via the river systems or the subsurface via
157 the alluvial formations. Because it is hydrologically closed, the lake mainly loses water through
158 evaporation. Previous studies reported lake level fluctuations over different time scales. It reached 4665
159 masl (85 m higher than the present level) prior to 25 ka BP and at the onset of the Holocene (11.9-9.5
160 ka BP), afterwards, the lake shrank gradually (Wünnemann et al., 2015). More recently, the lake level
161 decreased by 3.7 m between 1972 and 2015, losing 4.2% of its surface and 8.5% of its volume.
162 Measurements have been performed since the end of the 1970s and allow to accurately know the
163 evolution of the lake level until today (Lei et al., 2021, 2018), they are used in this study to validate our
164 hydrological results (Sect 3.2.1, Fig. 5D and 6B). At the seasonal scale, the lake level cycle has an
165 amplitude of ~ 0.4 m. It is marked by a strong increase during the monsoon period (JJAS) supported by
166 direct precipitation, glacier melt and land runoff. From October and until the next monsoon period,
167 evaporation dominates the lake mass budget and the level decreases rapidly until January and at a slower
168 rate afterwards (Lei et al., 2021).



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Figure 1. The Paiku Catchment. A: Topographic and hydrologic map of the catchment with the glaciers in white, the ephemeral rivers in dark blue and the lake in light blue (elevation: SRTM data courtesy of the U.S. Geological Survey). AWS: Automatic Weather Station. L13 and L15 are surface temperature loggers (Sect. 3.1). B: Localization of the catchment over the QTP. C: Monthly temperature and precipitation recorded at the AWS between October 2019 and September 2021.

175 **3. Material and methods**

176 3.1. Field measurements

177 An AWS was set up in October 2019 in the South of the catchment at an elevation of 5033 masl
178 (Fig. 1). It is equipped with various sensors which record air temperature, pressure, relative humidity,
179 wind speed, incoming and outgoing long and short wave radiations and precipitation every 15 minutes.
180 The meteorological record extends to September 2021 and covers a period of nearly 2 years. We used
181 it to evaluate and correct the distributed downscaled climatic forcing on which we rely in our modeling
182 framework (Sect. 3.2.2.).

183 Two temperature loggers recorded the surface temperature in the vicinity of the AWS location.
184 Logger 15 (L15) is located at 5055 masl, 6 km west of the AWS. Logger 13 (L13) is located at 5356
185 masl, 12 km west of the AWS (Fig. 1). Both loggers were buried 10 to 15 cm below the surface to avoid
186 direct solar radiation on the sensors and recorded surface temperature at a 20-minute timestep from
187 October 2017 to October 2018. These surface temperature records were used to evaluate the simulations
188 (Sect. 3.2.4.).

189 3.2. Catchment thermo-hydrological modeling

190 3.2.1. *Conceptual hydrological model for the catchment*

191 To understand the level variations of lake Paiku over the last 40 years (1980-2019 period), we
192 develop an approach at the catchment scale. Because the catchment is hydrologically closed, the lake
193 receives water input via direct precipitation, land surface and subsurface runoff, and glacier runoff.
194 Conversely, it loses mass via evaporation. Because the quantification of water flows between the lake
195 and potential aquifers surrounding it is difficult (Rosenberry et al., 2015), our approach assumes that
196 these flows are negligible. The present study requires quantification of the different terms of the
197 hydrological balance. Under these assumptions, the hydrological balance of the lake is given by the
198 following equation:

$$\Delta Z_{\text{Lake}} = \text{Precipitation}_{\text{Lake}} + \text{Runoff}_{\text{LandSurf}} + \text{Runoff}_{\text{LandSub}} + \text{Runoff}_{\text{Glacier}} - \text{Evaporation}_{\text{Lake}}$$

200 The production of forcing data for the catchment (including precipitation) is detailed in Sect. 3.2.2.

201 The land hydrology processes are quantified using the CryoGrid community model (version 1.0)

202 (Westermann et al., 2023) as described in Sect. 3.2.3. Distributed 1D simulations are used to quantify

203 land evaporation and runoff. The routing of water in the catchment is not represented and the runoff

204 computed for a given simulation is directly accounted as a water input for the lake. The evaporation

205 from the lake is simulated using the CryoGrid3-Flake model (Langer et al., 2016) as described in

206 Sect. 3.2.5. Glacier melt is not modeled, but estimated for the study period (1980-2019) from remote

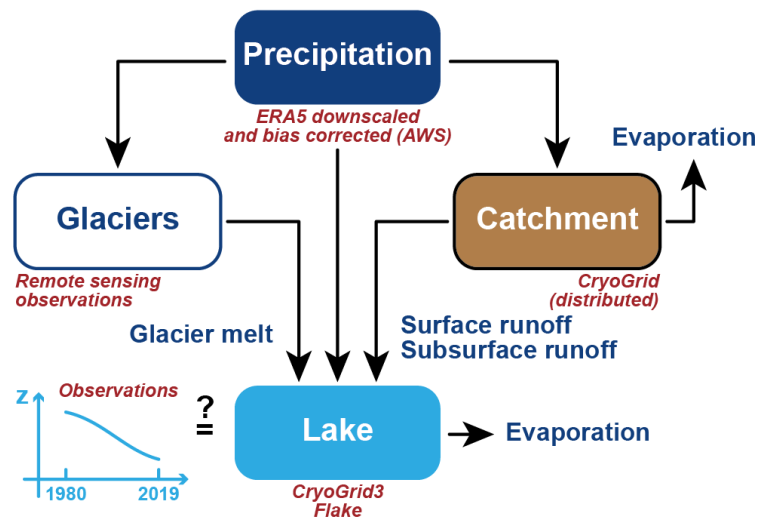
207 sensing observations. From these observations, glacier yield is calculated as described in Sect. 3.2.6.

208 Our catchment-scale approach to represent the hydrological balance of the lake is summarized in Fig. 2.

209 Based on this approach, we can evaluate the performance of our framework (Sect. 4.1.2.), by comparing

210 the simulated lake balance with the one derived from the detailed observations of lake level variations

211 over the study period (Lei et al., 2018, 2021).



212 *Figure 2. Conceptual hydrological framework for the study.*

214 3.2.2. Forcing data production and validation

216 In high mountain environments, topography creates strong spatial variability of temperature and

217 incoming radiation, which impact the surface energy balance (Klok and Oerlemans, 2002) and the

218 ground thermo-hydrological regime (Magnin et al., 2017). Our approach requires forcing data that (i)

219 captures this variability, (ii) includes numerous variables such as air temperature, incoming long and

220 short wave radiations, wind speed, specific humidity, rain and snowfall and (iii) covers the 40 years

221 study period at a sub-daily timestep. The TopoSCALE approach (Fiddes and Gruber, 2014) was
222 developed for this purpose and allows to downscale reanalysis products like ERA5 (Hersbach et al.,
223 2020) at high resolution (here $\sim 100 \times 100$ m).

224 Additionally, because working at a 10^{-2} km² spatial resolution over a 2400 km² catchment would
225 require more than 200,000 forcing files and simulations, we rely on the TopoSUB method (Fiddes and
226 Gruber, 2012) to reduce computational costs. This method uses a SRTM30 Digital Elevation Model to
227 explore redundancies in physiographic parameters of the study area such as elevation, aspect, slope and
228 sky-view factor and to identify groups of high-resolution pixels (100 x 100 m) sharing similar values
229 for these parameters. From there, all the high-resolution pixels belonging to such a group are only
230 described as a single TopoSUB point, for which climatic variables can be downscaled to create one
231 single dataset of climatic timeseries. The degree of similarity required by TopoSUB to identify groups
232 of high-resolution pixels with redundant physiographic parameters can be adjusted by choosing the final
233 number of TopoSUB points (and thus climate datasets) that should be used to cover the area
234 corresponding to one ERA5 pixel. The Paiku catchment intersects 8 ERA5 pixels at 30 km resolution
235 and we chose to use 50 TopoSUB points within each ERA5 pixel to cover the spatial variability created
236 by the topography on small-scale climate. Ultimately, 368 TopoSUB points are used to cover the
237 catchment. The average level of redundancy (i.e. the average number of high-resolution pixels
238 represented by a single TopoSUB point) is 723 ± 745 (1σ , median: 506, min: 1, max: 4347). Appendix
239 C, Fig. C1 shows the distribution of the TopoSUB points and a reconstruction of the topography of the
240 catchment based on this approach. The period covered by the forcing datasets starts on 1st January 1980
241 and ends on 31st August 2020 (40 years and 8 months).

242 In the TopoSCALE statistical downscaling approach, we do not rely on the AWS data and thus the
243 downscaled ERA5 data can be biased, as is often the case over Asia (Jiang et al., 2021, 2020; Jiao et
244 al., 2021; Orsolini et al., 2019). Comparison against the available AWS observations (Appendix D, Fig.
245 D1) indeed highlights notable differences in variables such as air temperature and precipitation. From
246 these differences, we derived monthly bias correction factors that we applied systematically to all of the
247 368 climate forcing datasets. The catchment averages for precipitation and air temperatures are shown
248 in Fig. 3. In this figure and across the rest of the study, we use p-values to evaluate the significance of

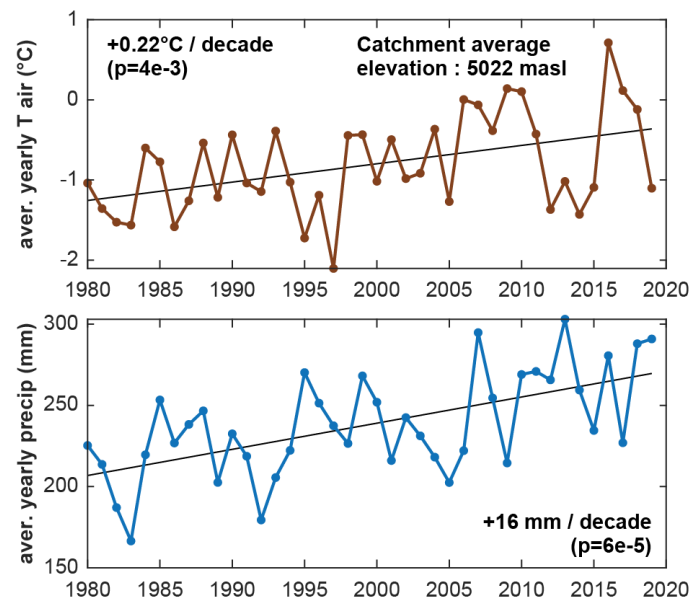
249 linear trends in the temporal evolution of certain variables (temperature, precipitation, evaporation...).

250 This p-value tests the null hypothesis which supposes that the value of the slope is equal to zero. The

251 hypothesis is tested using the Student's t-test, by comparing the distance between the estimated slope

252 and 0, relative to the standard error of the slope. We did not report trends when this p-value (probability

253 of a null slope) was higher than 0.005.



254

255 *Figure 3. Climate forcing data for the land and lake modeling. Annual catchment-average air*

256 *temperature (2 m above ground) and annual total precipitation for the study period. Note that the model*

257 *is also forced by incoming short and long wave radiations, humidity, windspeed and air pressure.*

258 *Details about the spatial and temporal resolution of the distributed forcing data are presented in Sect.*

259 *3.2.2.*

260

261 **3.2.3. The CryoGrid community model (version 1.0)**

262 To simulate the ground thermo-hydrological regime, we use the CryoGrid community model

263 (Westermann et al., 2023). The CryoGrid community model (CG) is a land surface model designed for

264 applications in cold regions where seasonal frozen ground or permafrost may occur. The model

265 implements heat transfer in a 1D soil column, accounting for freeze-thaw processes of soil water using

266 effective heat capacity (Nakano and Brown, 1972). To do so, soil freezing curves are based on

267 Dall'Amico et al. (2011) as detailed in Westermann et al. (2013). Vertical water movement in the soil

268 column is based on Richards equation (Richardson, 1922; Richards, 1931). The soil matric potential

269 and hydraulic conductivity follow van Genuchten (1980) and Mualem (1976). Additionally, to represent

270 the obstruction of connected porosity by ice formation, the hydraulic conductivity is reduced by a factor

271 dependent on the local ice content, following Dall'Amico et al. (2011). The model features the

272 snowpack module called *CG Crocus* described in Zweigel et al. (2021) that adapts the snow physics
273 parameterizations from the CROCUS scheme (Vionnet et al., 2012) to the native snow module of
274 CryoGrid3 (Westermann et al., 2016). At the surface, the model uses a surface energy balance module
275 to calculate the ground surface temperature and water content. The turbulent fluxes of sensible and
276 latent heat are calculated using a Monin–Obukhov approach (Monin and Obukhov, 1954). Evaporation
277 is derived from the latent heat fluxes using the latent heat of evaporation and is adjusted to the available
278 water in the soil. It occurs in the first grid cell only, but water can be drawn upwards due to matric
279 potential differences. Because vegetation is very scarce in the catchment, we do not expect transpiration
280 to have a strong imprint on evapotranspiration and our calculations do not unravel evaporation from
281 transpiration.

