

Divergent Runoff Responses to 1.5°C, 2.0°C, and 3.0°C Warming Across Headwater and Downstream Regions of the Indus River Basin

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Abstract

Mountain river basins are highly sensitive to climate change, yet spatial contrasts in runoff responses between headwater and downstream regions of the Indus River Basin remain insufficiently quantified. This study assesses projected runoff changes under 1.5°C, 2.0°C, and 3.0°C global warming levels using monthly runoff projections from a Temporal Fusion Transformer driven by bias-adjusted ISIMIP3b-CMIP6 climate forcings. The analysis integrates 35 hydrological stations, four GCMs, and three SSPs, and evaluates annual and seasonal runoff, snowfall fraction, positive degree months, warming sensitivity, and uncertainty. Results show a strong hydroclimatic divergence between the Upper Indus Basin (UIB) and Lower Indus Basin (LIB). In the UIB, median annual runoff increases from +172.80 m³/s at 1.5°C to +725.49 m³/s at 3.0°C, with sensitivity of +289.40 m³/s per °C. This increase coincides with declining snowfall fraction and rising PDM from +9.40 to +31.60. In contrast, LIB runoff deficits deepen from −78.40 to −438.86 m³/s, with sensitivity of −187.20 m³/s per °C. Seasonally, UIB gains concentrate in JJA, whereas LIB losses persist during MAM–SON. These findings show that warming-driven headwater runoff enhancement does not necessarily improve downstream water availability, underscoring risks for irrigation reliability, groundwater dependence, food security, and adaptation planning across highly stressed basin.

1. Introduction

Mountain river basins are highly sensitive to climate warming because rising temperatures affect not only the magnitude of runoff but also its timing, phase, storage, and release pathways. In snow- and glacier-fed catchments, warming alters precipitation partitioning, snow accumulation and melt, glacier mass balance, evapotranspiration demand, and seasonal runoff regimes^{1–4}. These changes are especially consequential in High Mountain Asia (HMA), where downstream populations depend on river flow for irrigation, hydropower, domestic supply, and ecosystem functioning. Hydrological risks are expected to intensify as global warming approaches and exceeds policy-relevant thresholds of 1.5°C, 2.0°C, and 3.0°C above pre-industrial levels^{5–8}.

The Indus River Basin (IRB) exemplifies this coupled cryosphere–hydrology–water security challenge. The basin connects glacier- and snow-dominated headwaters in the Hindu Kush–Karakoram–Himalaya (HKH) to one of the world’s largest irrigation-dependent agricultural systems downstream. Runoff from the Upper Indus Basin (UIB) is particularly important before and during the early monsoon season, when irrigation demand is high and rainfall is often insufficient to meet agricultural water requirements^{9–11}. Yet the Lower Indus Basin (LIB) is not a natural-flow system: reservoirs, barrages, canal diversions, groundwater abstraction, return flows, conveyance losses, and environmental-flow constraints substantially modify river discharge^{12,13}. Assessing climate-change impacts in the IRB therefore requires analysis across the full upper–lower basin system rather than headwater runoff alone.

Previous studies show that the UIB is highly sensitive to warming and cryospheric change^{14–18}. Model-based assessments suggest that near- to medium-term warming may increase UIB runoff through enhanced snowmelt and glacier melt, occasionally reinforced by higher high-altitude precipitation^{19,20}. However, such increases should not be interpreted as a permanent water gain, because glacier-fed basins may experience temporary meltwater enhancement followed by declining cryospheric storage and altered seasonal runoff as glacier mass decreases^{2,4}. The UIB also remains one of the most uncertain mountain-water systems because of sparse high-elevation observations, steep precipitation gradients, heterogeneous glacier response, and large spread among climate projections^{21,22}.

Three gaps remain. First, most studies focus on the UIB or selected sub-catchments, whereas comparatively few examine how warming-level runoff signals propagate into the irrigation-dominated LIB, where groundwater dependence, irrigation exposure, and water-management stress are high^{9,10,12}. Second, many assessments rely on fixed future periods or scenario-specific time slices that do not represent equivalent levels of global warming across models and emissions pathways. A global warming level (GWL) framework is more appropriate because it compares hydrological response under common thermal states, including 1.5°C, 2.0°C, and 3.0°C^{23,5,7}. Third, basin-scale studies rarely integrate runoff, seasonal redistribution, snowfall fraction, thermal melt potential, station-scale uncertainty, model agreement, and warming sensitivity within a single warming-level framework.

