



1 **Changing snowmelt and glacier-melt contributions reshape hydrological** 2 **dynamics and flood generation in a high-mountain headwater basin**

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10 **Abstract:** Climate warming has been reorganizing hydrological dynamics in high-mountain rivers by
 11 altering the relative contributions of snowmelt and glacier melt. However, inadequate representation of snow
 12 ablation processes can bias runoff-component partitioning and mislead the inferred roles of snowmelt and
 13 glacier melt in water availability and flooding. Here, we enhance the representation of snowpack dynamics
 14 in a cryosphere-hydrology model (SPHY-S) to quantify how changing snowmelt and glacier-melt
 15 contributions reshape flood regimes in the upper Yulongkashi River during historical (1980-2020) and future
 16 (2021-2050) periods. Results show that SPHY-S robustly reproduces observed snow-cover dynamics across
 17 monthly, seasonal, and interannual timescales, thereby improving the reliability of runoff-component
 18 estimates. By 2050, the relative contribution of snowmelt is projected to decrease by 2%-7% whereas that
 19 of glacier melt increases by 1%-6%, depending on the climate scenario. Under the high-emission scenario,
 20 glacier-melt runoff (increasing from 0.7 to 1.5 billion m³ yr⁻¹) will surpass snowmelt runoff (increasing from
 21 0.9 to 1.2 billion m³ yr⁻¹) to become the dominant contributor. Annual streamflow is projected to increase
 22 by 37%-79% and flood frequency is projected to increase by 84%-105% by 2050, depending on the climate
 23 scenario, with floods shifting from snowmelt-driven to glacier-melt-driven under the high-emission scenario.
 24 The growing contribution of glacier melt to floods creates more sustained high flows, posing challenges for
 25 downstream reservoir operation and disaster mitigation. Our findings highlight that improved representation
 26 of snow processes is essential for reliable runoff partitioning and diagnosis of shifts in flood regimes, thereby
 27 supporting proactive water management strategies in climate-sensitive mountain rivers.



28 1. Introduction

29 Glaciers and seasonal snow in mountainous regions act as natural reservoirs that store water during
30 cold periods and release it during warmer months, thereby regulating downstream streamflow (Immerzeel
31 et al., 2020). Warming-driven changes in timing, magnitude, and runoff components of meltwater-fed rivers
32 strongly influence downstream human and natural systems, particularly in arid and semi-arid regions where
33 river flow from mountain headwaters supports irrigation, hydropower generation, ecosystem maintenance,
34 and broader socioeconomic development (Li et al., 2022; Azam et al., 2021). By the middle of the 21st
35 century, nearly one quarter of lowland populations are projected to depend heavily on runoff originating
36 from mountain regions, compared with only 7% in the 1960s (Viviroli et al., 2020). Therefore, understanding
37 shifts in hydrological dynamics in cryosphere-fed basins has become increasingly important, not only for
38 identifying changes in seasonal water availability, but also for assessing their impacts on flood hazards
39 (Shukla & Sen, 2021).

40 Warming-induced cryosphere degradation alters the generation mechanisms of different runoff
41 components, with pronounced seasonal and regional heterogeneity (Huss et al., 2017; Feng & Gleason,
42 2024). Earlier snowmelt can advance the onset of flood season (Guo et al., 2025), while intensified glacier
43 melt can increase summer and late-season discharge (Lutz et al., 2016), together reshaping intra-annual flow
44 distribution and amplifying hydrological extremes (Han et al., 2024). In High Mountain Asia (HMA),
45 climate warming has already increased glacier meltwater to approximately 1.6 times the equilibrium
46 conditions (Pritchard, 2019). Concurrently, anticipated declines in snowmelt runoff will elevate the
47 vulnerability of irrigation systems that rely on seasonal meltwater, particularly in HMA, Central Asia, and
48 the western United States (Qin et al., 2020; Kraaijenbrink et al., 2021). Warming is also expected to intensify
49 flood risk in cold mountain regions by increasing meltwater inputs and making flood-generating
50 mechanisms more complex, especially where snowmelt, glacier melt, and rainfall interact (Nie et al., 2021;
51 Bai et al., 2025; Rizwan et al., 2023).

52 Despite these well-recognized impacts, disentangling the relative contributions of snowmelt and glacier
53 melt remains challenging, as runoff generation in high-mountain basins is governed by multiple cryospheric
54 processes that are largely simplified in hydrological models. Many glacio-hydrological modeling
55 frameworks have been developed to simulate snowmelt, glacier melt, and their hydrological impacts over
56 the past decades (Mohammadi et al., 2023; Zhang et al., 2025). Among them, the fully distributed Spatial



Processes in Hydrology (SPHY) model has been widely applied because of its robust performance and modest data requirements in data-scarce regions in HMA and the Andes (Terink et al., 2015; Zhu et al., 2023; Marinelli et al., 2025). Yet snow sublimation, an important source of uncertainty, remains marginally considered in many applications. As an important component of the water cycle in high-mountain regions, snow sublimation represents a major ablation pathway by removing a substantial fraction of seasonal snow storage before melt occurs and impacting snowmelt estimation (Strasser et al., 2008; Stigter et al., 2018). Field-based studies have shown that sublimation can account for up to 28% of winter snowfall losses in the Rocky Mountains (MacDonald et al., 2010) and 16% to 42% in parts of the Himalaya (Mandal et al., 2022). However, simulating sublimation has been challenged by coupled effects of energy availability, atmospheric demand, wind, topography, and snow conditions, all of which exhibit strong spatial heterogeneity in high mountains (Sexstone et al., 2018). Empirical relationships between sublimation and meteorological variables are commonly used to upscale sublimation from point observations to the catchment scale (Stigter et al., 2018), including temperature-based linear relationships (J. He et al., 2025) and radiation- or exposure-driven parameterizations (Khanal et al., 2021). Yet these approaches do not explicitly resolve energy balance and atmospheric demand, limiting their physical realism in describing sublimation processes.

Poor representation of snow sublimation can mislead the snowmelt estimation and the resultant runoff-component partitioning, distorting the understanding of changes in hydrological dynamics and flood generation. This issue is particularly important in the northern Tibetan Plateau, which is strongly influenced by the westerlies and characterized by considerable snow sublimation ranging from 0.2-1.8 mm day⁻¹ (Guo et al., 2022). Under climate warming, sublimation has been further enhanced (Guo et al., 2023) and has become a primary driver of winter snow-cover decline (Zheng et al., 2025) in this region. Previous studies have suggested that meltwater runoff dominates streamflow in this region, with snowmelt runoff accounting for over 50% of total discharge (Fang et al., 2018; Wei et al., 2022). However, these studies simplify snow ablation processes and neglect snow sublimation, which can introduce substantial uncertainty in overall streamflow simulation and potentially overestimate snowmelt runoff. This large uncertainty in meltwater simulating hampers, and may even mislead, the development of strategies for addressing future water security risks in these fragile mountain regions.

In this study, leveraging long-term hydrological observations from the upper Yulongkashi River Basin (YRB) in the northern Tibetan Plateau, we develop an enhanced SPHY model (SPHY-S) that explicitly represents snow sublimation and examine its effect on discharge simulation. Supported by SPHY-S, we



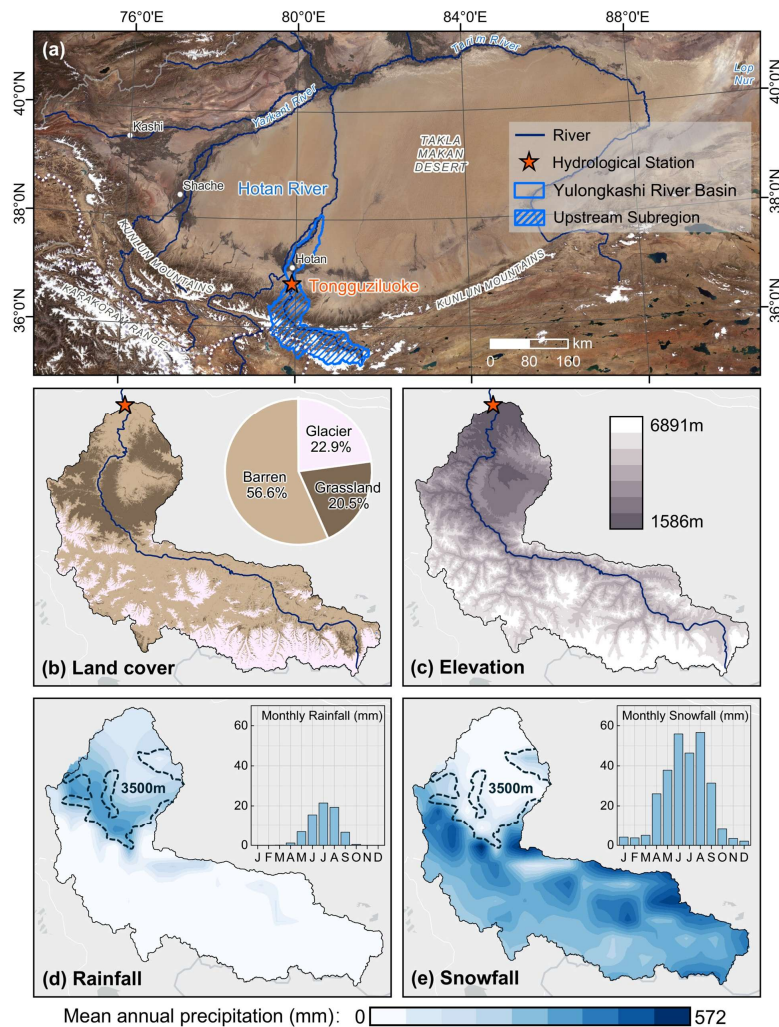
87 further investigate how competing snowmelt and glacier-melt contributions reshape hydrological dynamics
 88 and flood processes between the historical (1980-2020) and future (2021-2050) periods. Our findings
 89 provide insights into how warming-induced meltwater changes impact flood risks and agricultural irrigation
 90 security in cold mountain regions.