282 3.2.4. *Model setup and validation*

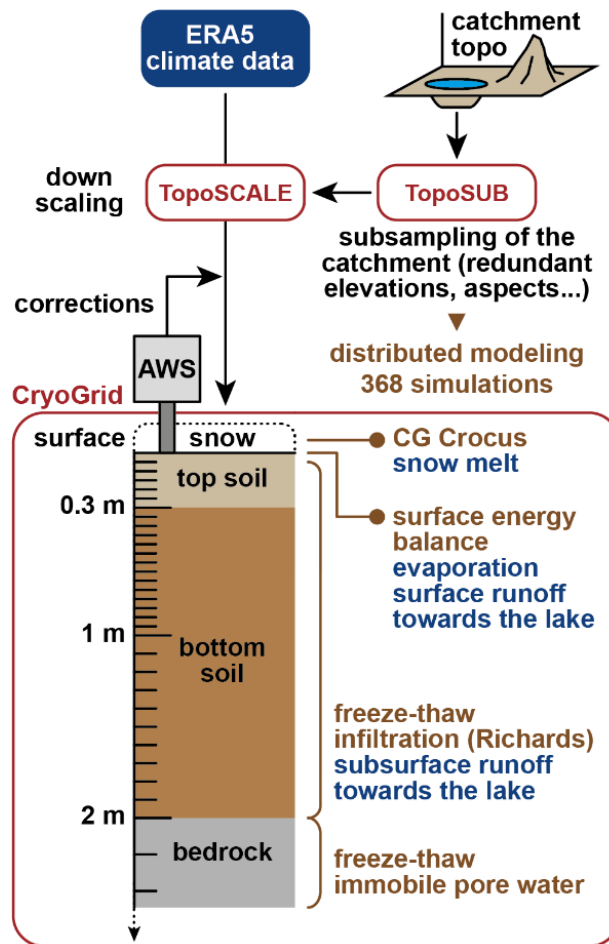
283 The setup of the CryoGrid community model for the land is presented in Fig. 4. To capture the
284 high spatial variability of mountainous climate, our approach relies on the 368 climate forcing datasets
285 to cover the catchment (see Sect. 3.2.2.). This approach enables us to perform spatially distributed
286 modeling. All of the 368 simulations are independent and use the same parameterization. In absence of
287 direct observation of the soil stratigraphy within the catchment, the soil column was designed to agree
288 with field observations in the region (Yuan et al., 2020; Wang et al., 2009; Hu et al., 2020; Luo et al.,
289 2020; Yang et al., 2014b; Wang et al., 2008), to be consistent with similar modeling approaches across
290 Tibet (Chen et al., 2018; Song et al., 2020) and to be consistent with input datasets (Shangguan et al.,
291 2013, 2017). Thus, the soil stratigraphy is divided into 3 units: a top soil (0.3 m thick), a bottom soil
292 (1.7 m thick), and a bedrock unit (extending beyond the depth of interest of the study). An overview of
293 the parameters for each unit, their source and the way they are calculated is presented in Appendix A,
294 Tab. A1.

295 Regarding the processes implemented in the model (Sect. 3.2.3.), infiltration according to Richards
296 equation only occurs in the top and bottom soil units. The bedrock unit has a static water content.
297 Unraveling surface from subsurface flow is an ongoing challenge in catchment-scale hydrology
298 (McDonnell, 2013) and this distinction is important in mountain terrains where these two flows can
299 behave differently due to the complex topography (Seibert et al., 2003; Gao et al., 2014; Hu et al.,

2020). For this study, we rely on a simple approach that is based on thresholds regarding the soil water content (porosity and field capacity). This kind of approaches are thus based on soil properties and have often been used in hydrological modeling studies (Vörösmarty et al., 1989; Shaman et al., 2002; Kelleners et al., 2010; Kampf, 2011; Samuel et al., 2008). In detail, we compute surface and subsurface flow as follows.

On the one hand, surface runoff is computed relative to the saturation level of the soil column. When the entire soil column is saturated ($WC = \text{porosity}$), additional water input from precipitation or snowmelt is directly counted as surface runoff. On the other hand, subsurface runoff is computed relative to the field capacity of the ground, which is an input parameter of the model. When the water content (WC) of a ground cell exceeds this field capacity (FC), the amount of water corresponding to $WC - FC$ is available to produce subsurface runoff. We use the lateral boundary condition `LAT_WATER_RESERVOIR` from the CryoGrid community model (Westermann et al., 2023) to account for this subsurface runoff. The speed at which this available water exits the soil column towards the lake is calculated with Darcy's law, using the hydrological conductivity of the ground and the mean slope of the catchment as hydraulic slope. Because the model couples thermal and hydrological fluxes, all of these changes in the soil water content can be driven by precipitation input, evaporation but also water phase change in the ground such as ice melt.

Because we do not have knowledge of the distributed thermal state with depth over the catchment at the beginning of the simulations, we assume temperature profiles were in equilibrium with the climate of the 5 first years of modeling (1980-1984). To do so, we start our simulations with a 60-year spin-up of these first 5 years (12 repetitions), which is sufficient to establish a stable temperature profile over the first 9 to 80 meters depending on the simulations, extending beyond the hydrologically active part of the ground (the first 2 meters).



323
 324 *Figure 4. Modeling framework for the land hydrology. ERA5 data are downscaled using the TopoSUB*
 325 *and TopoSCALE approaches (Fiddes and Gruber, 2014, 2012). The downscaled data are bias-*
 326 *corrected based on the AWS observations. Distributed 1D simulations are performed using the*
 327 *CryoGrid community model (Westermann et al., 2023). The vertical resolution is indicated with the tick*
 328 *marks on the depth axis.*

329
 330 To validate model simulations, the simulated ground surface temperatures (GST) are compared to
 331 the two temperature logger time series acquired in the vicinity of the AWS (Sect. 3.1.). We used this
 332 comparison to calibrate the surface roughness used for the surface energy balance calculations in the
 333 model.

334 The following method is used to produce area-averaged evaporation and runoff (in mm water
 335 equivalent) in a zone of interest. For a given TopoSUB point in this zone, the model produces
 336 hydrological values in m^3 using the area of a TopoSUB pixel on the catchment map. Then these values
 337 are multiplied by the number of pixels in the zone corresponding to this TopoSUB point in particular,
 338 and this for all the relevant TopoSUB points covering the zone (e.g. evaporation in warm permafrost).
 339 Then the area of interest is calculated by counting the number of pixels in the zone of interest and

340 multiplying this number by the area of a pixel. Then the total volume is divided by the total surface for
341 the zone of interest to obtain the final value in mm.

342 *3.2.5. Lake modeling*

343 The lake thermo-hydrological response to the climatic forcing data is simulated using the
344 CryoGrid3-Flake model (Langer et al., 2016). The two models were coupled by Langer et al. (2016) to
345 simulate the thermal regime of thermokarst lakes (including surficial water freezing and melting) and
346 underlying ground. Here we use the coupled models mainly to quantify evaporation at the lake surface.
347 In the coupled model, the native surface energy balance module of CryoGrid3 (Westermann et al., 2016)
348 was amended to account for processes tied to free water surface energy balance: (i) the dependence of
349 the albedo of a water surface to solar angle (and thus time of the day) and wind speed (and wave
350 formation), (ii) the dependence of the surface roughness length to wind speed (and wave formation) and
351 (iii) the exponential decay of incoming radiation with depth in the water column. Similar to the land
352 simulations, the lake simulations were forced by the downscaled ERA5 data (with the TopoSUB and
353 TopoSCALE methodology), with the corrections derived from the AWS data (Sect. 3.2.2.). The
354 simulations were initiated with a 20-year spin-up of the 1980-1984 climate. The simulation results
355 corresponding to the four ERA5 tiles covering the lake were then averaged using the respective spatial
356 footprint of each tile on the lake.

357 *3.2.6. Quantification of glacier mass change*

358 Multiple studies quantified the volume change of the glaciers located within the Paiku catchment
359 in the recent past (1970s to 2020). To our knowledge, there are no field based measurements of glacier
360 mass balance available in this catchment. As a consequence, we rely solely on geodetic mass balance
361 studies (Brun et al., 2017; Maurer et al., 2019; King et al., 2019; Shean et al., 2020; Hugonnet et al.,
362 2021). All these studies estimated glacier volume changes over periods of 20-30 years from satellite
363 derived DEMs. As a consequence, we can only estimate the average annual glacier mass balance, and
364 not the year-to-year variability. Glaciers occupy approximately 113 km² in the Paiku catchment. They
365 have shrunk for the past fifty years at a rate of 0.44 % y⁻¹, from an area of 132 km² in 1975 to 122 km²
366 around 2000 and to their current extent (King et al., 2019; Bolch et al., 2019). The average mass
367 balances for the period 1975-2000 and 2000-2020 are $-3.9 \pm 2.1 \times 10^{10}$ kg y⁻¹ and $-5.4 \pm 2.4 \times 10^{10}$ kg y⁻¹,

368 respectively ($-4.6 \pm 2.5 \cdot 10^7 \text{ m}^3$ and $-6.4 \pm 2.8 \cdot 10^7 \text{ m}^3$ with a 850 kg m^{-3} density). These mass balances
369 correspond to specific mass balances of $-0.31 \pm 0.17 \text{ m}$ of water equivalent per year (w.e. y^{-1}) and -0.47
370 $\pm 0.21 \text{ m w.e. y}^{-1}$, respectively.

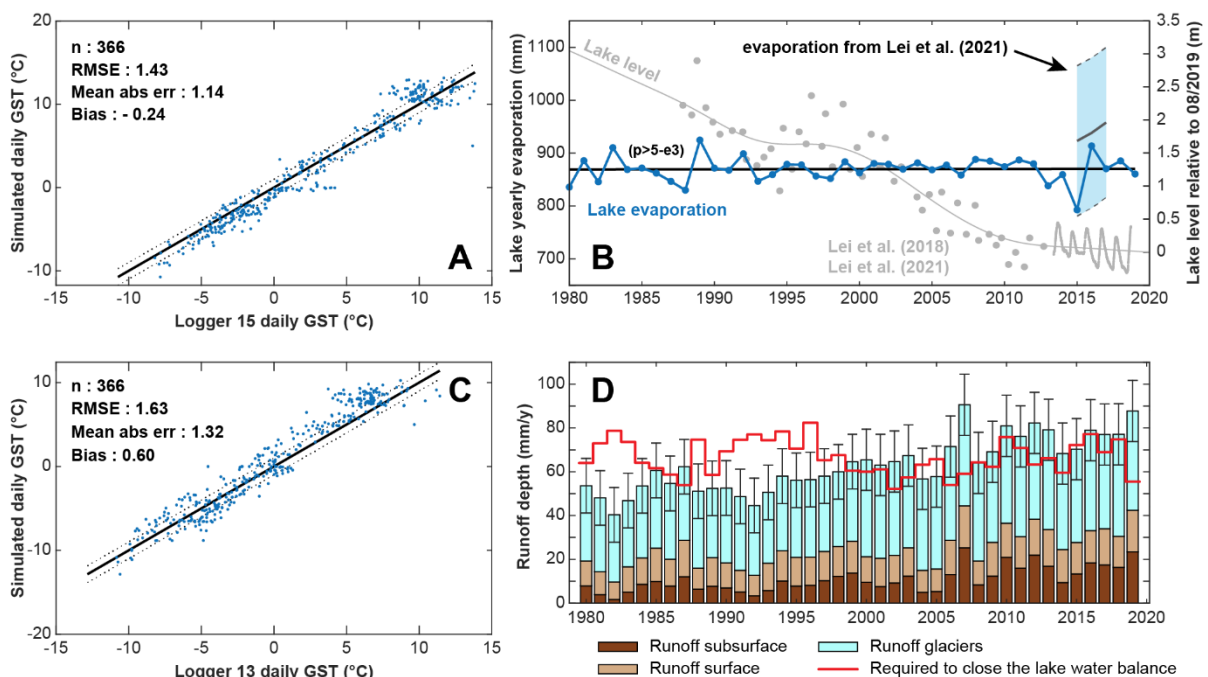
371 Regarding glacial runoff, it was estimated to $320 \pm 4 \text{ mm}$ per year for the 2001-2010 period by
372 Biskop et al. (2016) using a temperature-index approach for ice melt. For the 2000-2018 period, Zhang
373 et al. (2020b) derived a runoff value of $52 \pm 12 \text{ mm}$ per year ($1.24 \pm 0.29 \cdot 10^8 \text{ m}^3$ per year that we scaled
374 to the basin area). The value we derive of $39 \pm 13 \text{ mm}$ per year thus finds good consistency with the
375 latter one (Sect. 4.1.1.).