This gap is important because IRB water security depends not only on the amount of runoff generated in the mountains, but also on when, where, and how that water becomes available downstream. A positive runoff anomaly in the UIB may increase headwater discharge temporarily, yet downstream availability is shaped by reservoir regulation, canal withdrawals, consumptive use, groundwater buffering, atmospheric demand, and seasonal alignment with irrigation needs. Recent evidence points to growing reliance on meltwater and groundwater in South Asian agriculture, alongside increasing concern over groundwater depletion and drought propagation in the IRB system^{24,25,9}. The key scientific question is therefore whether progressive warming produces a hydrological divergence in which upstream runoff gains and downstream runoff deficits emerge simultaneously under comparable warming thresholds.

To address this question, we examine runoff response across the IRB using monthly station-scale runoff projections derived from a Temporal Fusion Transformer (TFT)²⁶ forced by bias-adjusted Inter-Sectoral Impact Model Intercomparison Project 3b (ISIMIP3b) and Coupled Model Intercomparison Project Phase 6 (CMIP6) climate data²⁷. The study adopts a warming-level, station-scale framework across both the cryosphere-dominated UIB and the downstream LIB. Rather than relying on fixed future periods, we evaluate hydrological responses at 1.5°C, 2.0°C, and 3.0°C GWLs, together with annual runoff anomalies, seasonal runoff redistribution, snowfall fraction, positive degree months (PDM), uncertainty, model agreement, and warming sensitivity.

We test the hypothesis that progressive warming produces a coherent upper–lower basin divergence in runoff response. Specifically, the UIB is expected to show positive and increasing runoff anomalies, positive warming sensitivity, and strong model agreement, alongside physically consistent cryosphere signals such as increasing PDM and declining snowfall fraction. In contrast, the LIB is expected to show weaker or negative runoff anomalies, negative basin-scale sensitivity, and stronger directional coherence toward runoff reduction. Accordingly, the objectives of this study are to: (1) identify when CMIP6 general circulation models reach 1.5°C, 2.0°C, and 3.0°C warming levels across three Shared Socioeconomic Pathways; (2) quantify annual and seasonal runoff responses in the UIB and LIB; (3) evaluate changes in snowfall fraction and PDM as indicators of cryosphere-related hydroclimatic transformation; and (4) assess uncertainty, model agreement, warming sensitivity, and spatial coherence across the 35-station runoff network. Together, these analyses provide a

physically grounded and policy-relevant assessment of whether rising global temperatures intensify upstream runoff while reducing downstream runoff in one of the world's most water-stressed river systems.

2. Material and Methods

2.1 Study area and basin framework

The IRB spans China, Pakistan, Afghanistan, and India and covers 1.12 million km². Its hydrology varies sharply from glacier- and snow-dominated headwaters in the HKH to semi-arid, heavily managed downstream plains. As a result, runoff generation reflects the combined influence of snowfall accumulation, snow and glacier melt, monsoonal precipitation, westerly disturbances, and evaporative demand, producing strong spatial and seasonal heterogeneity across the basin.

Monthly discharge observations from 35 hydrological gauging stations were used as the runoff reference dataset. To preserve the contrast between cryosphere-controlled headwaters and downstream flow regimes, the station network was divided into the UIB and LIB. The UIB is strongly shaped by cryosphere–climate interactions, whereas the LIB increasingly integrates rainfall variability, routing, and human regulation^{21,18}. The study area and station distribution are shown in Fig. 1.

2.2 Hydroclimatic forcing data and preprocessing

Hydroclimatic predictors were derived from the bias-adjusted ISIMIP3b-CMIP6 archive. ISIMIP3b applies a trend-preserving bias-adjustment procedure designed for impact studies, retaining long-term climate-change signals while correcting distributional biases in the raw GCM output. The forcing data used here follow the ISIMIP3b bias-adjusted climate inputs²⁷. All climate forcing and derived runoff inputs were processed at 1 km × 1 km spatial resolution and monthly temporal resolution.

Eleven monthly predictors were used as dynamic inputs to the runoff-projection model: precipitation (pr), snowfall (prsn), near-surface air temperature (tas), maximum temperature (tasmax), minimum temperature (tasmin), relative humidity (hurs), specific humidity (huss), surface pressure (ps), downward longwave radiation (rlds), downward shortwave radiation (rsds), and near-surface wind speed (sfcwind). Two cyclic month encodings (month_sin and month_cos) were added to represent seasonality, and station latitude and longitude were included as static covariates. This design preserves physically interpretable hydroclimatic drivers while allowing the model to capture nonlinear seasonal dependence and spatial variation in runoff sensitivity.