91 **2. Study area and materials**

92 **2.1 Study area**

93 The Yulongkashi River, originating on the northern slope of the Kunlun Mountains in southern Xinjiang,
 94 China, is a first-order tributary of the Hotan River and a second-order tributary of the Tarim River (**Figure**
 95 **1a**). Extending approximately 505 km, it constitutes one of the principal water sources for the Hotan region.
 96 This study is focused on the upper Yulongkashi River Basin (YRB), with Tongguziluoke hydrological station
 97 located at the basin outlet, encompassing an area of 14,799 km² (**Figure 1a**). Land cover is dominated by
 98 barren land (56.6%), with glaciers and grassland accounting for 22.9% and 20.5%, respectively (**Figure 1b**).
 99 The catchment exhibits pronounced topographic variation, with elevations ranging from 6891 m in the
 100 southern high mountains to 1586 m along the desert margin to the north (**Figure 1c**). Streamflow is strongly
 101 influenced by glacier melt, snowmelt, and summer precipitation, producing a highly seasonal regime in
 102 which 88.6% of annual discharge occurs between June and September, characteristic of cryosphere-fed
 103 basins (Fang et al., 2018). Precipitation displays a pronounced vertical gradient, being concentrated above
 104 3500 m where it falls almost exclusively as snow (**Figure 1e**). In contrast, areas below 3500 m receive
 105 limited rainfall, contributing only 20% of the annual basin total (**Figure 1d**). Under the influence of westerly
 106 circulation, glaciers in the Karakoram and western Kunlun ranges, including those within the study area,
 107 maintained an overall positive mass balance prior to 2010, a phenomenon widely recognized as the
 108 Karakoram-Kunlun anomaly, which contrasts with the prevailing retreat trends across High Mountain Asia
 109 (Farinotti et al., 2020).

110 Snow sublimation is substantial in this region ($\sim 115 \text{ mm yr}^{-1}$) due to strong westerlies and has further
 111 increased over recent decades (Guo et al., 2022). Previous studies have estimated that meltwater runoff
 112 dominates streamflow in the YRB, with snowmelt runoff accounting for approximately 60% of total
 113 discharge (Fang et al., 2018; Wei et al., 2022). However, these estimates often oversimplify snow ablation
 114 processes by neglecting sublimation, leading to an overestimation of snowmelt runoff and only moderate
 115 streamflow simulation performance, Nash-Sutcliffe efficiency (NSE) ranging from 0.55-0.63.



116
117 **Figure 1. Geographic and environmental characteristics of the study area (upstream subregion of the**
118 **Yulongkashi River Basin).** (a) Geographic setting, displaying a transition from glacierized mountains in
119 the upper reaches to desert plains downstream; major rivers and the Tongguziluoke hydrological station is
120 marked by a red star. (b) Land cover distribution with data source from Yang and Huang, (2021). (c)
121 Elevation map ranging from 1,586 m to 6,891 m, showing the pronounced relief from low-lying plains to
122 high mountains. (d-e) Spatial distributions of mean annual rainfall and snowfall, both delineated by the
123 3,500 m contour line; the insets show their respective monthly patterns.

124 **2.2 Materials**

125 Daily streamflow observations from 1980 to 2011 at the Tongguziluoke hydrological station are
126 provided in the Chinese Hydrological Data Yearbook, with gaps for the years 1990-2000, 2004, and 2006
127 due to limited data accessibility. Meteorological forcing data for historical hydrological simulations,



including near-surface air temperature, specific humidity, wind speed, downward shortwave and longwave radiation fluxes, and precipitation, were obtained from the China Meteorological Forcing Data (CMFD) 2.0 (B. He et al., 2024). CMFD 2.0 provides meteorological forcing at a native spatial resolution of 0.1° and a 3-hourly temporal resolution and integrates ECMWF Reanalysis v5 data, *in-situ* observations and two artificial intelligence datasets (ISCCP-ITP-CNN and TPHiPr). Among widely used gridded datasets in HMA, CMFD demonstrates the best overall performance for the snowfall variable, exhibiting the highest correlation ($CC = 0.790$) and the smallest bias ($BIAS = 0.048$ mm) (Sun et al., 2023). Therefore, it was selected as the forcing dataset in this study.

Soil moisture attributes required for model parameterization were extracted from the HiHydroSoil v2.0 dataset (Simons et al., 2020). Data on annual crop sowing areas were sourced from the Agricultural Statistical Yearbook provided by the Xinjiang Uygur Autonomous Region Bureau of Statistics. To better characterize the complex topography of alpine regions, topographic information was taken from the AW3D30 DSM dataset, which provides 30 m horizontal resolution and 5 m vertical accuracy (Tadono et al., 2014). Glacier distribution was characterized using the Randolph Glacier Inventory (RGI 7.0) (Pfeffer et al., 2014), with glacier thickness initialized from Millan et al. (2022) and annual glacier mass balance changes constrained by remotely sensed elevation change datasets (Hugonnet et al., 2021). Snow cover data for 2001–2020 were obtained from the MOD10A1.061 Terra Snow Cover Daily Global 500 m product (Hall et al., 2016). Land cover information for the study basin and annual cropland dynamics in the Hotan Oasis were taken from the China Land Cover Dataset (CLCD) (Yang & Huang, 2021).

To simulate future hydrological processes under different emission scenarios, we used the High-resolution CMIP6 Downscaled Daily Climate Projections over China (HiCPC) dataset (Yuan et al., 2024), which was generated by downscaling and bias correction of CMFD data. Among these, five general circulation models (GCMs) that showed superior performance in the western Kunlun Mountains were selected as forcing data for future scenario simulations (Table S1).

3. Improving snowpack representation in glacio-hydrological modelling

3.1. Structural modification of SPHY

The Spatial Processes in Hydrology (SPHY) model is designed to simulate key hydrological processes in cryosphere-fed catchments across varying spatial resolutions (Terink et al., 2015) and has been widely applied in recent glacio-hydrological studies (Singh et al., 2021; Budhathoki et al., 2023). Centered on the



157 PCRaster dynamic modeling framework, SPHY employs a conceptual representation of the soil water
 158 balance, snow and glacier dynamics, and surface-subsurface runoff interactions to quantify spatiotemporal
 159 variations in water storage and fluxes. A detailed description of the SPHY model structure, runoff component
 160 formulation, and evapotranspiration parameterization is provided in **Text S1**.

161 While SPHY represents snowmelt using a widely adopted degree-day formulation, it does not explicitly
 162 account for snow sublimation as an additional pathway of snowpack depletion. In mountainous regions,
 163 sublimation constitutes a substantial component of the seasonal snow mass balance, with reported losses
 164 ranging from 10% to 90% of total winter snowfall (Groot Zwaftink et al., 2011; Stigter et al., 2018; Lv &
 165 Pomeroy, 2020). Neglecting this process can lead to systematic overestimation of snow water equivalent
 166 (SWE) and consequent meltwater runoff, particularly in cold and wind-exposed high-elevation
 167 environments (MacDonald et al., 2010). To improve the representation of snowpack depletion and runoff
 168 partitioning, we modified the SPHY source code to develop SPHY-S by adding a sublimation module and
 169 revising the snow and glacier modules.

170 Snow sublimation has previously been estimated using energy-balance approaches, including the
 171 Penman-Monteith (PM) method (Stockert et al., 2025). The PM framework has also been suggested as more
 172 appropriate for high-mountain environments when radiation, wind speed, and humidity data are available
 173 (Shannon et al., 2023). We therefore calculate potential snow sublimation using a modified equation based
 174 on the structural framework of the FAO-56 PM formulation, with its surface and thermodynamic terms
 175 adapted to snow conditions (Eq. 1). Specifically, the net radiation term is reconstructed using snow-surface
 176 properties, explicitly accounting for the high albedo and emissivity of snow (Eq. S11). Saturation vapor
 177 pressure and its slope are replaced by ice-based formulations to represent phase-change processes associated
 178 with sublimation rather than evaporation (Eq. S8). The energy term further uses a latent-heat conversion
 179 factor appropriate for sublimation. Following Stigter et al. (2018), we set the surface resistance r_s , which
 180 represents bulk canopy resistance in the reference-crop formulation, to zero for the snow surface and allow
 181 the aerodynamic resistance r_a to be adjusted to represent unresolved variability in snow-surface roughness,
 182 wind exposure, and sub-grid turbulent vapor exchange. Because r_a is embedded in the aerodynamic term
 183 of the FAO-56 formulation, this adjustment is implemented through a dimensionless factor k_a .

$$184 \quad \text{Sublimation}_p = \frac{0.353 \times \Delta_{ice} \times \max(R_{n,snow}, 0) + k_a \times \gamma_s \times \frac{900}{T_a + 27} \times u_2 \times \max(e_{s,ice} - e_{a,0})}{\Delta_{ice} + \gamma_s} \quad (1)$$