376 4. Results

377 4.1. Model validation and hydrological budget of Lake Paiku

378 4.1.1. Model validation

379 Simulated daily ground surface temperatures are in good agreement with the observed ones,
 380 showing a bias of $-0.2\text{ }^{\circ}\text{C}$ and $0.6\text{ }^{\circ}\text{C}$ and a RMSE of $1.4\text{ }^{\circ}\text{C}$ and $1.6\text{ }^{\circ}\text{C}$ for loggers 15 and 13,
 381 respectively (Fig. 5A and 5C). Most of this RMSE is explained by a mismatch between model and
 382 observations in the tails of the temperature distribution, whereas intermediate temperatures exhibit the
 383 best agreement with observations.



384 *Figure 5. Model validation. A and C: modeled mean daily ground surface temperatures compared to*
 385 *measured ground surface temperatures for logger 15 and logger 13 (location on Fig. 1). B: modeled*
 386 *annual lake evaporation (blue curve) and comparison with values calculated by Lei et al. (2021) in the*
 387 *light blue zone. The gray curve shows the smoothed lake level relative to August 2019 based on*
 388 *observations from Lei et al. (2018) (gray points) and Lei et al. (2021) (gray oscillating line). D:*
 389 *Comparison between the runoffs required to reproduce the observed lake variations (red curve, derived*
 390 *from lake level, lake area, forcing data and lake evaporation) and the sum of the glacier and land runoff*
 391 *we derive from remote sensing observations and modeling respectively (Sect. 3.2.). Error bars are*
 392 *associated to the glacier values and come from the geodetic results. Runoff values are expressed as*
 393 *heights scaled to the land surface of the Paiku catchment.*
 394
 395

396 Annual lake evaporation mainly ranges between 800 and 900 mm per year (Fig. 5B), with a mean
 397 value of $870 \pm 23\text{ mm}$ (1σ). Lake evaporation does not exhibit a linear trend of increase or decrease and
 398 is mostly dominated by year-to-year variability. Though slightly lower, our evaporation results are in

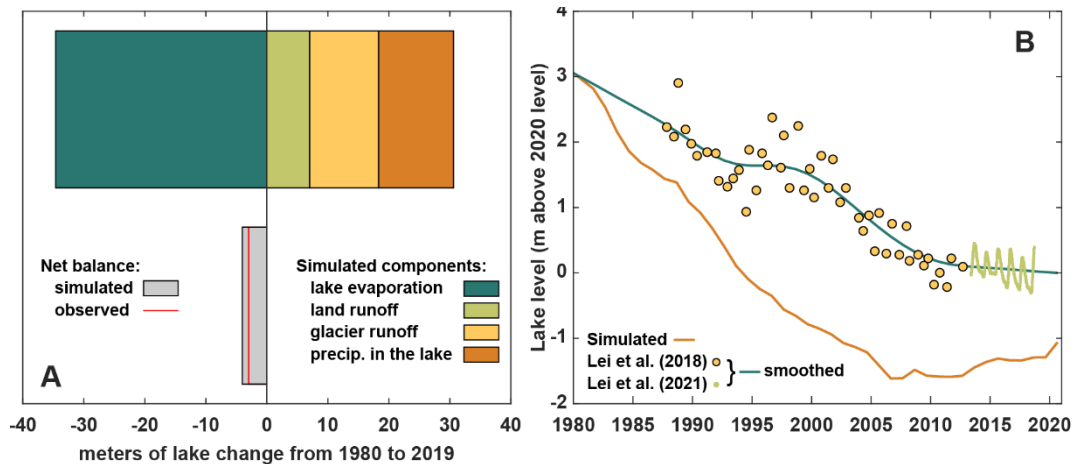
399 good agreement with the values from Lei et al. (2021), which are derived from local and regional
400 meteorological observation and lake budget calculation (Fig. 5B). We used the simulated evaporation
401 together with the lake level data and lake area data from Lei et al. (2018) and Lei et al. (2021) and the
402 precipitation forcing datasets (3.2.2) to derive the total runoff (land + glacier) required as an input to
403 the lake budget to reproduce the lake variations. This required runoff corresponds to the red line of
404 Fig. 5D. The required runoff volumes are scaled to the land area of the catchment to be comparable
405 with the other variables. Fig. 5D also presents the runoff values derived from the land cryo-hydrological
406 modeling and from the glacier remote sensing investigations. Annual volumes are expressed as mm
407 over the land part of the catchment (excluding the lake). As presented in Sect. 3.2.6., glacier mass
408 balance values are considered constant for the 1980-2000 period and the 2000-2019 period and are
409 respectively equal to $-4.6 \pm 2.5 \cdot 10^7$ and $-6.4 \pm 2.8 \cdot 10^7 \text{ m}^3$ per year. The addition of annual precipitation
410 to these values to quantify the total glacier runoff introduces year-to-year variability to the glacier
411 runoff. At the catchment scale, the average glacier runoff over the 40 years is 39 ± 13 mm per year.

412 Over the 40 years, the average annual land runoff value (surface + subsurface) we model is 24 ± 8
413 mm. Summed together, the land and glacier runoff find a partial agreement with the runoff that is
414 required to close the lake water balance. Annual values are compatible within error bars for 28 out of
415 the 40 years of simulations. The glacier and land runoff are slightly too small to close the lake water
416 balance during the first 20 years and slightly too large for the last 20 years of simulation. Over the whole
417 period, the sum of the glaciers + land runoff produces 95% of the required runoff. Land runoff is further
418 described in Sect. 4.3. and lake results in the following section.

419 *4.1.2. Hydrological budget of Lake Paiku*

420 Our observations, climate data, simulations, geodetic data and the lake level data from Lei et al.
421 (2018, 2021) enable us to quantify the different terms of the lake hydrological budget. We present these
422 results in m of lake level change based on the average slope of the Volume = f(level) relationship
423 (Fig. 6). As the unique output term, evaporation dominates the lake budget with an average annual value
424 of 0.86 m (34.6 m per 40 years, Fig. 6A). Direct precipitation in the lake is the dominant input with an
425 average annual value of 0.31 m (12.3 m per 40 years), followed by glacier runoff (0.28 m per year, 11.3

426 m per 40 years) and land runoff (0.18 m per year, 7.0 m per 40 years). When compared with lake volume
 427 observations over the 40 years of the simulation, the simulated lake budget is 1.04 m too negative.



428
 429 *Figure 6. Budget and level of lake Paiku for the simulation period (1980-2019). A. The different*
 430 *components of the hydrological budget of the lake according to our framework. Results are given in m*
 431 *of lake change based on the average slope of the $Volume = f(level)$ relationship. B. Lake level data.*
 432 *Points correspond to observations from Lei et al. (2018, 2021) that we smoothed (green curve, based*
 433 *also on observation points older than 1980). The simulated lake level appears in orange.*
 434

435 Based on our results, we also reconstructed lake level variations that we compare with the observed
 436 variations (Fig. 6B). Following our framework, our values are presented at an annual timestep. They
 437 qualitatively reproduce the overall lake level decrease but tend to overestimate this decrease and show
 438 an increasing mismatch with the observations from 0 in 1980 to 2 meters in 2005. This mismatch is
 439 later compensated by an increasing lake level trend in our simulation from 2005 to 2019. At the end of
 440 the simulation period, the mismatch is 1.04 m, consistent with the budget values (Fig. 6A) and the fact
 441 that our approach provides 95% of the required runoff to close the lake budget (Sect. 4.1.1.). This pattern
 442 of a too strong decrease followed by an increase is consistent with the comparison between simulated
 443 and required runoff presented on Fig. 5D.

444 4.2. Ground thermal results

445 Based on our temperature results, we define four categories of ground thermal regimes (Fig. 7A).
 446 *Cold permafrost* are the areas of the catchment for which the deepest thaw depth did not exceed 1 m
 447 over the 40 years of simulation. For cold permafrost, frozen conditions dominate the first meters of the
 448 ground most of the year and surficial thawing during summer can be interrupted by ground freezing

449 from the surface to the top of the permafrost at night. *Warm permafrost* are the areas of the catchment
 450 presenting permafrost for the whole duration of the simulation and which are not part of the *cold*
 451 *permafrost*. These areas are characterized by a distinct seasonal pattern of frozen ground in winter and
 452 an active layer in summer. *Disappearing permafrost* are the areas of the catchment presenting
 453 permafrost at the beginning of the simulation and not at the end. *No permafrost* are the areas without
 454 permafrost at the onset of the simulation. The geographical characteristics of each ground category are
 455 presented in Tab. 1, and their distribution throughout the catchment is shown on Fig. 7A. These different
 456 ground categories are subsequently used to compare their cryo-hydrological behaviors during the
 457 simulation (consistent color code).

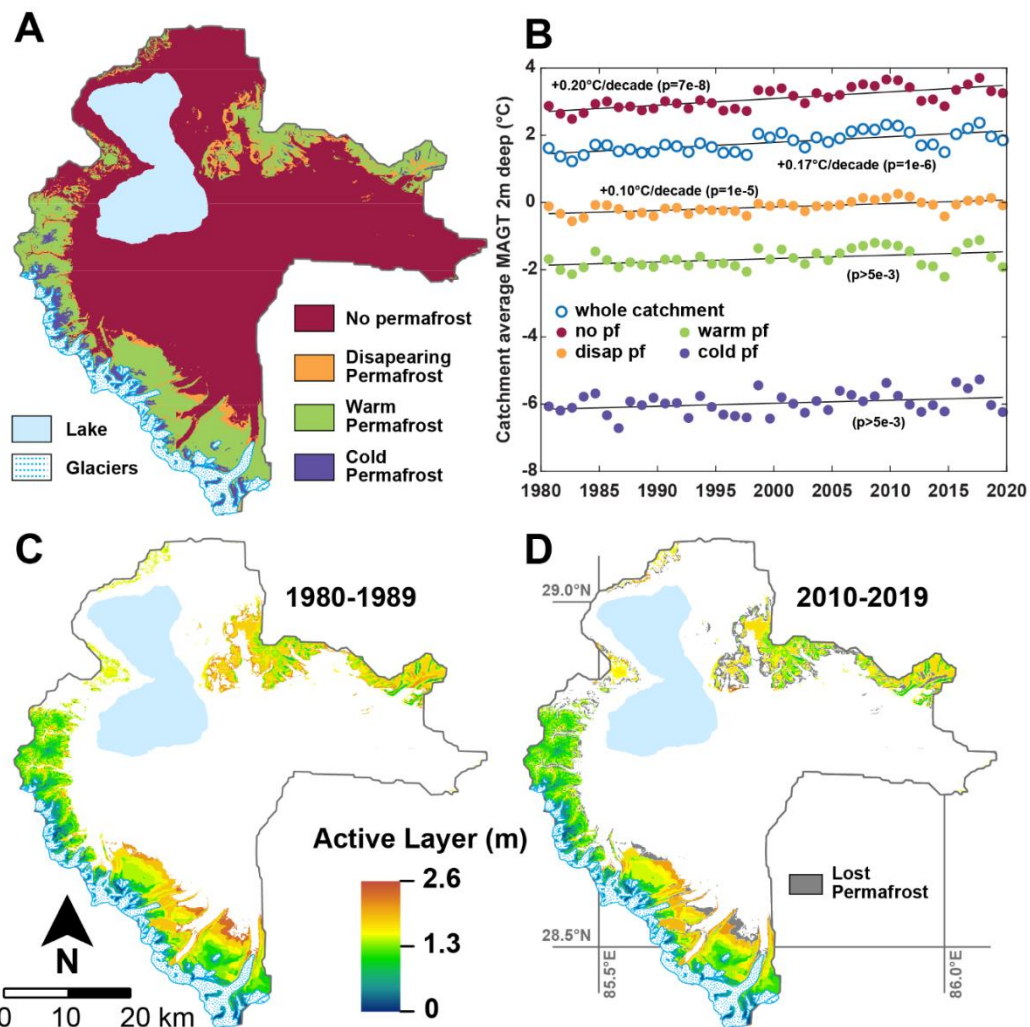
458 *Table 1. Cryological classification of the catchment based on the modeled ground temperatures.*

Name	Characteristics	% of the catchment area	Elevation mean (masl)	Elevation range (masl)	Slope mean (°)
Cold permafrost	Max thaw depth over the 40 years < 1m	3%	6068	6946 5213	35±13
Warm Permafrost	Max thaw depth > 1 m and permafrost present over the 40 years	19%	5480	5921 4877	20±9
Disappearing permafrost	Permafrost present in 1980 but disappears during the simulation	5%	5274	5552 4882	18±9
No permafrost	No permafrost from 1980 to 2019	73%	4900	5463 4580	10±8

459 At the catchment scale, the 2 m depth temperature (Fig. 7B) shows a pronounced warming trend
 460 of 0.17 °C per decade ($p=1\times 10^{-6}$). This trend is mainly supported by the *no permafrost* areas, which
 461 underwent a slightly stronger warming trend of 0.2 °C per decade ($p=7\times 10^{-8}$). Areas with disappearing
 462 permafrost, warm permafrost and cold permafrost exhibit smaller trends around 0.1 °C per decade with
 463 decreasing p-values (respectively 0.00001, 0.006 and 0.05, i.e. non-significant for the last two).