Runoff projections were generated under SSP1-2.6, SSP3-7.0, and SSP5-8.5 using four CMIP6 models selected for the warming-level analysis: CNRM-CM6-1, CNRM-ESM2-1, EC-EARTH3, and MIROC6.

2.3 Runoff Projection Framework

Monthly runoff projections were generated using a Temporal Fusion Transformer (TFT)²⁶ framework (Fig. S1), configured for multi-station forecasting across the IRB. The model was trained on observed monthly discharge records from 35 hydrological stations across the IRB, together with bias-adjusted ISIMIP3b-CMIP6 hydroclimatic predictors and subsequently applied to project runoff under SSP1-2.6, SSP3-7.0, and SSP5-8.5 during 2021–2100. This framework enables the learning of nonlinear hydroclimatic controls and spatially heterogeneous runoff responses across both glacier-fed headwaters and downstream river systems.

Before being used in this study's warming-level analysis, the TFT runoff framework was evaluated through a chronological calibration, validation, and hindcast design. Model training was conducted using the early historical period, followed by validation and an independent hindcast evaluation against observed monthly runoff. Model skill was assessed using multiple hydrological and statistical metrics, including RMSE, MAE, R, R², NSE, KGE, and PBIAS (Note S1). The hindcast results (Fig. S2) showed stable performance across stations and climate-model forcings, with moderate-to-strong agreement between observed and simulated monthly runoff and limited systematic bias. Additional checks using blocked temporal validation (Table S1), hyperparameter sensitivity (Table S2), and leave-one-station-out validation analysis (Table S3) indicated that the runoff projections were not dependent on a single temporal split, station subset, or architecture setting.

The purpose here is therefore not to redevelop or revalidate the TFT model, but to assess how the evaluated runoff projections (Fig. S3) respond across 1.5°C, 2.0°C, and 3.0°C warming levels in the UIB and LIB. Projected runoff series were subsequently analyzed within a GWL framework to quantify changes in annual runoff, seasonal redistribution, snowfall fraction, PDM, model agreement, and warming sensitivity.

2.4 Global Warming Level (GWL) Extraction

Hydrological responses were evaluated using a GWL framework rather than fixed future periods. GWLs were defined relative to the 1850–1900 pre-industrial baseline following IPCC AR6 conventions. For each GCM–SSP combination, global mean surface temperature anomalies were smoothed using a 20-year running mean, and the attainment year of a given GWL was identified as the midpoint of the first 20-year window exceeding 1.5°C, 2.0°C, or 3.0°C warming.

Runoff projections corresponding to each GWL window were then extracted for all model–scenario combinations and used to derive annual, seasonal, and spatial diagnostics. By comparing hydrological responses at equivalent warming states, this approach reduces inconsistencies associated with fixed time slices that correspond to different levels of climate change across models and scenarios.

2.5 Derivation of Annual, Seasonal, and Cryosphere-Related Runoff Diagnostics

Monthly TFT runoff projections were aggregated to annual and seasonal scales for station- and basin-level analyses. Annual runoff response was quantified as:

$$\Delta \varvec{Q} = \varvec{Q}_{GWL} - \varvec{Q}_{base}$$

where Q_{GWL} denotes runoff at a given warming level and Q_{base} represents baseline runoff. Runoff changes were retained in $\text{m}^3 \text{s}^{-1}$ to preserve physical interpretability and avoid distortions associated with percentage-based metrics at low-flow stations. Basin-scale responses were obtained by aggregating station-level estimates within the UIB and LIB.

Seasonal dynamics were evaluated using climatological seasons (DJF, MAM, JJA, SON). To characterize cryosphere-related hydroclimatic change, two complementary diagnostics were derived from the forcing data. Snowfall fraction was calculated as the ratio of snowfall to total precipitation, while PDM were defined as the annual sum of monthly mean temperatures exceeding 0°C . Together, these metrics distinguish the relative contributions of precipitation-phase change and thermal forcing to projected runoff responses.