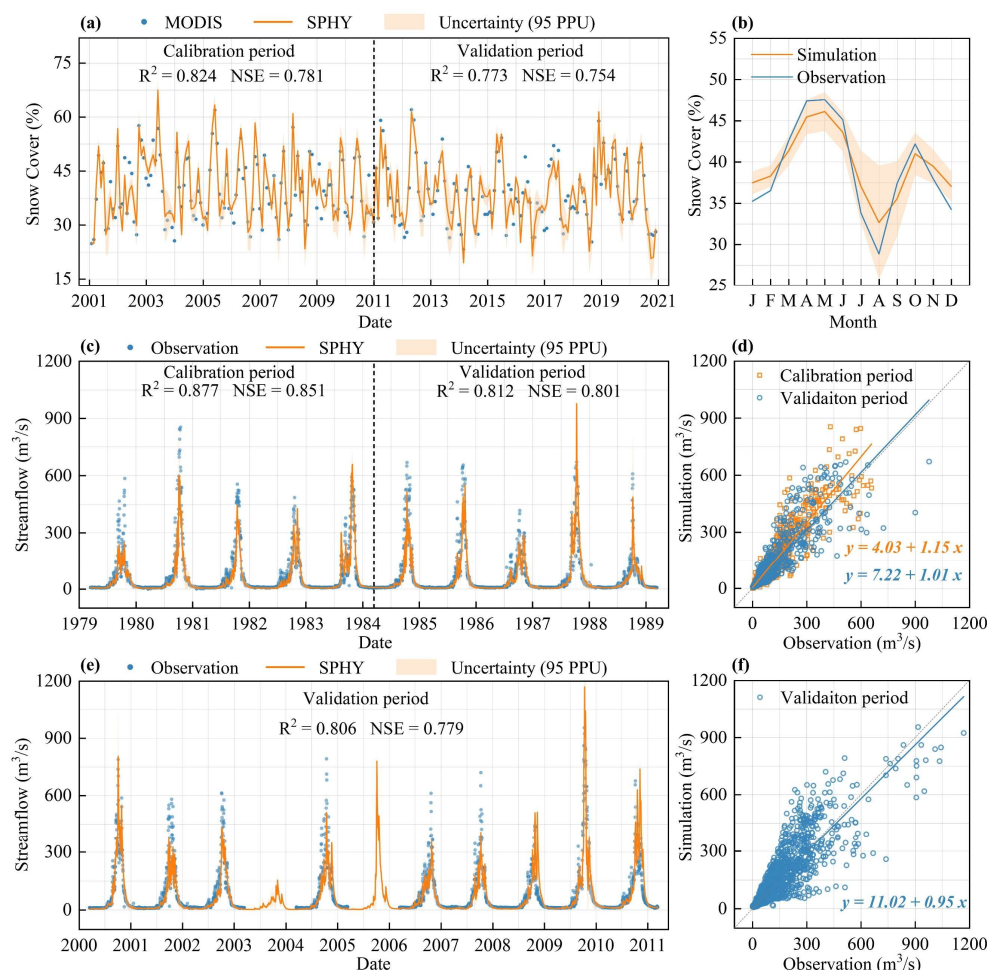
185 where $Sublimation_p$ is the potential snow sublimation rate (mm d^{-1}); Δ_{ice} is the slope of the saturation
 186 vapor pressure curve over ice ($\text{kPa } ^\circ\text{C}^{-1}$); $R_{n,snow}$ is the net radiation at the snow surface ($\text{MJ m}^{-2}\text{d}^{-1}$); γ_s
 187 is the psychrometric constant for sublimation ($\text{kPa } ^\circ\text{C}^{-1}$); T_a is the near-surface air temperature ($^\circ\text{C}$); u_2
 188 is wind speed at 2 m (m s^{-1}); $e_{s,ice}$ is the saturation vapor pressure over ice (kPa); e_a is the actual vapor
 189 pressure of air (kPa). The coefficient 0.353 represents the latent heat conversion factor for sublimation,
 190 distinguishing it from the standard value used for evaporation in the FAO-56 PM equation. The
 191 dimensionless parameter k_a is an adjustment factor for the aerodynamic resistance r_a and is calibrated
 192 together with other snow-related parameters during the snow-process calibration using observed snow-
 193 cover dynamics. The detailed formulations of auxiliary variables, including vapor pressure, radiation
 194 components, and thermodynamic parameters, are provided in **Text S2**. In SPHY-S, potential snow
 195 sublimation is passed to both the snow and glacier modules, where it contributes to the simulation of
 196 snowpack dynamics over non-glacier snow-covered cells and glacier surfaces, respectively. Within each
 197 module, sublimation is constrained to zero when grid-cell air temperature exceeds 0°C , as snow ablation
 198 is assumed to be dominated by melt processes under above-freezing conditions (Mott et al., 2018). Actual
 199 sublimation is further constrained by the available snow water equivalent (SWE) within each grid-cell
 200 snowpack, thereby ensuring mass conservation.

201 **3.2. Model calibration and validation**

202 To mitigate parameter non-identifiability and reduce process compensation that may lead to equifinality,
 203 we adopted a staged, hierarchical calibration strategy where cryosphere-related processes were constrained
 204 before streamflow fitting. Specifically, remotely sensed snow cover was first used to constrain snow-process
 205 parameters. Glacier melt parameters were subsequently calibrated against satellite-derived glacier surface
 206 elevation changes. Only after these cryospheric parameters were fixed were the hydrological parameters
 207 governing runoff generation, flow routing, and baseflow processes calibrated against observed discharge.
 208 Parameter sampling and optimization were conducted using the Statistical Parameter Optimization Tool for
 209 Python (SPOTPY), with consistent prior parameter ranges and a fixed random-seed configuration applied
 210 to ensure reproducibility (Houska et al., 2015). The Nash-Sutcliffe efficiency (NSE) was adopted as the
 211 objective function for parameter optimization, with model performance assessed using both NSE and the
 212 coefficient of determination (R^2). All simulations include a two-year spin-up period and are performed at a



213 spatial resolution of 250 m with daily time steps. Input data were preprocessed before model calibration
 214 (Text S3), and the fitted parameter values are provided in Table S2.



215
 216 **Figure 2. Comparison between observed and SPHY-simulated snow cover and streamflow.** (a) Time
 217 series of MODIS-observed and simulated snow cover fraction during calibration (2001–2010) and validation
 218 (2011–2020). (b) Seasonal cycle of snow cover fraction from observations and simulations. (c, e) Daily
 219 streamflow series for calibration (1980–1984) and validation (1985–2011). (d, f) Scatter plots of observed
 220 versus simulated streamflow during 1980 to 1989 and 2001 to 2012, respectively. Shaded bands indicate the
 221 95% prediction uncertainty (95PPU), defined by the 2.5th and 97.5th percentiles of 500 model simulations.

222 Snow-related parameters were constrained using MOD10A1 V6.1 snow-cover products. Daily
 223 Normalized Difference Snow Index (NDSI) data were obtained via Google Earth Engine, with snow cover
 224 identified using an NDSI threshold of 0.4 and subsequently aggregated to monthly basin-scale snow-cover
 225 fraction (Riggs et al., 2015). In the model, grid cells with simulated SWE exceeding 2 mm were classified



226 as snow-covered to ensure structural consistency with the remote-sensing definition. The temperature
 227 threshold for snowfall (T_{crit}), the degree-day factor for snowmelt (DDFS), and the aerodynamic-resistance
 228 adjustment factor (k_a) governing snow sublimation were calibrated against observed monthly snow-cover
 229 fraction using a split-sample design, with 2001-2010 used for calibration and 2011-2020 reserved for
 230 independent validation. As DDFS is closely coupled with runoff dynamics, this step only constrained its
 231 parameter range, with final optimization performed during the subsequent discharge calibration, whereas
 232 T_{crit} and k_a were fixed after snow-cover calibration. The enhanced SPHY-S reproduced monthly snow-cover
 233 variability with good fidelity, yielding an R^2 of 0.82 for the calibration period and 0.77 for the validation
 234 period (**Figure 2a**), and consistently capturing the seasonal evolution of snow cover without substantial loss
 235 of performance across periods (**Figure 2b**).

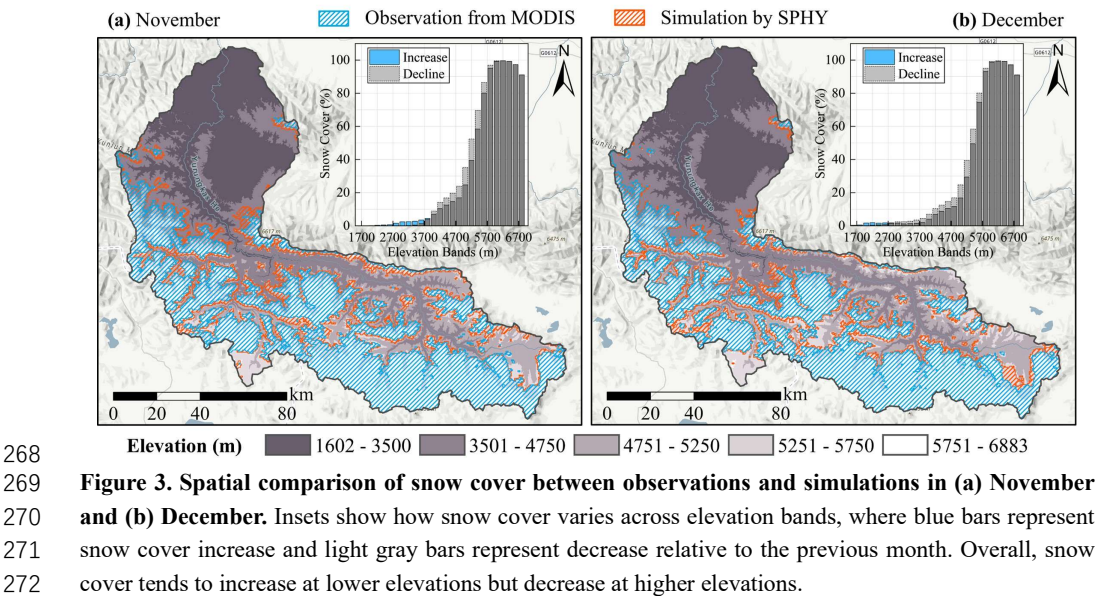
236 Glacier melt degree-day factors for clean ice (DDFG) and debris-covered ice (DDFDG) were calibrated
 237 using glacier surface elevation change data derived from satellite observations (Hugonnet et al. 2021).
 238 During this process, grid cells with glacier coverage exceeding 90% over the 2000-2020 period were
 239 selected, and glacier density was assumed to be 850 kg m^{-3} to convert elevation change into glacier mass
 240 balance. At the basin scale, remote-sensing-based geodetic estimates indicate that glaciers in the upstream
 241 YRB remained close to mass balance, with a mean value ($\pm$ spatial standard deviation) of $0.031 \pm 0.02 \text{ m}$
 242 w.e. yr^{-1} during 2000-2020, consistent with regional assessments reported for the West Kunlun Mountains
 243 (Cao et al., 2020). The SPHY-S simulated glacier mass balance was $0.019 \pm 0.07 \text{ m w.e. yr}^{-1}$, which was
 244 comparable to the remote-sensing estimates. Both results indicate a slightly positive glacier mass balance,
 245 consistent with the reported Karakoram Anomaly in this region (Farinotti et al., 2020).