464 From 1980 to 1989, permafrost covers 27% of the catchment and the mean active layer thickness
 465 (ALT) is 1.36 ± 0.51 m (1σ , minimum: 0.11 m and maximum: 2.37 m, Fig. 7C). From 2010 to 2019,
 466 permafrost covers 22% of the catchment. At the scale of the initial permafrost area, this change
 467 corresponds to a loss of 19%. The mean ALT is 1.29 ± 0.49 m (1σ , minimum: 0.11 m and maximum:
 468 2.55 m, Fig. 7D) for this period. Permafrost disappearance (grey zones in Fig. 7D) mainly happens for

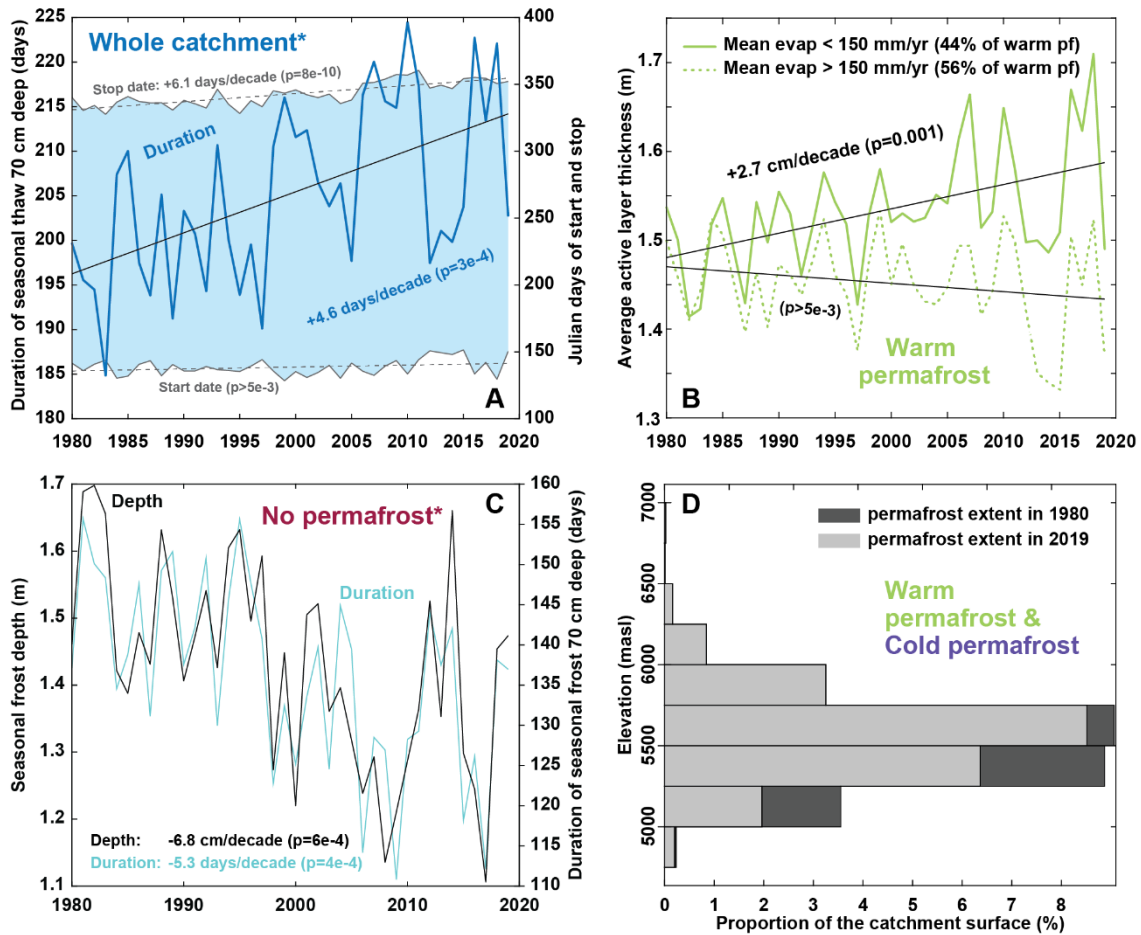
469 low-lying permafrost of the south and the center of the catchment. It occurs for the most part on the
 470 outer slopes of the permafrost regions and at the bottom of steep glacial valleys.



471
 472 *Figure 7. A: Different cryological states of the ground throughout the catchment for the 1980-2019*
 473 *period (see Tab. 1). B: Mean Annual Ground Temperature (MAGT) 2 m deep, averaged for the whole*
 474 *catchment and for the different cryological states of the ground. C: Average active layer depth over the*
 475 *1980-1989 period. D: Average active layer depth over the 2010-2019 period. Only locations presenting*
 476 *permafrost at the end of the simulation are assigned a color on the map on C and D. Locations where*
 477 *permafrost has disappeared are shown in gray on D.*
 478

479 We also present the average duration of seasonal thaw at a depth of 70 cm averaged over the
 480 catchment (Fig. 8A). Because at this depth some areas might present two (or more) consecutive years
 481 without thawing (highest locations) or without freezing (lowest locations), these areas were excluded
 482 from the averaging. In the end, the averaged results account for 89% of the catchment land area (i.e.
 483 excluding glaciers and lake Paiku). The results show an increasing trend in the duration of the seasonal
 484 thaw of +4.6 days per decade ($p=3 \times 10^{-4}$, blue line on Fig. 8A). When looking at the average start and
 485 stop days of the seasonal thaw (Fig. 8A, grey lines) in the Julian calendar (day 150 is the 30th of May

486 and day 300 is the 27th of October), we note that this increase is mainly caused by a later ending date of
 487 the thaw season (*Stop date* on Fig. 8A, +6.1 days per decade, $p=8\times 10^{-10}$) and not by an earlier starting
 488 date (non-significant trend).



489
 490 *Figure 8. A: Duration of seasonal thaw 70 cm deep averaged over the catchment. The asterisk indicates*
 491 *that the presented curves average 89% of the surface of the catchment (Sect. 4.2.). The gray curves and*
 492 *the light blue area are associated with the right axis and indicate the average start and stop day of the*
 493 *seasonal thaw in the Julian calendar. Values higher than 365 indicate that freezing conditions came*
 494 *back after the 31st of December. B: Active Layer Thickness (ALT) evolution for warm permafrost. The*
 495 *solid line shows the ALT for simulations experiencing an annual evaporation lower than 150 mm when*
 496 *averaged over the 40 years. The dashed line shows the ALT for simulations with annual evaporation*
 497 *higher than 150 mm. C: Temporal trends for seasonally frozen ground where there is no permafrost.*
 498 *The asterisk indicates that simulations were excluded if one of the simulated years did not present*
 499 *freezing conditions 70 cm deep (persistence of thawed conditions from one year to another). The*
 500 *presented curves thus average 88% of the total permafrost-free areas of the catchment. D: Altitudinal*
 501 *distribution of permafrost in 1980 and 2019. This distribution includes both cold and warm permafrost.*
 502

503 Within *warm permafrost*, we distinguished ALT for locations experiencing an average evaporation
 504 lower or higher than 150 mm per year during the simulations (Fig. 8B). Whereas locations with average
 505 evaporation below 150 mm per year record an active layer deepening trend of 2.7 cm per decade
 506 ($p=0.001$), it is not the case for locations with an average evaporation higher than 150 mm per year

507 (non-significative trend). This threshold value of 150 mm per year is based on further investigations on
508 the relationships between evaporation and ALT provided in Sect. 5.3.1.

509 In the permafrost-free areas of the catchment, seasonal frozen ground (Fig. 8C) reaches a depth of
510 1.43 ± 0.15 m on average and shows a decreasing trend of -6.8 cm per decade ($p=6 \times 10^{-4}$). At a 70 cm
511 depth, the average duration of seasonally frozen ground is 136 ± 12 days with a decreasing trend of -
512 5.3 days per decade ($p=4 \times 10^{-4}$). These values average 88% of the no permafrost areas since locations
513 showing persistent thawed conditions at this depth from one year to another were excluded (i.e. minimal
514 seasonal freezing depth over the 40 years lower than 70 cm).

515 When comparing permafrost spatial distribution between 1980 and 2019 (Fig. 8D), our results
516 show that permafrost distribution above 5750 masl has not been modified during the simulation.
517 Permafrost disappearance has mainly occurred between 5000 and 5750 masl, with the largest loss
518 reaching 2.5% of the catchment area between 5250 and 5500 masl.

519 4.3. Hydrological results for the land

520 The mean annual evaporation (land area only) over the simulation time is 180 ± 19 mm (1σ , Fig.
521 9A). Evaporation shows an increasing trend over the 40 years of $+1.01$ mm per decade ($p=3 \times 10^{-7}$).
522 Average total runoff over the 40 years is 24 ± 8 mm per year (Fig. 9B) and exhibits an increasing trend
523 of $+4.8$ mm per decade ($p=8 \times 10^{-7}$). Similarly, surface runoff (13 ± 3 mm per year) and subsurface runoff
524 (11 ± 6 mm per year) show increasing trends of $+1.3$ and $+3.5$ mm per decade ($p=6 \times 10^{-5}$ and 3×10^{-7})
525 respectively (Fig. 9B). The surface runoff presented on Fig. 9B includes the snow melt that did not
526 infiltrate the ground. These linear trends we report are high compared to the absolute values of the
527 variables and their extrapolation backward in time would lead to null values in the recent past which is
528 unrealistic. This suggests a non-linear evolution of these variables over the XXth century.

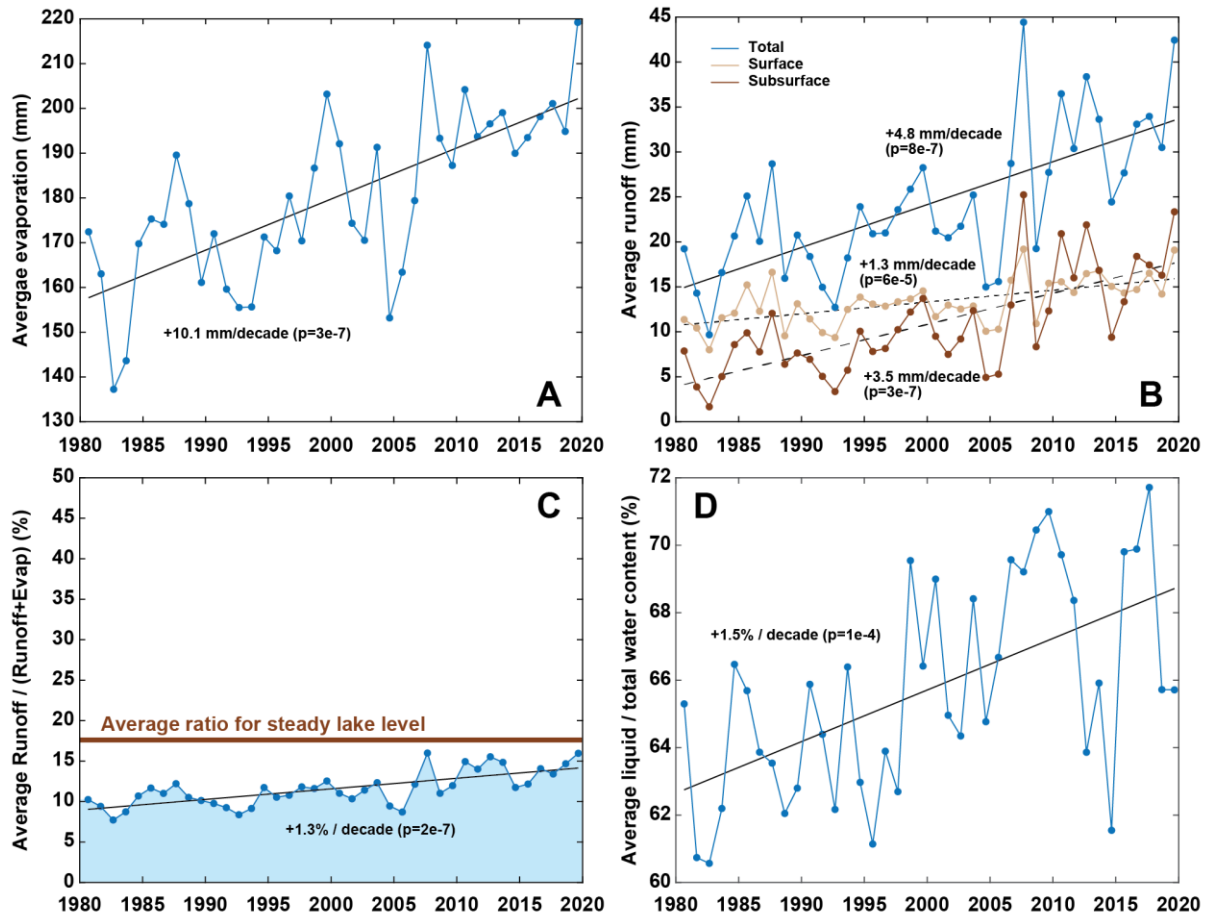
529 We also present the catchment average of the $runoff / (runoff + evaporation)$ ratio (Fig. 9C), which
530 is equivalent to $runoff / (rain + snow - snow\ sublimation)$ given the negligible contribution of soil
531 storage variations. Hence it is the proportion of the water input to the ground surface that is converted

532 into runoff. This proportion is $11 \pm 2\%$ over the simulation time and shows an increasing trend of
533 $+1.23\%$ per decade ($p=2 \times 10^{-7}$). Fig. 9C also shows the average theoretical ratio to maintain a steady
534 lake level (of 17.6%). This ratio was obtained under the following hypothesis:

- 535 • Same climate forcing data, hence same lake evaporation
- 536 • The glacier contribution is (i) considered the same for the historical simulation and this
537 scenario and (ii) taken as the difference between the total land surface runoff and the red
538 curve of *required runoff* in Fig. 5, therefore independent of remotely sensed estimates.
- 539 • Under these conditions, the runoff increase needed to maintain the lake level is only
540 supplied by land runoff (surface and subsurface) by shifting the *runoff / (runoff +*
541 *evaporation)* ratio.

542 The ratio from the historical simulation starts significantly below the theoretical steady lake ratio
543 ($10.2\% < 17.6\%$, Fig. 9C) and increases progressively to 16.0% in 2019.

544 Finally, Fig. 9D shows the annual proportion of *liquid / total* water averaged for the whole
545 catchment. The value was computed based on the daily water content (liquid and frozen) of the first
546 2 m of the soil column (the hydrologically active part of the column, Sect. 3.2.4.) from which annual
547 averages were derived and used to compute a catchment scale average. The graph shows that the
548 proportion of liquid water in the total water content increases at around $+1.41\%$ per decade ($p=1 \times 10^{-4}$),
549 indicating that water spends more and more time in the ground in a liquid form, being thus increasingly
550 available for hydrological processes such as evaporation or runoff.

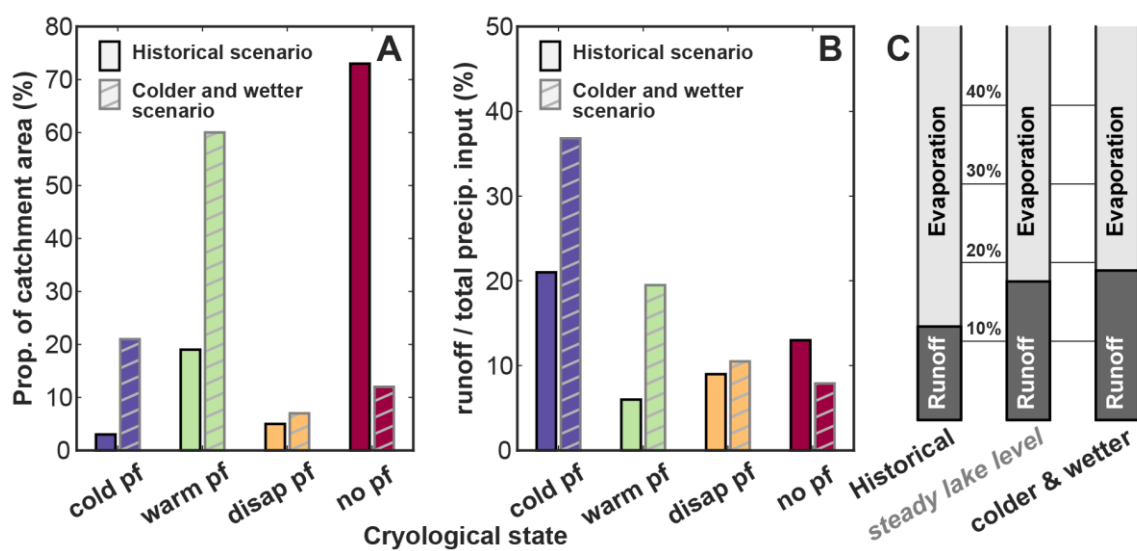


551
 552 *Figure 9. Hydrological results. A: Annual evaporation averaged over the whole catchment (land area*
 553 *only). B: Annual runoff averaged over the whole catchment (land area only). The blue curve sums the*
 554 *surface and subsurface runoff. C: Ratio between runoff and (evaporation + runoff) averaged over the*
 555 *whole catchment (land area only). The brown line indicates the theoretical average ratio needed to*
 556 *maintain a steady lake level when considering an identical glacier contribution to runoff (details in*
 557 *Sect. 4.3.). D: Annual mean of the (liquid water)/(total water) ratio over the first 2 meters of ground,*
 558 *averaged over the whole catchment (land area only).*

559 4.4. Sensitivity of evaporation and runoff

560 We conducted a simple sensitivity test on the climatic conditions (i.e. not a full-scale sensitivity
 561 test). We ran the same 40 years of simulations (with thermal initialization) for a climate 1 °C cooler and
 562 30% wetter (more precipitation) than the historical scenario. We call this new scenario *colder and wetter*
 563 (to be compared with the *historical scenario*, i.e. the results of the present study presented in the rest of
 564 Sect. 4.). Results of this experiment are presented in Fig. 10 and Table 2. Because of the difference in
 565 climate forcing, the *colder and wetter* scenario produced a greater amount of *cold* and *warm permafrost*
 566 areas than the historical scenario, as presented on Fig. 10A. Fig. 10B shows the proportion of the

567 precipitation reaching the surface (rain + snow – snow sublimation) that produces runoff compared to
 568 evaporation for the Paiku catchment. Fig. 10C aggregates over the whole catchment the distribution of
 569 such precipitation input to the ground between runoff and evaporation for both scenarios. In between
 570 them, it also includes the distribution associated with the steady lake level scenario of Fig. 9C, which
 571 is based on the hypothesis listed as bullet points in Sect. 4.3. (climate forcing of the historical scenario,
 572 same glacier contribution, only land runoff increases).



573
 574 *Figure 10. Sensitivity of the distribution between runoff and evaporation to climate. A: distribution of*
 575 *the different cryological states of the ground for the historical scenario (presented in Sect. 4.1. to Sect.*
 576 *4.3.) and for an alternative scenario where the climate is 1 °C colder and brings 30% more*
 577 *precipitation. B: runoff as a proportion of the precipitation input to the land (rainfall + snowfall – snow*
 578 *sublimation) for the different cryological states of the ground and for the 2 climatic scenarios. C:*
 579 *catchment scale ratio between runoff and evaporation for (i) the historical scenario, (ii) for a steady*
 580 *lake level with the same glacier contribution (same as Fig. 9C), and (iii) for the colder and wetter*
 581 *scenario.*

582
 583 The *historical scenario* shows that *cold permafrost* areas produce the highest proportion of runoff,
 584 which we attribute to the fact that the ground in these areas is most of the time frozen, turning a
 585 substantial part of the snow melt and rainfall into surface runoff. When considering grounds with a
 586 hydrologically active subsurface (*warm permafrost*, *disappearing permafrost* and *no permafrost*) in the
 587 historical scenario, the proportion of runoff increases slightly from *warm permafrost* to *no permafrost*.
 588 Such an evolution then corroborates the idea that the presence of permafrost tends to increase
 589 evaporation at the expense of runoff, as modeled by Sjöberg et al. (2021). Yet, for the *colder and wetter*
 590 scenario, runoff shows a regular decrease from *cold* to *no permafrost* with a more pronounced trend
 591 than the historical scenario. Several factors can be at play in this transition and most likely involve (i)

592 a different extent and altitudinal distribution for each cryological type of ground, (ii) an overall reduced
 593 intensity of evaporation due to cooler surface temperatures, (iii) a higher soil water content driven by
 594 higher precipitation and (iv) difference in the seasonal timings. Altogether, these processes substantially
 595 change the proportion of water that ends up as runoff water available for the lake, as highlighted by Fig.
 596 10C.

597 *Table 2. Distribution between runoff and evaporation for the 2 scenarios*

Ground cryological type	<i>Historical Scenario</i>			<i>Colder and wetter scenario</i>		
	Precipitation input ¹	Runoff	Evaporation	Precipitation input ¹	Runoff	Evaporation
Cold permafrost	100% 117 mm	21% 24 mm	79% 93 mm	100% 234 mm	37% 86 mm	63% 148 mm
Warm permafrost	100% 183 mm	6% 10 mm	94% 173 mm	100% 281 mm	20% 55 mm	80% 226 mm
Disappearing permafrost	100% 211 mm	9% 19 mm	91% 192 mm	100% 211 mm	10% 22 mm	90% 189 mm
No permafrost	100% 218 mm	13% 28 mm	87% 189 m	100% 200 mm	8% 16 mm	92% 184 mm

598 ¹. Precipitation input is the input to the ground, counted as rainfall + snowfall – snow sublimation

599 **5. Discussion**

600 5.1. Limitation and potential of the approach

601 *5.1.1. Data usage within the conceptual framework and data scarcity*

602 Our approach relies on a variety of data regarding their scientific focus (glaciers, ground, lake,
603 atmosphere), their type (in situ observations, remotely sensed data, reanalysis data), their characteristics
604 (point wise data, distributed data, constant or with various time resolution) and the way they interact
605 with our models (model parameters, forcing data, validation data, result data in case of the glacier
606 runoff). Such a diversity arises from our goal to quantify both the ground thermo-hydrological regime
607 and the different terms of the lake budget. This variety also makes it challenging to consistently merge
608 these data into a unique framework. For example, our quantification of the glacier mass change
609 reconstruction is made of two constant values for the study period (1975-2000 and 2000-2020), which
610 limits the relevance of the comparison between the observed lake level variations and the simulated
611 ones.

612 Yet, the lake level variations are the only hydrological observations available to evaluate the
613 robustness of the runoff we compute. Therefore, we had to combine lake level observations with our
614 precipitation forcing data and lake evaporation quantifications in a simple mass conservation
615 calculation, to derive the land runoff to the lake required to reproduce the level variations (red curve on
616 Fig. 5D). In this regard, the sum of the glacier and land runoff we derive over the 40 years correspond
617 to 95% of the required runoff to the lake, indicating that the magnitude of our reconstruction is correct.
618 Year-to-year comparison is less accurate and we suggest that this is the consequence of the
619 aforementioned limitations and also of our modeling strategy as detailed below.

620 A main limitation regarding our usage of the data is related to the limited amount of available field
621 observations required to provide robust model parameterizing, climate forcing and in-depth validation
622 of the simulations, both hydrologically and thermally. Regarding climatic forcing data, our AWS
623 measurement offers sound observations to evaluate and adjust the ERA5 data processed with TopoSUB
624 and downscaled with TopoSCALE. Yet, a period of observations longer than 2 years would have
625 enabled more robust corrections and could have allowed us to perform a more advanced statistical

626 downscaling approach, e.g. quantile mapping (Thiemeßl et al., 2011). As such, the spatiotemporal
627 domain of relevance of these corrections is insufficient to correct data for the whole catchment and the
628 40 years of simulations. Overall, considering the strong bias we observe in the raw ERA5 data (Fig.
629 D1), these corrections do represent an important first-order improvement.

630 Additionally, in absence of borehole data that would allow us to anchor our parameters into
631 observations, we rely on gridded values designed for hydrological and/or land surface modeling (Sect.
632 3.2.4. and Appendix A). Because these values might be less reliable than field observations, we chose
633 to average them over the catchment to derive some more robust values. Altogether, this scarcity of field
634 observations is likely to bring significant uncertainties to our analysis. Future efforts should focus on
635 acquiring additional data or developing validation methods based on remotely sensed observations.

636 *5.1.2. Modeling strategy*

637 A limitation in our study is that lateral water flows between land simulation units is ignored. By
638 giving access to the timing of water transport across the catchment, water routing would allow to
639 investigate temporal hydrological patterns at a monthly or seasonal scale. Because we work at annual
640 and decadal time scales, this limitation has limited consequences on our results. The main consequence
641 is to ignore potential storage effects on the land that would delay the arrival of runoff to the lake. We
642 suggest that it is possible that this limitation partly explains the limited match between computed and
643 required runoff at the annual time scale (Fig. 5). Yet, our subdivision of the catchment based on the
644 different cryological states of the ground allows us to work with hydrological units that are smaller than
645 the catchment and thus present shorter hydrological response time to precipitation.

646 Additionally, our approach regarding the modeling of runoff is relatively simple, i.e. partition
647 between subsurface and surface runoff based on comparison between the soil water content and field
648 capacity and porosity, respectively. More complex approaches split runoff into more sophisticated
649 categories such as Horton overland flow, Dunne overland flow, subsurface stormflow... (e.g. Savenije,
650 2010; Gao et al., 2014; Mirus and Loague, 2013). However, over the last decade, the relevance of this
651 type of partitioning between different types of runoff has been questioned (McDonnell, 2013; Gao et
652 al., 2023). In the frame of our study, we find it important to distinguish between surface and subsurface

653 runoff because they generate flows with very contrasted speed. In a general perspective, this significant
654 difference in flow velocities impacts the hydrological system as a whole (e.g. river discharge,
655 evaporation...) and has various consequences throughout the catchment, such as the water availability
656 for vegetation, erosion and sediment transport.