246 Building upon the constrained snow and glacier parameters, fourteen hydrological parameters were
 247 calibrated against daily discharge observations from the Tongguziluoke hydrological station for 1980-1984
 248 and subsequently validated for 1985-2011. These parameters included key controls on runoff routing and
 249 groundwater processes (**Table S2**), such as the flow recession coefficient (k_x) and the groundwater recharge
 250 delay time (δ_{gw}). At the daily scale, NSE values reached 0.85 in the calibration period and remained at 0.78
 251 across the validation period, demonstrating robust performance in long-term streamflow simulations
 252 (**Figure 2c-f**). Furthermore, the flow duration curve indicates that the model captures high-flow conditions
 253 with moderate negative biases, with the highest 20% of flows underestimated by about 12%, suggesting that
 254 flood magnitudes are represented with reasonable fidelity (**Figure S6**).



255 **3.3. Impacts of snow sublimation**

256 Snow cover in the upstream YRB exhibits pronounced bimodal seasonal variability, with a gradual
257 spring replenishment phase and a distinct autumn peak (**Figure 2b**). During March-May, air temperatures
258 in mountain regions above 3500 m remain persistently below freezing, causing most precipitation to fall as
259 snow and thereby sustaining continuous snowpack accumulation (**Figure 1e, Figure S1**). As temperatures
260 rise, snow cover gradually declines and reaches its minimum in August, covering *ca.* 30% of the catchment
261 (**Figure 2b**). With the onset of autumn, snowfall begins to exceed snowmelt, initiating snow accumulation
262 again. Notably, although snowmelt is largely completed by the end of September, snow cover declines again
263 after October (**Figure 2b**) driven by sublimation and wind-driven redistribution (Kraaijenbrink et al., 2021;
264 Fugger et al., 2024; C. Li et al., 2025). This interpretation is further supported by our analyses, which show
265 that during November–December, snow cover reduction occurs predominantly above 3500 m, whereas
266 lower-elevation areas exhibit a slight increase (**Figure 3**). This spatially redistributed pattern of snow cover
267 and its bimodal seasonal cycle are well reproduced by the SPHY-S (**Figure 2b**).



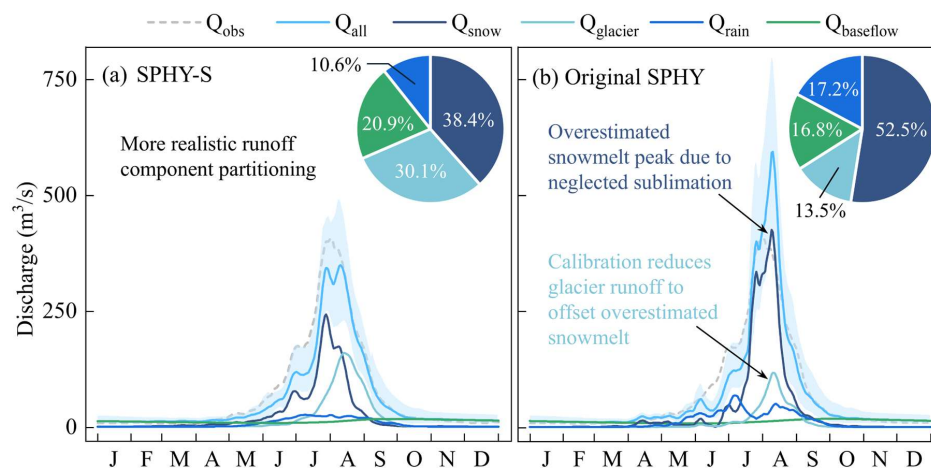


Figure 4. Influence of incorporating snow sublimation on simulated runoff components. (a) Results from the SPHY-S model that explicitly represents snow sublimation. (b) Results from the original SPHY model, where neglecting sublimation leads to an overestimated snowmelt runoff. Curves show the day-of-year (DOY) climatology of runoff components derived from 500 calibration runs for each model. Solid lines represent the median, and the shaded envelope indicates the interquartile range (IQR) of total discharge (Q_{all}). Uncertainty shading for individual runoff components is omitted for clarity. Insets show the relative contribution of each runoff component to total discharge. To ensure consistency, the analysis was restricted to the periods 1980-1989 and 2001-2011, when discharge observations were available. The DOY curves were smoothed using a 7-day moving average.

By improving the representation of snowpack dynamics, SPHY-S thereby provides a more physically reliable basis for subsequent streamflow calibration (Figure 2c-f) and runoff partitioning (Figure 4a). Although melt-related processes were constrained using remote-sensing snow cover and/or glacier mass balance data, key melt-related parameters still need to be calibrated based on fine-time-scale streamflow observations to reduce hydrological uncertainty (Avesani et al., 2025). However, in hydrograph-based calibration, different parameter combinations can reproduce similar streamflow yet rely on unrealistic compensation among runoff components, leading to biased runoff partitioning (Nijzink et al., 2018). Here, we noted that, when calibration was performed without accounting for snow sublimation in the original SPHY model, abnormally high snowmelt estimations (exceeding $300 \text{ m}^3/\text{s}$) emerged during August (Figure 4b), leading to overestimated total runoff and exceeding observations by 21.2%. In some periods, the simulated snowmelt runoff approached or even exceeded the observed discharge, demonstrating a pronounced overestimation of snowmelt. To compensate for this excessive runoff and maintain agreement with the observed hydrograph, the calibration process favored parameter combinations that suppressed the



296 other runoff components, for example by reducing the glacier-melt runoff. When snow sublimation was not
 297 considered, glacier-melt runoff contributed only 13.5% to total discharge in the upstream YRB (**Figure 4b**),
 298 substantially lower than previous estimates of roughly one-third reported in earlier studies (Fang et al., 2018;
 299 Wei et al., 2022) and the estimate in SPHY-S. Moreover, the excessive snowmelt runoff simulated during
 300 August-September led to lower calibrated values of DDFS, which consequently reduced snowmelt
 301 simulations during May-July. This adjustment distorted rainfall-runoff parameters, increasing the
 302 contribution of rainfall runoff to offset underestimated snowmelt runoff in these periods.

303 4. Results

304 4.1. Runoff component partitioning and temporal variability

305 Meltwater dominates the streamflow in the upstream YRB. Snowmelt runoff (36.8%) was the largest
 306 contributor to streamflow during 1980-2020, followed by glacier-melt runoff (30.6%) and baseflow (23.0%),
 307 whereas rainfall runoff contributed only 9.7% (**Figure 5**), as most precipitation falls as snow at high
 308 elevations while rainfall occurs mainly in the downstream reaches located within the hyper-arid Taklamakan
 309 Desert ($<50 \text{ mm yr}^{-1}$) (**Figure 1d-e**). However, this runoff partitioning varied markedly between warm and
 310 cold years. For instance, in 1989, when the basin-mean annual temperature dropped to $-9.21 \text{ }^{\circ}\text{C}$, glacier-
 311 melt runoff accounted for only 5.1% of total discharge (**Figure 5a**). In contrast, in 2013, when temperature
 312 reached a 40-year maximum ($-6.39 \text{ }^{\circ}\text{C}$), the glacier-melt contribution increased to 49.9%. In cold and
 313 relatively dry years, the contribution of snowmelt runoff also tends to be lower. For example, in 2014 (333.8
 314 mm, $-8.75 \text{ }^{\circ}\text{C}$), snowmelt runoff accounted for only 20.0% of total discharge, substantially below the long-
 315 term average of 36.8% (**Figure 5c**). Notably, glacier-melt runoff exhibited the largest interannual variability
 316 among all runoff components, with a coefficient of variation (CV) of 0.88, even greater than the variability
 317 of total streamflow (0.58).