657 In the particular case of a cryo-hydrological study, separating surface from subsurface runoff is
658 particularly relevant because both flows do not react in the same way to ground temperature changes.
659 As such, we see our approach as a middle way that allows us to make this distinction based on simple
660 hydrological considerations. Yet, we acknowledge that the classification and quantification of the
661 different types of runoff represent a valuable direction for future investigation on catchment-scale cryo-
662 hydrology in Tibet. Another potential improvement in our modeling approach could be to unravel
663 evaporation from transpiration. However, since vegetation is extremely scarce in the Paiku catchment,
664 which is largely dominated by barren lands, we suggest that this would not significantly affect our
665 results. However, this limitation should be explored in future field and modeling studies.

666 Conversely, our approach also conveys several important advantages regarding our goal to
667 describe and quantify the ground thermo-hydrological regime of the catchment. The use of TopoSUB
668 enables us to produce results at a resolution of 100 x 100 m over an area of nearly 2400 km² with
669 calculation costs 700 times lower than if each 100 x 100 m pixel was treated individually. Yet, thanks
670 to the clustering method used to produce the forcing dataset (Sect. 3.2.2.), the strong spatial variability
671 of the physiography and its impact on the climate and incoming radiations is significant in the forcing
672 data and has a major influence on the ground thermo-hydrological results, as exemplified by the strong
673 spatial variability of ground temperatures (Fig. 7). Beyond elevation, other physiographic parameters
674 such as aspect also influence the results. The mean values of 2 m-deep temperature and evaporation
675 over the 40 years for north-facing areas (averaged over the whole catchment and over the 40 years) are
676 1.3 °C and 163 mm while they reach 2.9 °C and 197 mm for the south-facing ones. This strong
677 dependence of modeled results on physiography highlights the necessity to take it into account when
678 modeling the thermo-hydrological regime of the ground in high mountainous environments. Finally,
679 our approach allows us to couple the physical processes governing both energy and water fluxes at the
680 surface and subsurface and highlight their interplay, as developed in Sect. 5.3.1.

681 5.2. Trends in the catchment and across the QTP

682 5.2.1. *Lake hydrological budget and level variations*

683 The total lake level change we simulate is a decrease of 4.11 m. This is qualitatively consistent
684 with the overall observed trend. The mismatch with the observations is limited to a 1.04 m excess in
685 the simulated level drop (Fig. 9A). Our reconstruction shows a decrease of 4.66 m from 1980 to 2007,
686 which is an overestimation of the initial drop. Afterwards, while observations indicate a gradual
687 slowdown of the lake level decrease, we simulate a stabilization followed by a slight increase (0.55 up
688 between 2013 and 2019). The reason for the overall mismatch of 1.04 m can arise from bias (i) in the
689 forcing data (and mainly in the precipitation) used for the land and lake simulations, (ii) in the glacier
690 mass balance estimate and/or (iii) in the quantification of hydrological processes for the land or for the
691 lake (evaporation, runoff). On top of these potential biases, the difference in trends for the end of the
692 simulation time can be influenced by (i) our estimates of glacier mass changes, which are made of two
693 time averages (one for the 1980-2000 period and one for the 2000-2020 period) and therefore produce
694 very smoothed glacier runoff values that cannot capture variations at the scale of the decade of less and
695 (ii) the absence of water routing that prevent us from accounting for delays of storage effects on the
696 water supply from the land to the lake.

697 Additionally, our approach ignores potential water fluxes between the lake and a surrounding
698 aquifer. This can be a possible reason for this mismatch. In the context of a decreasing lake level, an
699 aquifer surrounding the lake can create an additional water inflow when the lake level passes below the
700 piezometric level of the aquifer (Yechieli et al., 1995). We suggest that such an inflow could mitigate
701 the lake level decrease and thus explain the missing water in our reconstruction (Fig. 6B). It could also
702 explain the gradual stabilization of the lake level that our model does not reproduce. This flow is not
703 part of our conceptual hydrological framework even though it likely exists in reality, especially since
704 there is no permafrost near the lake (as we simulate it here), allowing for the existence of such an aquifer
705 (Walvoord and Kurylyk, 2016). Groundwater has been identified as a potential contributor to lake level
706 rise in other regions of the QTP (Lei et al., 2022). In the long run, lake-aquifer systems commonly
707 follow oscillations of the net atmospheric flux of water (Precipitation – Evaporation) and of the runoff

708 that forces its mass balance (Watras et al., 2014). During these oscillations, the lake can “pump” water
709 from the aquifer or feed it depending on the relative difference of piezometric level between them
710 (Almendinger, 1990; Liefert et al., 2018). Yet, this potential effect is difficult to account for and its
711 magnitude remains unclear. Therefore, the reasons for the mismatch between observed and simulated
712 lake levels could also be connected to other aspects of our methodology such as bias in the climatic
713 forcing data and other shortcomings arising from the lack of field data, or hydrological processes, as
714 developed in Sect. 5.1.1. and 5.1.2.

715 Our reconstruction of the lake budget is informative regarding the respective contribution of the
716 different inputs and outputs. Regarding lake evaporation, our mean value of 870 ± 23 mm is close to
717 the one modelled by Yang et al. (2016) with the Flake model for lake Nam (832 ± 69 mm) for the period
718 1980-2014 but we do not report a significant increasing trend in our results. Yet for the same lake (Nam
719 Co) and a similar period (1980-2016) Zhong et al., (2020) reported an average value of 1149 ± 71 mm
720 (along with an increasing temporal trend) using the Penman formula (Penman, 1948), thus highlighting
721 the potential dependence of the results to the methodology. In our results, direct precipitation to the lake
722 represents 40% of the inputs, followed by glacial runoff (35%) and land runoff (25%). Glaciers are
723 therefore a particularly important contributor to the runoff towards the lake (60% of the total runoff, vs.
724 40% for land runoff), what contrasts with the results from Biskop et al. (2016) who calculated that the
725 runoff input to the lake Paiku was dominated by land runoff (70% vs. 30% for the glacier contribution).
726 Here again, these differences likely arises from important differences in input data and methodologies
727 to quantify the different hydrological processes (evaporation, runoff, snow and glacier melt). Yao et al.
728 (2018) reported that, at the QTP scale, the balance between precipitation and evaporation (over land
729 and lake) was dominant over glacier melt to understand both lake storage increases and decreases. Our
730 reconstruction does not give us access to significant temporal variation of the glacier contribution but
731 the above-mentioned proportions in the contributions to the lake (40%, 35% and 25%) show that the
732 glacier contribution does not dominate the input terms. At the catchment scale, these proportions can
733 vary significantly depending on the glacier coverage. For Lake Selin, Zhou et al. (2015) reported that
734 runoff towards the lake, evaporation from the lake and on-lake precipitation altogether explained 90%

735 of the lake storage variations for the 2003-2012 period. The catchment of lake Selin has a very limited
736 glacier coverage, corresponding to 0,63% of its area (Lei et al., 2013), compared to the Paiku (5%).

737 *5.2.2. Permafrost and ground temperature trends*

738 Our results indicate that permafrost coverage in the Paiku catchment evolved from 27 to 22% of
739 the land area during the simulated period. Such a coverage corresponds to sporadic permafrost (10-50%
740 of the area) and is consistent with recent large-scale estimates of permafrost in the Northern Hemisphere
741 (Obu et al., 2019) and across the QTP (Zou et al., 2017; Ran et al., 2018). This decrease corresponds to
742 a 19% shrinkage of the 1980 permafrost area, which is higher than the 9% reported by Gao et al. (2018),
743 a value determined by catchment-scale numerical modeling in the upper Heihe catchment (northeastern
744 QTP) over a similar period. It is also slightly higher than the 13% decrease modeled from 1971 to 2015
745 for the Qinghai Lake catchment with a similar approach by Wang and Gao (2022). Yet, it is smaller
746 than the 34% loss modeled by Qin et al. (2017) from 1981 to 2015 for the Yellow River Source Region
747 (YRSR, North Eastern QTP).

748 Active layer (AL) evolution is contrasting throughout the catchment and a deepening signal is only
749 visible for the locations with limited evaporation (<150 mm per year). Given the strong drive of summer
750 climate on ALT, this overall lack of a deepening trend highlights how evaporation can act as an energy
751 intake at the surface (Yang et al., 2014a), limiting the surface and subsurface heat fluxes and thus AL
752 deepening. In this regard, our results fall in line with the conclusions of Fisher et al. (2016) when
753 observing evapotranspiration and ALTs in boreal forests and also confirm the modeling experiments of
754 Zhang et al. (2021b) on permafrost wetting in arid regions of the QTP. Besides, the lack of an overall
755 deepening trend is consistent with observations from Luo et al. (2018) in the YRSR over the last decade
756 and with the modeled AL from Zhang et al. (2019) at the scale of the QTP for the last 40 years. Where
757 evaporation is limited, we report an AL deepening trend of 2.7 cm per decade, which is smaller than
758 the 4.8 cm per decade trend modeled by Song et al. (2020) for the YRSR for the same period, and
759 smaller than the 4.3 cm modeled by Gao et al (2018) in the upper Heihe catchment. Yet it is comparable
760 to the 2 cm per decade value modeled by Wang and Gao (2022) for the Qinghai Lake catchment from
761 1971 to 2015. Connection between AL deepening and evaporation are discussed in Sect. 5.3.1.

762 In *no permafrost* areas, our simulations show that the thickness of seasonally frozen ground shrinks
763 at a rate of 6.8 cm per decade. This rate is faster than the rate of 3.1 cm per decade quantified by Qin et
764 al. (2018) using the Stefan solution for the YRSR (1961-2016) and faster than the 3.2 cm per decade
765 modeled by Gao et al. (2018, Heihe catchment). However, it is similar to the 6 cm per decade rate
766 modeled by Wang and Gao (2022) in the Qinghai Lake catchment from 1971 to 2015 and smaller than
767 the 12 cm per decade modeled by Qin et al. (2017) for the YRSR (1981-2015). All these values fall
768 within the wide range of 3 to 29 cm per decade reported by Wang et al. (2020a) when studying
769 seasonally frozen ground over the whole QTP with in-situ observations. Regarding timing, we report a
770 decreasing trend of 5.3 days of frozen conditions (70 cm deep) per decade which is consistent with the
771 decrease of 6.7 days per decade reported by Wang et al. (2020a) just below the surface.

772 Regarding the timing of seasonal ground thaw, our results highlight that the increase in the duration
773 in the seasonal ground thaw (at 70 cm) is mostly driven by a progressive delay of the end date of the
774 thaw period. This result contrasts with those from Song et al. (2020) for the same period in the YRSR
775 who also modeled an increase of the seasonal thaw (at a 2 cm depth), although driven by an advancing
776 trend of the start date of the seasonal thaw.

777 Our warming trends at a 4 m depth for permafrost areas is 0.1 °C per decade, which is substantially
778 smaller than the 0.43 °C per decade observed at this depth between 1996 and 2006 in permafrost
779 boreholes along the Qinghai-Tibetan Highway in the North East of the QTP (Wu and Zhang, 2008).
780 Zhang et al. (2019) reported a 0.13 °C per decade of warming of the permafrost top during winter that
781 is consistent with the trend of 0.14 °C per decade we observe at 2 m depth (mean AL between 1.4 and
782 1.7 m in our simulations) for the months of December, January and February.

783 5.2.3. *Evaporation and runoff trends*

784 Our results are characterized by (i) an increase of both evaporation and runoff (Fig. 9A and 9B),
785 mainly driven by an increase in precipitation (Fig. 3 bottom), (ii) a runoff/(runoff+evaporation) ratio
786 exhibiting an increasing trend as a result of ground warming and permafrost disappearance that both
787 enable more subsurface runoff along time (Fig. 9C and 10D) and (iii) an increase in the proportion of
788 liquid water in the ground compared to ice (Fig. 9D). Regarding all these points, our results find a good

789 consistency with the evolution reported by Gao et al. (2018) for the upper Heihe catchment
790 (northeastern QTP) using a similar approach for a comparable period (1971-2013). The increasing
791 trends in evaporation and runoff they report for the thawing season (dominant period for both processes)
792 are comparable with the annual values we report: +10.0 mm per decade for evaporation (our study:
793 +10.1 mm per decade) and +3.3 mm per decade for runoff (our study: +4.8 mm per decade). Similar
794 evolutions are also reported by Wang and Gao (2022) for the Qinghai Lake catchment and by Qin et al.
795 (2017) for the YRSR (1981-2015). These increases in runoff (especially surface runoff) are likely to
796 have an influence on sediment transport. For instance, Li et al. (2021) showed that current precipitation
797 increase over High Mountain Asia is driving a runoff increase, which contributes to a significant rise in
798 fluvial sediment fluxes. Regarding differences, Qin et al. (2017) modeled a stronger evaporation
799 increase (14.3 mm per decade) linked to a decreasing runoff coefficient. Similar to Li et al. (2019), we
800 see that an important part of snow melt (49%) infiltrates the ground and later contributes to runoff and
801 evaporation.

802 5.3. Cryo-hydrological couplings at catchment scale and implication for 803 lake level variations

804 5.3.1. *Interdependence of thermal and hydrological variables*

805 Our simulation results enable us to explore the interplay between the fluxes of energy and water at
806 the surface and subsurface. In this regard, we tested the correlation of evaporation with the proportion
807 of liquid/total water in the ground for cold and warm permafrost, as well as the correlation between
808 evaporation and the duration of seasonal thaw at a 70 cm depth (Fig. 11, A and B). For permafrost areas
809 (*cold permafrost* and *warm permafrost*), evaporation shows a strong correlation with the seasonal
810 distribution between liquid and frozen water, similar to previous modeling works for the region (Cuo
811 et al., 2015). As such, this correlation suggests that the intensity of seasonal ground thaw plays a role
812 in enabling higher or lower evaporative fluxes. This is likely due to cold surface temperatures strongly
813 reducing water loss from the surface and because moisture delivery to the surface is inhibited when the

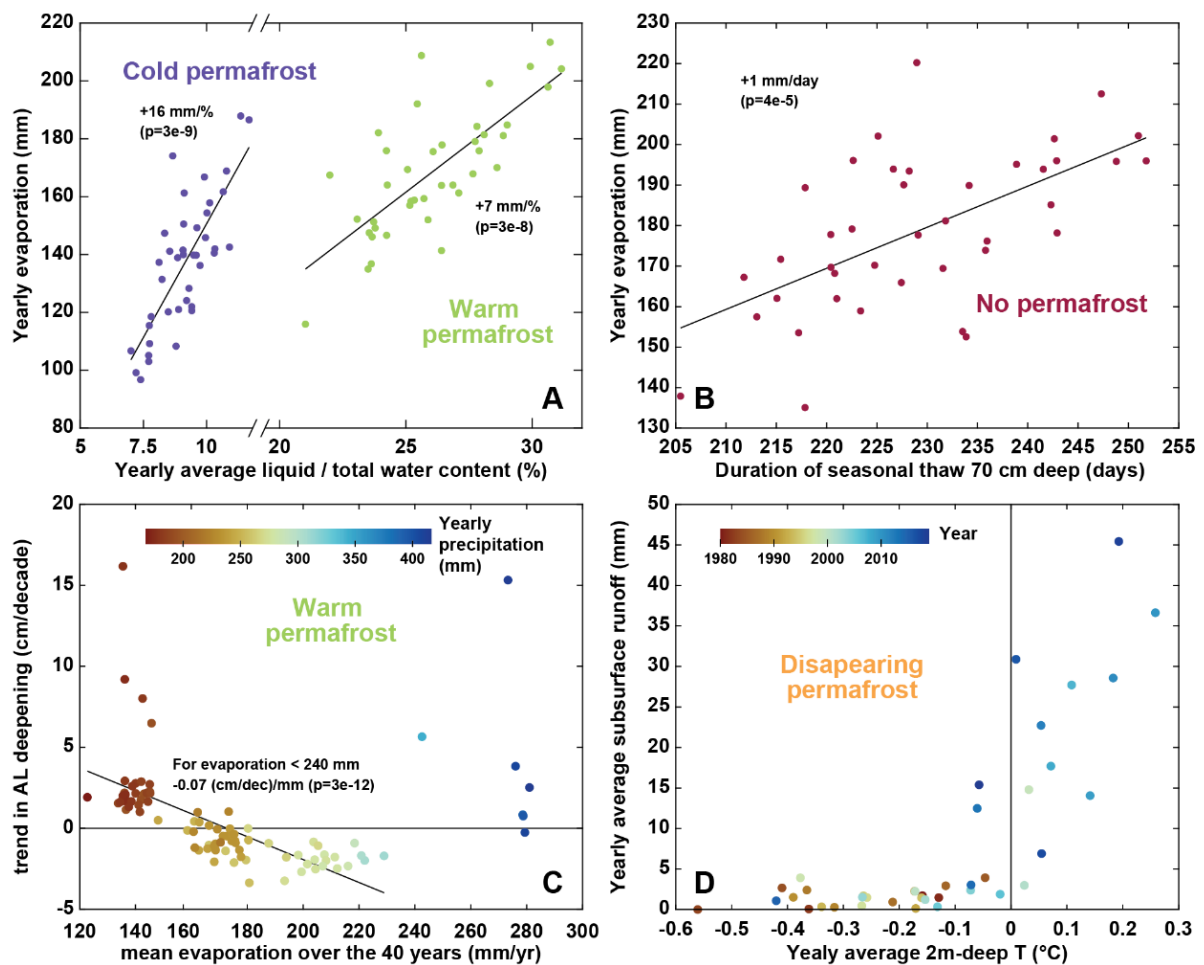
814 ground is frozen. We suggest that this dependence is particularly important in the Paiku Catchment
815 because evaporation is strong (88% of the precipitation input to the surface evaporates on average) and
816 because frozen water is the dominant form of water in the ground in permafrost areas (Fig. 11A, the
817 calculation includes the first 2 meters below the surface).

818 Similarly, evaporation in *no permafrost* areas shows a significant correlation with the duration of
819 the seasonal thaw (Fig. 11B). We suggest that this result arises from the fact that frozen ground limits
820 the evaporative fluxes and thus years during which the subsurface seasonal thaw is shorter are associated
821 with reduced evaporative fluxes. We also tested the relationship between the linear trend of active layer
822 deepening and the mean evaporation (over the 40 years of simulation) for *warm permafrost* areas (Fig.
823 11C). Thus, this graph does not present annual values and one point corresponds to one of the 92
824 TopoSUB points classified as *warm permafrost* (values averaging the 40 years). The graph highlights
825 that TopoSUB points showing an Active Layer (AL) deepening trend are associated with low
826 evaporation and precipitation. From there, TopoSUB points with stronger evaporation show no
827 deepening trend or even a shrinkage of the AL. This relationship is contradicted by the highest level of
828 evaporation (>240 mm per year) observed for *warm permafrost*, for which AL deepening is observed
829 again (dark blue points of the graph). These TopoSUB points with the highest levels of evaporation also
830 correspond to those receiving the largest amount of precipitation.

831 Runoff also shows a strong connection with the ground thermal regime (Fig. 11D). At the
832 beginning of the simulation, years with an average 2 m-deep temperature below 0 °C are associated
833 with limited subsurface runoff (< 5 mm per year). Over the years, as the ground warms up and
834 permafrost disappears, subsurface runoff increases and can reach 20 to 45 mm per year. This result is
835 consistent with increased subsurface connectivity expected when permafrost thaws (Kurylyk et al.,
836 2014; Gao et al., 2021) that has been both observed (Niu et al., 2016) and modeled (Lamontagne-Hallé
837 et al., 2018; Huang et al., 2020; Gao et al., 2018). We suggest that these substantial changes in
838 subsurface runoff, associated with changes in the ground temperature in Fig. 11D support the hypothesis
839 of a modification in the hydrological pathways as permafrost thaws.

840 Altogether, these results suggest a dependence of key variables quantifying the catchment
841 hydrological balance (evaporation, runoff) to the seasonal characteristics and interannual trends of the

842 ground thermal regime (temperature, liquid vs frozen water content). Similar to previous studies (Ding
 843 et al., 2020; Wang and Gao, 2022), we think these results advocate for the necessity to couple thermal
 844 and hydrological modeling to improve our ability to understand and quantify changes in the
 845 hydrological balance of high mountain catchments. To our best knowledge, along with Gao et al. (2022),
 846 our study represents to date the most complete effort to include the variety of coupled climatological,
 847 surface and subsurface processes characterizing the climate, hydrology and ground thermal regime of
 848 high-mountain catchments in Tibet at a small scale with a high spatial resolution.



849
 850 *Figure 11. Thermo-hydrological couplings. A: Annual evaporation vs. annual mean of the liquid / total*
 851 *water ratio over the first 2 meters of ground, averaged for simulations corresponding to cold permafrost*
 852 *and warm permafrost (one dot per year for each permafrost category). B: Annual evaporation vs.*
 853 *duration of seasonal thaw at a 70 cm depth averaged for simulations corresponding to locations without*
 854 *permafrost (one dot per year). C: Active layer deepening trend vs. mean evaporation over the 40-year*
 855 *for each simulation corresponding to warm permafrost (here one dot corresponds to one TopoSUB*
 856 *point). The color of the dots shows the precipitations averaged over the 40 years for each simulation.*
 857 *The linear regression excludes simulations exhibiting annual evaporation higher than 240 mm. D:*
 858 *Annual subsurface runoff vs Annual 2 m-deep temperature averaged for simulations corresponding to*
 859 *locations with disappearing permafrost (one dot per year). The color of the dot indicates the year of*
 860 *the simulation.*

861 5.3.2. *Influence of the ground thermal regime on the distribution between runoff*
862 *and evaporation*

863 Our results indicate that evaporation is particularly strong in the Paiku catchment. Over the 40
864 years of simulation, 10% of the total precipitation is converted into runoff, and the rest of the water is
865 either directly returned to the atmosphere from the snowpack via snow sublimation or from the ground
866 surface via evaporation. Comparatively, Gao et al. (2018) observed and modeled a ratio of around 35%
867 for the Heihe catchment; Qin et al. (2017) reported an average ratio of 33% for the YRSR and Li et al.
868 (2014), a ratio of 83% for the Qugaqie catchment (central QTP) but modeling hydrological fluxes only.

869 Our sensitivity test on evaporation and runoff for a slightly different climates (Sect. 4.4.) highlights
870 the fact that the role of permafrost regarding the runoff/evaporation distribution is a complex question,
871 as it has already been discussed in the literature (e.g. Bring et al., 2016). Some studies have suggested
872 that landscape-scale permafrost thaw would trigger more evaporation (Walvoord and Kurylyk, 2016).
873 This phenomenon was modeled by Wang et al. (2018) in the upper Heihe River Catchment, for which
874 they reported that the thickening of the active layer increased the ground storage capacity and led to a
875 decrease in runoff and an increase in evapotranspiration. Wang et al. (2020b) also reported that
876 permafrost thawing accelerated evapotranspiration (1961-2014).

877 Conversely, Zhang et al. (2003) and Carey and Woo (1999) reported that shallow frozen ground
878 conditions (such as a shallow active layer) maintain higher water contents close to the surface,
879 promoting higher evaporation. Sjöberg et al. (2021) modeled this phenomenon with a fully coupled
880 cryo-hydrological model including surface energy balance calculation. They modeled a slope with a
881 simplified geometry in 2D for different permafrost coverages. They found that hillslopes with
882 continuous permafrost have twice as high rates of evapotranspiration compared to hillslopes with no
883 permafrost.

884 As such, the interplay between the runoff/evaporation distribution and the ground thermal regime
885 in areas where permafrost coverage shows a spatiotemporal variability is difficult to apprehend (Fig
886 10). This complexity is most likely due to a strong sensitivity to the drainage conditions (fast flows of
887 steep mountain environments vs. slow flows of lowland catchments) and to the climate setting, both at

888 the annual scale (arid regions vs. wet regions) and at the seasonal time scale (relative timing of
889 temperature variations, rainfall, snowfall, snow melt and ground freeze/thaw).

890 Because it can both promote evaporation or runoff depending on the setting, the ground thermal
891 regime of the catchment seems to have the possibility to create a positive feedback, both towards lake
892 level decrease or increase. Further studies should therefore focus on comparing the thermo-hydrological
893 regime of different Tibetan catchments with contrasting lake level changes and permafrost coverage, to
894 test to which extent these differences can contribute to explain the spatial patterns of lake level changes
895 across the QTP.

896 5.4. Implications for lake level changes

897 At the scale of the Paiku catchment and in regard of lake level variations, the results we present
898 highlight that:

- 899 • The sum of the direct precipitation in the lake, the land runoff and the glacier runoff are not
900 enough to compensate for the lake evaporation over the study period, hence driving the
901 observed lake level decrease.
- 902 • Long-term hydrological trends in the catchment are led by trends in climate; and
903 precipitation increase, jointly with glacier melt, provides enough water to drive a
904 concomitant increase of runoff and evaporation.
- 905 • Ground thermal changes increase the distribution of liquid vs. frozen water in the ground
906 and the duration of seasonal thaw, correlations suggest that these modifications increase
907 evaporation. The warming of the ground is also related to the increase of subsurface runoff
908 towards the lake.
- 909 • Ground warming and permafrost thawing promote subsurface runoff over time,
910 contributing to an increase in the runoff/evaporation ratio of the catchment.
- 911 • Over the 40 years we studied, the presence of permafrost seems to promote evaporation at
912 the expense of runoff. Yet this trend appears to be climate-dependent and the cryological

913 state of the ground might shift the runoff/evaporation distribution in the other direction
914 under colder and wetter climates.