318 From P1 (1980-1999) to P2 (2000-2020), together with a warming-wetting climate in the upstream
 319 YRB, the contributions of snowmelt and rainfall increased from 34.7% and 8.3% in P1 to 38.1% and 10.6%
 320 in P2, respectively, whereas the relative contributions of glacier melt and baseflow declined (**Figure 5**). In
 321 the upstream YRB, Pettitt change-point analysis further identified significant abrupt changes in temperature,
 322 total streamflow, snowmelt runoff, and rainfall runoff around 1997, glacier-melt runoff in 1996, and
 323 precipitation around 2000 (**Figure 5 a, c; Table S4**). These hydroclimatic changes are broadly consistent
 324 with the late-1990s abrupt warming-wetting regime shift reported across the wide High Mountain Asia



(Zhang et al., 2020; Liu et al., 2022). Interestingly, although climate warming increased the fraction of precipitation falling as rain, the total amount of precipitation also increased, and much of this additional precipitation remained fell predominantly as snow in the mountainous regions of the YRB (**Figure S1**). As a result, the increase in total precipitation offsets the reduced snow fraction and leads to a net increase in basin-scale snowfall, thereby enhancing seasonal snow storage and the contribution of snowmelt to streamflow.

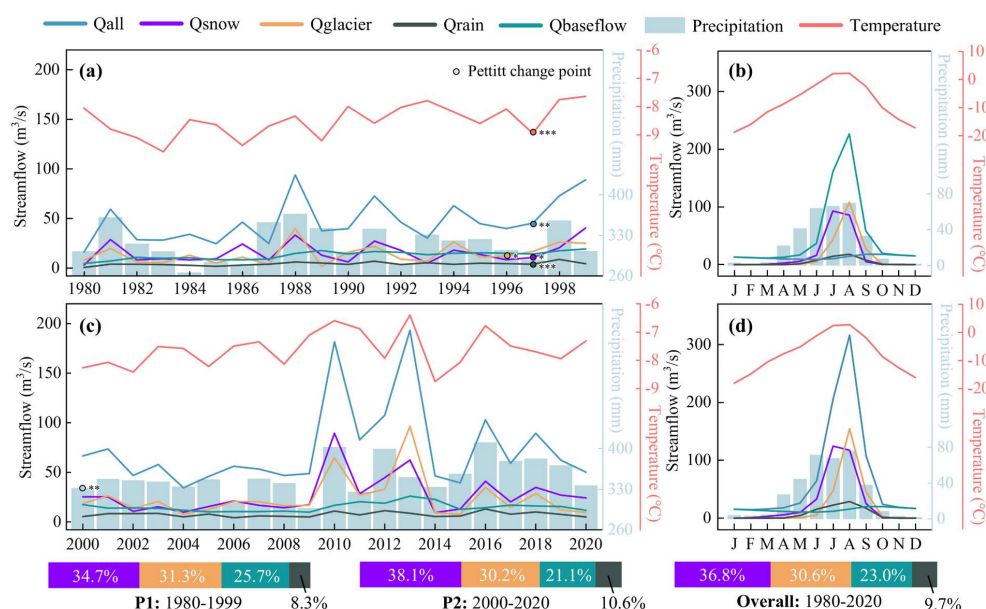


Figure 5. Streamflow components (Q_{snow} , Q_{glacier} , Q_{rain} , Q_{baseflow}) and climatic drivers (temperature and precipitation) in the upstream YRB during 1980-2020. Annual time series are shown in panels (a) and (c) for Period 1 (P1: 1980-1999) and Period 2 (P2: 2000-2020), respectively, while intra-annual (monthly) distributions are presented in panels (b) and (d). Pettitt change-point locations are marked by colored filled circles, with colors corresponding to different variables, and asterisks indicating statistical significance levels. No statistically significant change point was detected for baseflow. Colored bars at the bottom summarize the mean relative contributions of each streamflow component for P1, P2, and the overall period (1980-2020).

From the intra-annual perspective, streamflow typically peaks during June-September (**Figure 6a-d**), coinciding with the warmest months of the year (**Figure 5b-d**). While more pronounced increases in streamflow are observed in spring and autumn. After 2000, due to earlier snowmelt (**Figure S2**) and increased rainfall (**Figure 6g**), streamflow during spring in P2 increased by 45% compared with P1 (**Figure 6h**). In autumn, the increase in streamflow (+58.3%) outpaced that in spring (**Figure 6e-h**), leading to higher



streamflow extremes, especially in September (**Figure 6a**). In summer, mean streamflow in July and August
 increased from 190.5 and 260.3 m³/s to 235.1 and 309.7 m³/s, respectively, while the August 95th percentile
 rose from 587 to 897 m³/s (+52.8%), indicating an intensification of extreme streamflow events (**Figure 6a**).
 Glacier-melt runoff plays an increasingly important role in these high-flow events, with the 95th percentile
 of glacier-melt runoff in August increased from 278 to 626 m³/s (+125%) (**Figure 6c**).

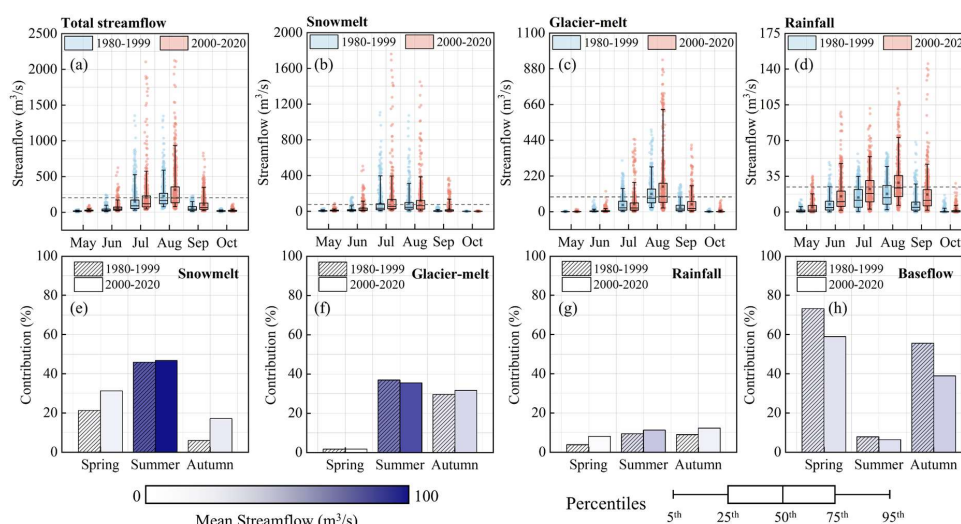


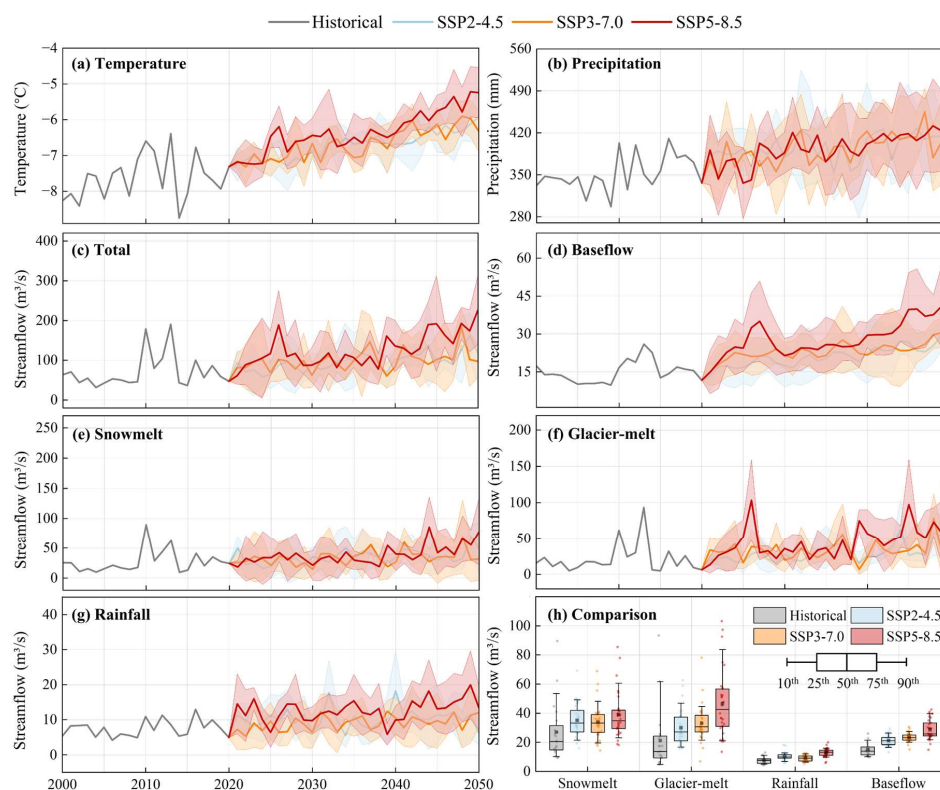
Figure 6. Shifts in seasonal streamflow of the Yulongkashi River between 1980-1999 (P1) and 2000-2020 (P2). (a-d) Seasonal variations of total, snowmelt, glacier-melt, and rainfall runoff simulated by SPHY, showing amplified extremes during P2. The dotted dark grey lines show the 95%-percentile thresholds of daily total, snowmelt, glacier-melt and rainfall runoff during the P1, at 204.8 m³/s, 81.7 m³/s, 90.7 m³/s and 24.5 m³/s, respectively. (e-h) Seasonal percentage contributions of snowmelt, glacier-melt, rainfall, and baseflow to total streamflow, showing increasing shares of snowmelt and rainfall.

4.2. Projection for future changes in runoff components

Continued warming is projected to sustain the increase in streamflow in the upstream YRB at least until 2050 (**Figure 7a-c**), accompanied by an amplified role of glacier melt (**Figure 7d-h**). Under three Shared Socioeconomic Pathways (SSPs) scenarios, basin-scale mean temperature during 2021-2050 is projected to rise by 1.1-1.5 °C, while annual precipitation will increase by 53.6-55.4 mm. Relative to the 2000-2020 average (2.2 billion m³), total streamflow is expected to increase by 36.8%, 40.4%, and 79.4% under SSP2-4.5, SSP3-7.0, and SSP5-8.5, respectively, revealing a nonlinear increase in discharge under high-emission forcing (**Figure 7**). Under SSP2-4.5, the scenario closest to the current emission trajectory, snowmelt runoff, glacier-melt runoff, and rainfall runoff are projected to increase by 30.2%, 38.4%, and 37.6%, respectively, with glacier-melt runoff showing the largest increase (0.3 billion m³). The relative



367 contribution of glacier-melt runoff will increase from 30.4% to 30.9%, whereas that of snowmelt runoff will
 368 decrease from 38.1% to 36.4%. This pattern reflects fundamental differences in the response mechanisms
 369 of glacier and snow melt to climate change, with glacier melt being energy-limited and snowmelt largely
 370 supply-limited once seasonal snowpack is exhausted.



371 **Figure 7. Annual temperature, precipitation, and streamflow components in the Yulongkashi River**
 372 **during the historical period (2000-2020) and the future period (2021-2050) under CMIP6 scenarios.**
 373 (a-g) Time-series line plots showing annual variations, where the gray lines represent the historical period
 374 (2000-2020), and the blue, orange, and red lines correspond to future projections (2021-2050) under the
 375 SSP2-4.5, SSP3-7.0, and SSP5-8.5 scenarios, respectively. The shaded areas denote inter-model uncertainty
 376 (± 1 standard deviation) among five GCMs. (h) Boxplots of runoff components with overlaid annual means
 377 (dots).
 378

379 In the SSP5-8.5 scenario, the mean annual glacier-melt runoff will rise from 0.67 billion m³ in the
 380 historical period to 1.46 billion m³ by 2050, raising its contribution to 36.6% and surpassing snowmelt runoff
 381 (30.7%) as the dominant water source (**Figure 7h**). The interannual variabilities of glacier-melt runoff (CV
 382 = 0.50) and snowmelt runoff (CV = 0.41) are substantially higher than under lower-emission scenarios (0.24-
 383 0.32), indicating greater instability of meltwater with intensified warming. The upper quantiles of annual



glacier-melt runoff increase markedly, implying both higher intensity and greater frequency of glacier-melt-driven high-flow events. Concurrently, rainfall-driven streamflow will increase from 0.24 to 0.40 billion m³ by 2050 (**Figure 7g**). This overlap of short-duration heavy rainfall and ice/snow melting during summer and autumn will further amplify streamflow variability by 2050. Although SSP5-8.5 is increasingly viewed as unlikely, it approximates a potential temperature overshoot and therefore serves as an upper-bound stress test for assessing the maximum hydrological response to climate change. Under such conditions, the role of glacier-melt runoff becomes more clearly manifested. A consistent shift in runoff composition is identified across all emission scenarios, where the contribution of glacier-melt runoff will increase while that of snowmelt runoff will decline. In the SSP3-7.0 scenario, glacier-melt runoff will increase by 53.1%, with its contribution approaching that of snowmelt runoff.

4.3. Shifts in flood frequency, magnitude, and timing

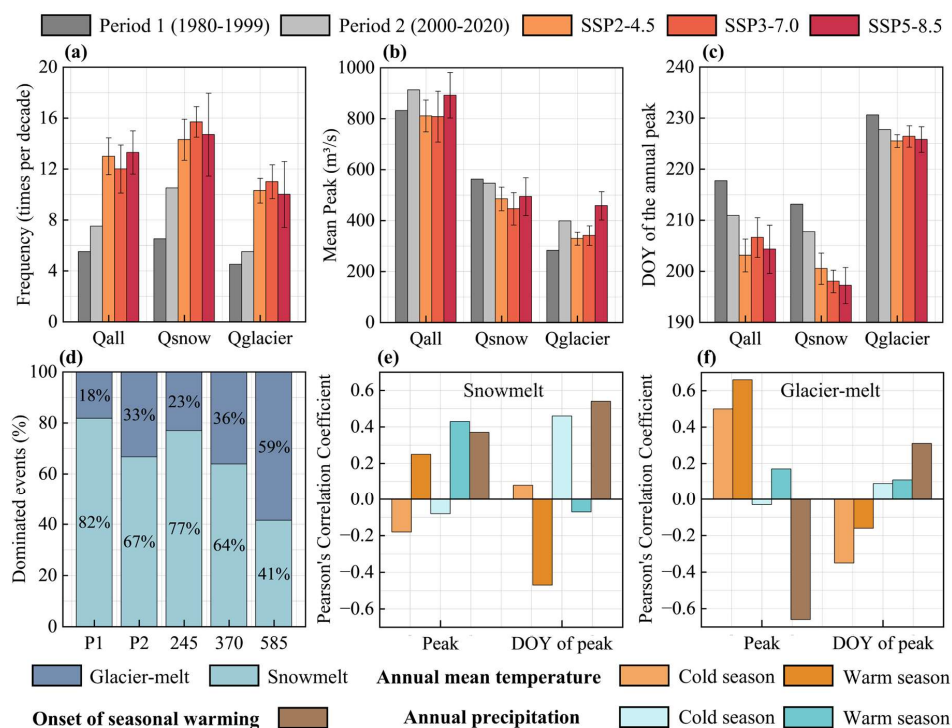
Climate warming increased flood frequency and shifted flood types in the upstream YRB (**Figure 8a, d**). Flood events were identified using the 95th-percentile threshold of daily discharge during P1 (1980-1999) (**Text S4**), consistent with Guo et al. (2025). The flood frequency rose from 5.5 times per decade during P1 to 7.5 times per decade in P2, and further to 12.0-13.3 times per decade by 2050 depending on the emissions scenario. During the historical period, floods were mainly snowmelt-dominated, accounting for 82% of events in P1 and 67% in P2 (**Figure 8d**). Under future scenarios, however, the glacier-melt contribution increased progressively, reaching 59% under SSP5-8.5, demonstrating a shift from snowmelt- to glacier-melt-dominated flood generation.

Changes in flood magnitude are more complicated (**Figure 8b**). Mean peak discharge increased from P1 (832 m³/s) to P2 (913 m³/s), but it declined relative to P2 under SSP2-4.5 (811 ± 62.5 m³/s) and SSP3-7.0 (808 ± 99.8 m³/s), while SSP5-8.5 (892 ± 89.4 m³/s) maintained comparatively high flood peaks. The mean peak of glacier-melt runoff shows a similar pattern under three scenarios. Under SSP5-8.5, the mean peak of glacier-melt runoff is projected to increase by 15.1% (458 ± 55.1 m³/s) relative to P2 (398 m³/s), whereas the mean peak of snowmelt runoff will decline by 9.5% relative to P2 (546 m³/s), showing the increasingly important glacier-melt runoff in floods.

Temperature warming is also shifting flood timing toward an earlier and prolonged flood season (**Figure 8c**). The timing of annual peak discharge shifts from early August in P1 to July under future scenarios, advancing by 11.1-14.6 days depending on the emissions scenario. The advance of snowmelt



runoff peaks (~15 days earlier) during the historical period was greater than that of glacier-melt runoff peaks (~5 days earlier), showing a more rapid response of seasonal snowpack to warming than glaciers. Using the 80th percentile of daily discharge during P1 as the threshold, the onset of the flood season advances by ~22 days in P2, shifting to July, and further to early June and late May under SSP3-7.0 and SSP5-8.5, respectively (**Figure S5**). Meanwhile, the duration from flood-season onset to the annual peak discharge increases from 37 days in P1 to 59 days under SSP5-8.5. Correlation analysis further shows that snowmelt flood peaks were mostly related to warm-season precipitation and the onset of seasonal warming, whereas glacier-melt flood peaks were more closely controlled by temperature, particularly during the warm season (**Figure 8e, f**). Together, these results show that floods in the upstream YRB is occurring more often, arriving earlier, and becoming increasingly associated with warming-driven glacier melt rather than snowmelt alone.



423

424 **Figure 8. Warming-induced shifts in the (a-c) frequency, magnitude, and timing of extreme flood**
 425 **events.** (d) Dominant driver of extreme floods, defined as the streamflow component with the largest
 426 contribution to the total event volume. (e-f) Correlations between annual peak discharge, timing of peak
 427 discharge, and annual mean temperature (light/dark yellow for cold/warm seasons) and precipitation
 428 (light/dark blue for cold/warm seasons) during 1980-2020.



429 5. Discussion

430 5.1. Sublimation reduces snowmelt from high mountains

431 Sublimation represents a major but often overlooked pathway of cryospheric water loss in high
 432 mountain headwaters (Sexstone et al., 2018), and this omission can severely distort estimations of runoff
 433 components and create deceptively reasonable hydrographs arising from compensating errors in runoff
 434 partitioning (**Figure 9**). In the upstream YRB, strong vapor pressure deficits and persistent high-elevation
 435 winds remove nearly 50 mm of SWE annually through sublimation prior to melt (**Figure S3**), reducing the
 436 snowmelt runoff by $\sim 7.5 \times 10^8 \text{ m}^3$, equivalent to about one-third of the catchment's annual streamflow. In
 437 this study, the original SPHY model, which does not account for snow sublimation, produced unrealistically
 438 high snowmelt runoff because snow that would otherwise be removed through sublimation was instead
 439 released as meltwater (Type 1 in **Figure 9**), consequently resulting in poor total runoff simulation
 440 performance (**Figure 4b**). During discharge calibration, DDFS had already reached the lowest value allowed
 441 under the constraint imposed by the remote-sensing snow-cover observations, leaving little room for further
 442 reduction in snowmelt runoff. To offset the resulting overestimation of discharge, the calibration algorithm
 443 favored relatively low values of DDFG and DDFDG, thereby suppressing glacier melt contributions. At the
 444 same time, excessive snowpack carry-over further increased the snow available for melt in the subsequent
 445 hydrological year, reinforcing the overestimation of snowmelt runoff due to cumulative errors.