915 At the scale of the QTP, these results have several implications. First, a better understanding of the
916 recent and future lake level variations will come with a better knowledge of spatial patterns and
917 temporal trends in precipitation. Second, climate changes are modifying the ground thermal regime of
918 Tibetan catchments. Ground warming may lead to active layer deepening, permafrost disappearance
919 and/or changes in the seasonal freeze/thaw cycles, affecting evaporation, runoff volumes and pathways
920 and overall, changing the hydrological functioning of Tibetan catchments (and the waterflow provided
921 to the lakes). Finally, the effect of permafrost on the distribution between evaporation and runoff seems
922 to be dependent on the climate settings and the permafrost coverage of the catchment. Further studies
923 should investigate this phenomenon and how it might contribute to explaining the contrasting lake level
924 evolutions across the QTP.

925 **6. Conclusion**

926 Our study quantifies the different terms of the Lake Paiku budget over the past 40 years. Direct
927 precipitation to the lake represents 40% of the inputs, followed by glacial runoff (35%) and land runoff
928 (25%). Glaciers are therefore a particularly important contributor to the runoff towards the lake.

929 We also confirm that the ground of the Paiku catchment presents different types of cryological
930 states, from seasonally frozen ground to permafrost. Permafrost coverage shrinks from 27 to 22% of
931 the land area of the catchment from the 1980s to the 2010s (19% loss of the 1980 permafrost area). The
932 whole catchment warms up at a rate of 0.17 °C per decade (2 m deep), with a substantial elevation-
933 dependent variability. This warming is concomitant with an increase in the duration of the seasonal
934 thaw, mainly supported by a progressive delay of the end date of the thaw period. Where permafrost is
935 present, active layer deepening is only observed where evaporation is relatively low ($<150 \text{ mm yr}^{-1}$).

936 Over the simulation period, we also report an increase in evaporation (+10.1 mm per decade),
937 surface and subsurface runoff (+1.3 and +3.5 mm per decade respectively). Together, this leads towards
938 an increase of the runoff/(runoff + evaporation) ratio of +1.2% per decade. Our results also highlights
939 the strong interdependence between the ground thermal and hydrological regimes and the necessity to
940 jointly represent them to accurately quantify evaporation and runoff in this type of environment.

941 Over the last 40 years, the presence of permafrost seems to promote evaporation at the expense of
942 runoff. Yet this trend appears to be climate-dependent and the cryological state of the ground might
943 shift the runoff/evaporation distribution in the other directions under colder and wetter climates. Further
944 studies should investigate this phenomenon and how it might contribute to explain the contrasted lake
945 level evolutions across the QTP.

946

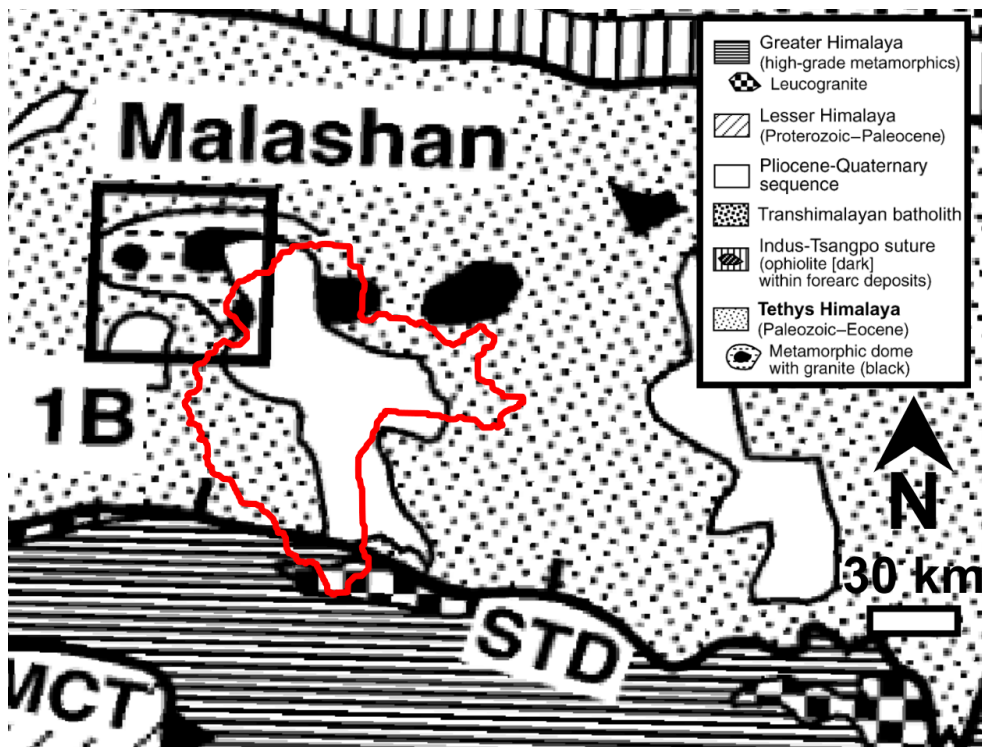
947 **Appendix A: model parameters**

948 *Table A1. Parameters of the model.*

Depth	Layer	Parameter	Values	Source	Calculation
0.0 m	Surface	Albedo	0.24	Modis MCD43A3.006	November mean, 4600-5100 masl
		Emissivity	0.95	Modis MCD43A3.006	November mean, 4600-5100 masl
		Roughness	0.024	-	Adjusted to fit loggers T values
0.0 m		Thickness	0.30 m	HiHydro Soil v1.0	modeling framework
		Porosity	0.5	Shangguann et al. 2013	mean
		Organic	8.60%	HiHydro Soil v1.0	catchment mean
		Mineral	41.40%	-	subtraction (100 - porosity - orga)
0.3 m	Top soil	Soil type	Sand	Shangguann et al. 2013	dominant fraction
		Field capacity	0.32	HiHydro Soil v1.0	catchment mean
		Hydro cond	0.000030 m s ⁻¹	HiHydro Soil v1.0	catchment mean
0.3 m		Alpha	0.028 cm ⁻¹	HiHydro Soil v1.0	catchment mean
		n	1.481	HiHydro Soil v1.0	catchment mean
0.3 m		Thickness	1.70 m	Shangguan et al. 2017	truncation, consistent with literature
		Porosity	0.4	Shangguann et al. 2013	catchment mean
		Organic	4.20%	HiHydro Soil v1.0	catchment mean
		Mineral	55.80%	-	subtraction (100 - porosity - orga)
1.7 m	Bottom soil	Soil type	Sand	Shangguann et al. 2013	dominant fraction
		Field capacity	0.32	HiHydro Soil v1.0	catchment mean
		Hydro cond	0.000016 m s ⁻¹	HiHydro Soil v1.0	catchment mean
		Alpha	0.062 cm ⁻¹	HiHydro Soil v1.0	catchment mean
2.0 m		n	1.707	HiHydro Soil v1.0	catchment mean
2.0 m		Thickness	98.3 m	-	-
		Porosity	0.03	-	-
98 m	Bedrock	Organic	0%	-	-
		Mineral	97%	-	-
100 m		Soil type	Sand	-	-
		Field Capacity	0.03	-	equal to porosity

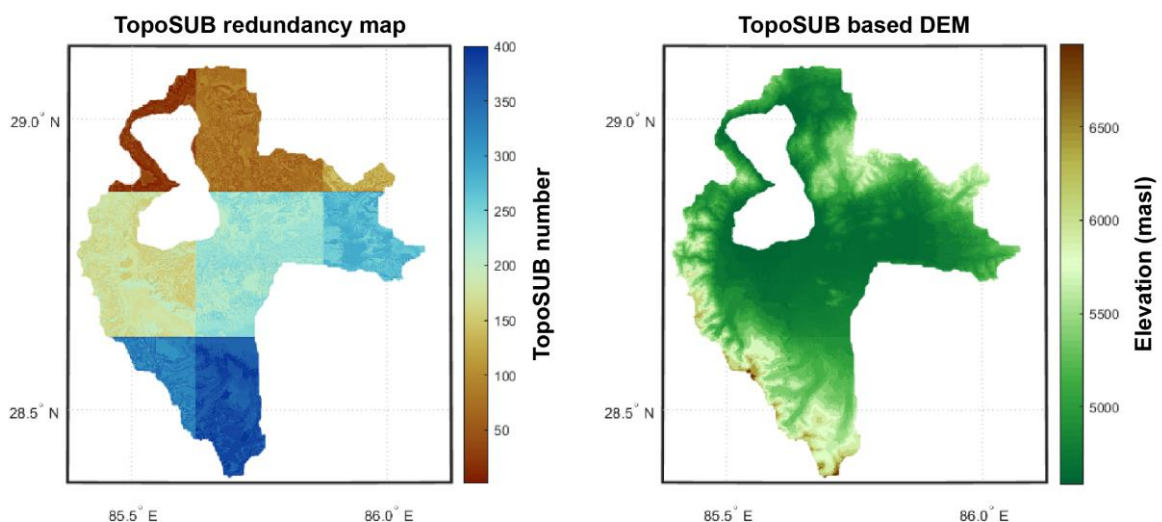
949

950 **Appendix B: Geological map of the catchment**



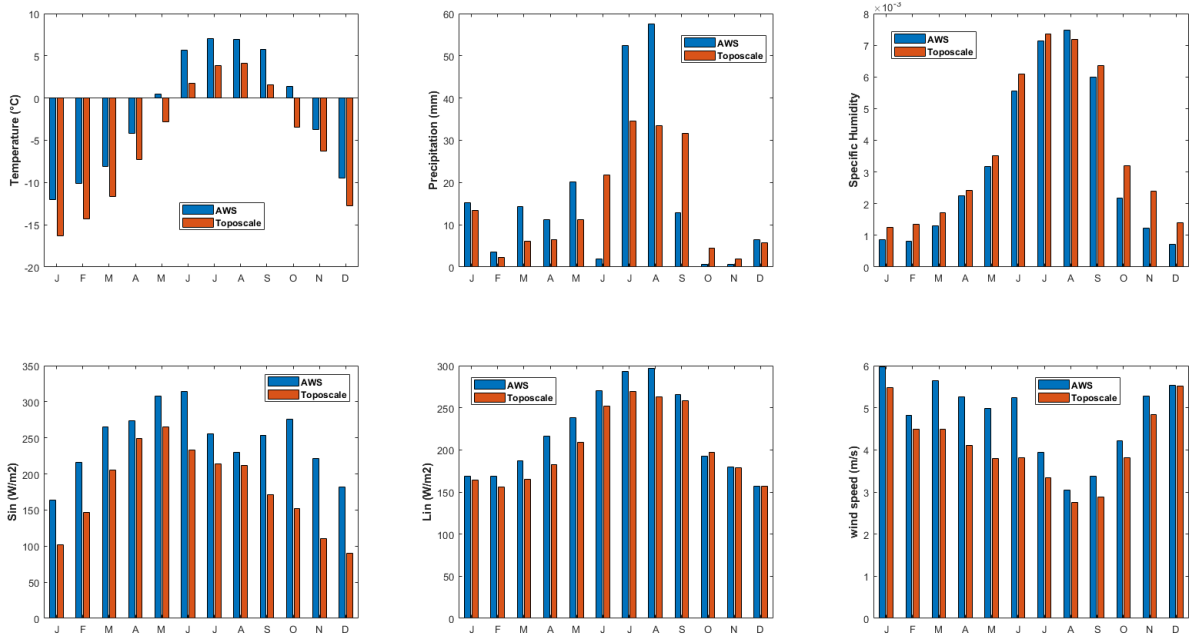
951
 952 *Figure B1. Geology of the catchment. Modified from Aoya et al. (2015). The red contour indicates the*
 953 *limits of the Paiku catchment.*

954 **Appendix C: TopoSUB subsampling of the catchment**



955
 956 *Figure C1. Application of the TopoSUB clustering method (Fiddes and Gruber, 2012) in the Paiku*
 957 *catchment. Left: number of the TopoSUB points. Strong color changes reflect the footprint of the 8*
 958 *ERA5 pixels that the catchment intersects. Small color changes within a given of these zones show the*
 959 *distribution of the 50 TopoSUB points covering each tile (Sect. 3.2.2.) B: topographic map*
 960 *reconstructed using the TopoSUB approach.*

961 **Appendix D: Evaluation of forcing data**



962
 963 *Figure D1. Comparison between the AWS data and the model forcing data downscaled from ERA5 with*
 964 *the TopoSCALE and TopoSUB approaches. Based on the AWS data, a monthly correction factor is*
 965 *applied to the downscaled data so that monthly data matches for the observed period for each variable*
 966 *(methodological details in Sect. 3.2.2.).*

Code availability. The CryoGrid community model (version 1.0) and related documentation are available at: https://github.com/CryoGrid/CryoGridCommunity_source.

Data availability. Field data have been saved on Zenodo.org and will be published with a DOI upon acceptance of the manuscript.

Author contribution. L.M, W. I. and S.W. designed the study. L.M. and M.M. conducted the numerical simulations. S.W., M.L. and L.M. contributed to the model development. F.B., W.I., Y.L. and S.A. acquired field data. L.M., F.B., M.M., P.K., Y.L. and T.M. analyzed and processed the data. J.F. provided downscaled forcing data for the model. All authors contributed to result interpretation and to manuscript preparation.

Competing interests. The authors declare that they have no conflict of interest.

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