446 However, the impacts of omitting snow sublimation are not always tangible as we identified in this
 447 high-sublimation study area (Type 1 in **Figure 9**). In other cases, model calibration may achieve good
 448 agreement with observed discharge, while substantial errors hidden in the partitioning of runoff components.
 449 For instance, in regions with moderate snow sublimation, the magnitude of snowmelt runoff overestimation
 450 remains limited (Nedělkov et al., 2025), allowing the calibration algorithm to fully compensate by reducing
 451 glacier-melt runoff (Type 2 in **Figure 9**). In such cases, the erroneously retained snowpack can only be
 452 depleted through excessive snowmelt during early snowmelt season. The resulting spurious hydrograph
 453 agreement can mask substantial errors in runoff partitioning, which could nevertheless be revealed by
 454 independent constraints on runoff partitioning, such as isotope-based evidence where available (Avesani et
 455 al., 2025). In regions characterized by low snow sublimation, the retained snowpack due to neglecting
 456 sublimation can be relatively small (Type 3 in **Figure 9**). During calibration, slightly lowering the DDFS is
 457 sufficient to suppress excessive melt release and maintain a reasonable fit to the discharge hydrograph.



Because the surplus snowpack is limited, short-term deviations in runoff partitioning remain inconspicuous (Zhou et al., 2014). However, the reduced DDFS prevents this retained snowpack from being fully depleted within each melt season, leading to progressive snowpack carry-over. Over time, the persistent snow cover delays glacier ice exposure and shortens the effective glacier-melt period (Stigter et al., 2018), resulting in a gradual drift in runoff partitioning toward increasing snowmelt and declining glacier-melt contributions.

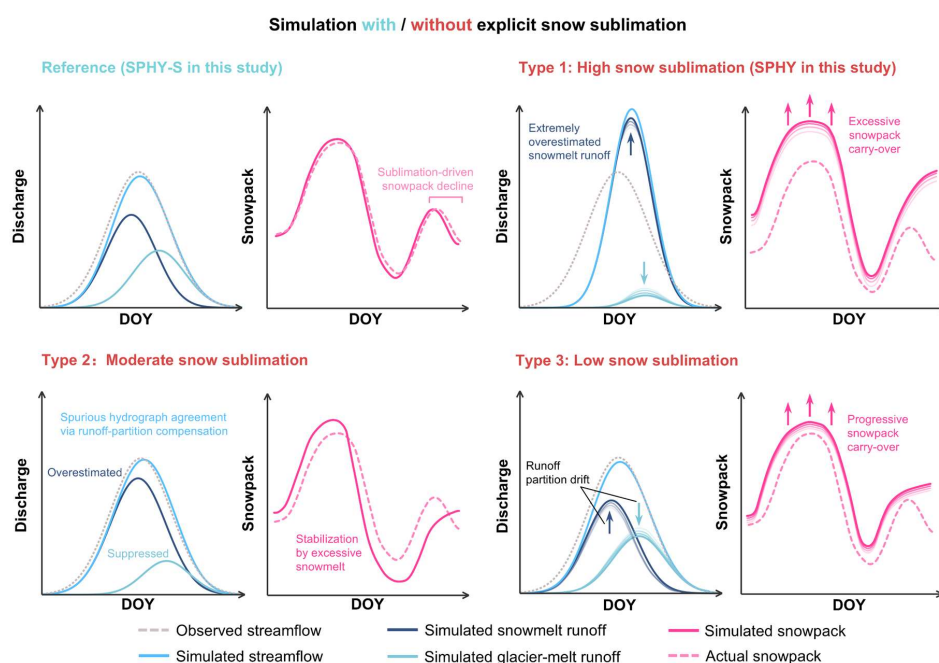


Figure 9. Schematic diagram showing how omission of snow sublimation affects discharge simulation and snowpack evolution across regions with different sublimation intensities. The performance of SPHY-S in this study serves as the reference, while the behavior of SPHY in this study corresponds to Type 1. Types 2 and 3 represent generalized responses to sublimation omission inferred from our analyses and from prior glacio-hydrological studies (e.g., Avesani et al., 2025; Nedělkov et al., 2025; Stigter et al., 2018; Zhou et al., 2014). Omitting sublimation can lead to three distinct outcomes, including severely overestimated snowmelt runoff and degraded model performance under high sublimation, spurious hydrograph agreement arising from runoff-partition compensation under moderate sublimation, and progressive snowpack carry-over with long-term runoff-partition drift under low sublimation.

These runoff-partitioning biases can misrepresent the roles of snowmelt and glacier melt in regulating seasonal water availability (Zha et al., 2025) and flood-generation mechanisms (Khanal et al., 2021). This issue is particularly critical in High Mountain Asia, where irrigation and crop production rely heavily on meltwater and groundwater (Lutz et al., 2022), while sublimation represents an important component of the basin water balance (Buri et al., 2023). By reducing snowpack across mid- to high-elevation zones (**Figure**



3; **Figure S3**), sublimation limits the volume of snowmelt available during the flood season and dampens snowmelt-driven discharge peaks. Meanwhile, reduced snow cover shifts surplus energy toward glacier ablation, increasing the relative contribution of glacier melt to high flows under warming conditions. By explicitly incorporating snow sublimation into SPHY, we developed SPHY-S which successfully reproduces the observed snow-cover dynamics ($NSE = 0.77$; **Figure 2a-b**) and thus robustly separates the runoff components (Reference in **Figure 9**). It also outperformed the state-of-the-art GRDR runoff product (Feng & Gleason, 2024), achieving substantially higher accuracy in seasonal discharge simulation ($NSE = 0.82$ vs. 0.55) and lower bias ($PBIAS = -9.25\%$ vs. $+18.58\%$), particularly in spring and autumn (**Figure S4**). SPHY-S enables a more robust assessment of meltwater availability and supports water-security adaptation in cryosphere-fed basins.

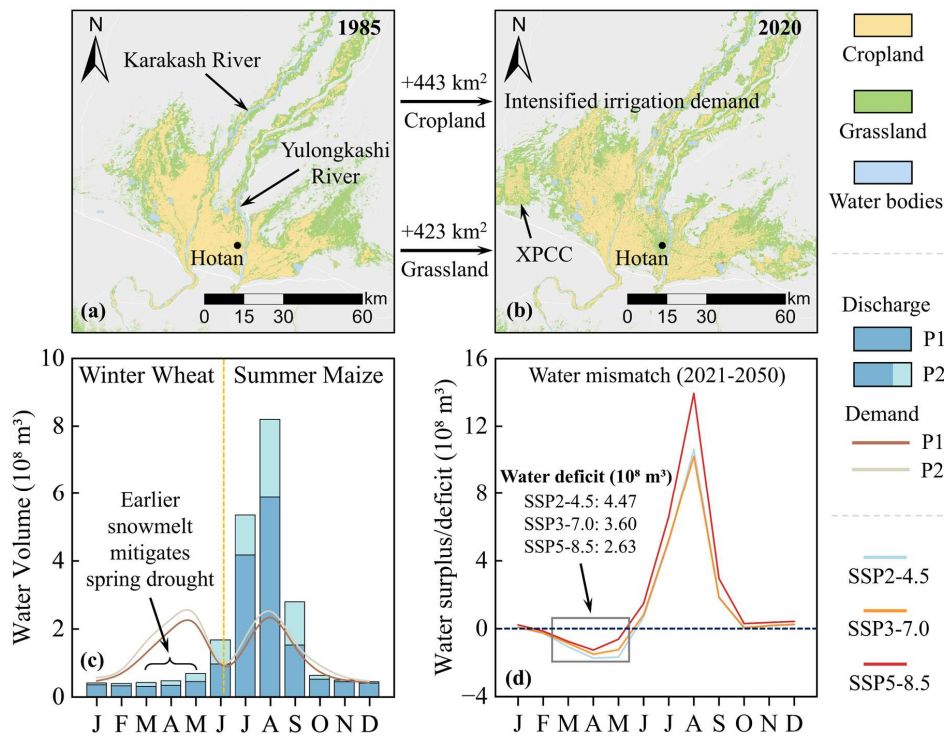
5.2. Implications of shifting runoff generation mechanisms

Meltwater is pivotal for local communities in high mountain regions, bridging seasonal water-demand gaps while also driving flood hazards. Climate warming is reorganizing meltwater generation in the upstream YRB, with glacier-melt runoff playing an increasingly important role relative to snowmelt runoff (**Figure 8**). Similar transitions are also reported in other high mountain regions, including the Himalaya and the Tien Shan (Zhang et al., 2023; B. He et al., 2025; Kraaijenbrink et al., 2021). This shift reflects the contrasting responses of the two meltwater sources to climate change, with snowmelt being supply-limited and glacier melt being energy-limited. Snowmelt is constrained by the finite seasonal snow reservoir, whereas glacier melt remains responsive to increasing melt energy (Guo et al., 2025). Consequently, once seasonal snowpack is depleted, additional warming no longer yields substantially more snowmelt but can still enhance glacier ablation, sustaining flood flows as reported in the southern Tibetan Plateau (Sun et al., 2024).

The muted or weakened response of flood magnitude under moderate emission scenarios (**Figure 8b**) likely results from the combined effects of snow supply limitation and glacier retreat (Xie et al., 2024). Low elevation glacier shrinkage reduces the area capable of producing intense short-term ablation near the outlet (Pradhananga & Pomeroy, 2021), while expansion of the ablation zone toward higher and more distant areas delays meltwater delivery. These delayed inputs tend to broaden the hydrograph rather than sharpen the flood peak. Together with warming-extended snowmelt season (**Figure S5**), the increase in persistence of high flows becomes more pronounced than that in magnitude of peak flow, sustaining high flows later into



507 summer and autumn (**Figure 8c**). Future flood risks in high mountain regions, exemplified as the YRB, may
508 therefore arise less from isolated sharp peaks and more from prolonged periods of elevated discharge,
509 potentially intensified by compound rain-on-snow or rain-on-ice events (Roy et al., 2024; Y. Li et al., 2025).
510 Such sustained high flows can trigger cascading effects, including channel instability and failures within
511 downstream socio-hydrological systems, while simultaneously increasing the exposure of populations and
512 economic assets (Merz et al., 2021).



513
514 **Figure 10. Expansion of cropland and grassland and the seasonal mismatch between irrigation**
515 **demand and streamflow supply.** Land cover data source from Yang and Huang, (2021). (a-b) Changes in
516 cropland and grassland patterns from 1985 to 2020, where XPCC refers to the 224th Regiment of the 14th
517 Division, Xinjiang Production and Construction Corps. (c) Comparison between crop irrigation demand and
518 streamflow during P1 (1980-1999) and P2 (2000-2020), where the irrigation demand is proportionally
519 calculated for the Yulong Kashi River. (d) Future summer water surplus and spring water deficit under three
520 climate scenarios, assuming unchanged irrigation demand in the future (conservative estimate).

521 Beyond flood risks, shifts in runoff components also influence downstream agriculture water security,
522 as snowmelt and glacier melt regulate seasonal water availability and help compensate for spring irrigation
523 deficits (Pritchard, 2019), especially for water-intensive staples such as wheat and maize. In the YRB,



524 increased cryospheric meltwater has helped sustain agricultural expansion in the Hotan Oasis over recent
 525 decades (**Figure 10a**). Following the establishment of the 224th Regiment of the Xinjiang Production and
 526 Construction Corps (XPCC) in 2004, large-scale land reclamation converted hundreds of square kilometers
 527 of desert and grassland into irrigated farmland (**Figure 10b**), substantially intensifying irrigation water
 528 demand. Although annual streamflow increased from 1.56 to 2.19 billion m³ after 2000 (**Figure 5**), a
 529 pronounced seasonal mismatch between water supply and irrigation demand remains (**Figure 10c**). River
 530 discharge during April-May accounts for only about 7% of the annual total, whereas the first major irrigation
 531 peak occurs during the regreening stage of winter wheat, when water demand (**Text S5**) reaches about 0.58
 532 billion m³ but streamflow is only 0.13 billion m³. As a result, spring drought remains a major constraint on
 533 agricultural productivity in the Hotan Oasis, driving increasing dependence on groundwater abstraction (Fu
 534 et al., 2020) and exacerbating groundwater depletion already reported across the Tarim Basin (Wang et al.,
 535 2023). Earlier snowmelt under climate warming (**Figure S2**) has partly alleviated this stress, as spring
 536 streamflow increased by 45% from P1 to P2 (**Figure 6**), with snowmelt runoff accounting for 47.7% of the
 537 increase (**Table S3**). By contrast, glacier melt becomes increasingly important from midsummer onward by
 538 sustaining high flows later in the season (**Figure 6f**; **Figure 8c**), although this contribution is accompanied
 539 by greater flood exposure.

540 Overall, the YRB exemplifies a broad implication of meltwater changes and hydrological regime shifts
 541 shared by many cryosphere-fed basins (Khanal et al., 2021; Wang et al., 2024; Liu et al., 2025), where
 542 seasonal imbalances in water availability pose dual challenges for irrigation supply and flood management.
 543 Warming-enhanced meltwater intensifies flood-season runoff and its variability, and the resulting discharge
 544 pulses in July-August can exceed the design capacity of irrigation channels and levees, damaging cropland
 545 and infrastructure (Merz et al., 2021). At the same time, spring irrigation deficits persist despite rising annual
 546 discharge. With cropland area and crop composition held constant, earlier snowmelt and increased
 547 streamflow reduce the spring (March-May) water deficit to about 0.447 billion m³ under SSP2-4.5 and 0.263
 548 billion m³ under SSP5-8.5 (**Figure 10d**). However, this partial relief is likely transient. Across the Kunlun
 549 Mountains and the wider Tibetan Plateau, glacier meltwater is projected to peak around 2060 and
 550 subsequently decline (Huss & Hock, 2018), while precipitation is expected to become increasingly important
 551 and make downstream supply increasingly dependent on monsoon years. These results suggest that natural
 552 hydrological regulation alone is insufficient to manage the combined pressures of irrigation demand and
 553 flood risk. Multipurpose hydraulic infrastructure with both storage and flood-control functions will therefore



554 be essential to enhance basin resilience (Upadhyay & Bhatia, 2025), with their operation explicitly
 555 accounting for cryosphere-regulated hydrological shifts. For example, in the YRB, reservoirs would require
 556 a minimum regulation capacity of approximately 0.5 billion m³ to alleviate spring water shortages while
 557 attenuating summer monsoon flood peaks.

558 **6. Conclusions**

559 Climate warming is driving shifts in the hydrological dynamics and amplifying water-related risks to
 560 downstream communities. Yet these changes remain poorly understood, partly because snow sublimation
 561 remains insufficiently represented in hydrological models, which can bias runoff partitioning and
 562 misrepresent the relative roles of snowmelt and glacier melt. By incorporating snow sublimation into the
 563 SPHY framework, we developed SPHY-S, a model that improves the representation of snowpack dynamics
 564 and enables more physically consistent simulations of runoff components. Using this model, we quantified
 565 changes in runoff components in the upstream YRB during the historical (1980-2020) and future (2021-
 566 2050) periods, showing that continued climate warming increases total meltwater runoff, amplifies the
 567 contribution of glacier melt to streamflow, and intensifies flood risk.

568 During 1980-2020, snowmelt runoff dominated streamflow in the upstream YRB (36.8%), followed
 569 by glacier-melt runoff (30.6%), and this composition exhibited substantial interannual variability,
 570 particularly between warm-wet and cold-dry years. By 2050, the relative contribution of snowmelt runoff
 571 will decrease consistently in different climate scenarios, while glacier-melt runoff becomes increasingly
 572 important. Glacier-melt runoff nearly matches snowmelt runoff under SSP3-7.0 and overtakes it under
 573 SSP5-8.5, becoming the dominant source of streamflow. Meanwhile, floods are becoming more frequent,
 574 with longer durations but lower peak magnitudes. Earlier and stronger snowmelt runoff pulses increased
 575 spring streamflow by approximately 45%, partly easing irrigation stress. Nevertheless, the seasonal
 576 mismatch between water supply and agricultural demand persists. These findings showcase how warming-
 577 induced hydrological regime shifts in high mountain headwaters impact downstream agriculture and
 578 infrastructure, underscoring the need for hydraulic infrastructure that improves seasonal water allocation,
 579 mitigates flood risks, and strengthens the sustainability of water and food security.

580



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585 **Author contribution**

586 Yu Qi: Conceptualization, Methodology, Software, Formal analysis, Data curation, Visualization, and
 587 Writing - original draft. Ting Zhang: Conceptualization, Methodology, Supervision, Project administration,
 588 and Writing - review and editing. Philip D. A. Kraaijenbrink: Methodology, Interpretation of results, and
 589 Writing - review and editing. Liu Liu: Writing - review and editing. Xiaoya Deng: Writing - review and
 590 editing.

591 **Data availability**

592 The daily CMFD 2.0 data are available from the National Tibetan Plateau / Third Pole Environment Data Center
 593 at <https://doi.org/10.11888/Atmos.tpdc.302088>. Soil moisture attributes data are available from FutureWater at
 594 <https://www.futurewater.eu/projects/hihydrosol/>. Topography data are available from Earth Engine Data Catalog
 595 at https://developers.google.cn/earth-engine/datasets/catalog/JAXA_ALOS_AW3D30_V4_1. Glacier
 596 distribution data are available from National Snow and Ice Data Center at [https://nsidc.org/data/nsidc-](https://nsidc.org/data/nsidc-0770/versions/7)
 597 [0770/versions/7](https://nsidc.org/data/nsidc-0770/versions/7). China Land Cover Dataset can be accessed at <https://zenodo.org/records/15853565>. High-
 598 resolution CMIP6 Downscaled Daily Climate Projections over China (HiCPC) dataset is available from the
 599 National Tibetan Plateau / Third Pole Environment Data Center at <https://doi.org/10.11888/Atmos.tpdc.301122>.
 600 Glacier thickness data can be accessed at <https://doi.org/10.6096/1007>. Glacier elevation change information can
 601 be accessed at <https://doi.org/10.6096/13>. Snow cover data are available from Earth Engine Data Catalog at
 602 https://developers.google.com/earth-engine/datasets/catalog/MODIS_061_MOD10A1.

603 **Competing interests**

604 The authors declare there are no conflicts of interest for this manuscript.

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