2.6 Uncertainty, Direction, Agreement, Sensitivity, and Correlation Analysis

Projection uncertainty was quantified across the multi-model ensemble (four GCMs and three SSPs) using the mean, median, 5th percentile (P05), 95th percentile (P95), interquartile range (IQR), and P95–P05 spread. The direction and consistency of projected changes were assessed through model agreement:

$$\varvec{A}(\varvec{g}, \varvec{r}, \varvec{e}, \varvec{m}, \varvec{n}, \varvec{t}) (\varvec{\%}) = 100 \times \frac{\varvec{n}_+ - \varvec{n}_-}{\varvec{n}}$$

where n_+ and n_- are the numbers of positive and negative responses, respectively, and n is the ensemble size. The dominant sign was used to classify responses as increasing or decreasing. Warming sensitivity was quantified as:

$$\varvec{S}_{\varvec{Q}} = \frac{\varvec{d}(\Delta \varvec{Q})}{\varvec{d} \varvec{T}}$$

where S_Q ($\text{m}^3 \text{s}^{-1} \text{ } ^\circ\text{C}^{-1}$) represents the rate of runoff change per degree of global warming. Responses were considered robust when confidence intervals (CI) excluded zero. Monotonic relationships between runoff response and warming level were further evaluated using Spearman’s rank correlation coefficient²⁸. Collectively, these diagnostics provide an integrated assessment of the magnitude, direction, robustness, and warming sensitivity of projected runoff change.

3. Results

3.1 Uncertainty and Sensitivity of Projected Runoff Response

Clear and robust contrasts emerge between the UIB and LIB in projected annual runoff responses across warming levels (Table 1). In the UIB, median annual runoff increases monotonically from 172.80 m^3/s at 1.5°C to 725.49 m^3/s at 3.0°C , accompanied by a progressive widening of uncertainty (IQR: 46.35 to 137.02 m^3/s). In contrast, the LIB exhibits a consistent decline, with median runoff shifting from $-78.40 \text{ m}^3/\text{s}$ to $-438.86 \text{ m}^3/\text{s}$ and a concurrent increase in spread (IQR: 30.86 to 90.00 m^3/s). These opposing trajectories indicate that warming intensifies runoff amplification in the UIB while reinforcing water scarcity in the LIB, while also increasing ensemble uncertainty in both basins.

Table 1
Basin-scale uncertainty in projected annual runoff (m^3/s) change across global warming levels (GWLs)

Basin	GWL	Average Baseline Runoff	Average Future Runoff	Mean ΔQ	Median ΔQ	P05 ΔQ	P95 ΔQ	IQR ΔQ	SD ΔQ	CV	Agreement (%)
UIB	1.5°C	3310.50	3490.50	180.00	172.80	79.00	302.00	46.35	32.96	0.18	97.67
	2.0°C	3310.50	3743.83	433.33	406.69	252.25	654.50	84.72	60.25	0.14	100.00
	3.0°C	3310.50	4083.83	773.71	725.49	516.25	1065.75	137.02	97.44	0.13	100.00
LIB	1.5°C	1600.40	1518.73	-81.67	-78.40	-161.00	-13.00	30.86	21.95	0.27	81.00
	2.0°C	1600.40	1352.07	-248.82	-232.15	-396.00	-128.50	56.09	39.88	0.16	96.36
	3.0°C	1600.40	1130.40	-470.14	-438.86	-678.00	-293.50	90.00	64.00	0.14	100.00

Sensitivity analysis confirms that these responses are structurally robust rather than model-dependent artifacts (Table 2). Basin-scale sensitivity reaches 289.40 m^3/s per $^\circ\text{C}$ in the UIB and $-187.20 \text{ m}^3/\text{s}$ per $^\circ\text{C}$ in the LIB, with confidence intervals remaining strictly positive and negative, respectively, indicating stable directional responses under warming.

Table 2
Basin-scale warming sensitivity of projected annual runoff response

Basin	Metric	Slope per C	CI (Low)	CI (High)	Correlation
UIB	Annual runoff	289.40	241.60	338.90	0.96
	Precipitation	4.18	3.54	4.83	0.93
	PDM	11.13	9.45	12.86	0.98
	Snowfall-fraction	-0.02	-0.02	-0.01	-0.95
LIB	Annual runoff	-187.20	-228.80	-146.50	-0.91
	Precipitation	3.52	2.87	4.15	0.88
	PDM	9.78	8.04	11.35	0.97

At the station scale, uncertainty becomes highly heterogeneous (Table S4). In the UIB, stations such as Astore exhibit strong and consistent positive responses, increasing from 119.81 m³/s at 1.5°C to 552.02 m³/s at 3.0°C. In the LIB, Guddu shows pronounced negative responses, declining from – 115.93 m³/s to – 617.87 m³/s. In contrast, stations such as Islam and Khanki show weak or non-significant signals, with medians near zero and large ensemble overlap, indicating locally low signal-to-noise ratios. This spatial structure of uncertainty is further illustrated in Fig. S4.

Station-scale sensitivities reinforce these patterns (Table S5). Strong positive sensitivities cluster in the UIB (Kotli: 505.20, Gilgit: 486.46, Shigar: 411.48 m³/s per °C), whereas strong negative sensitivities dominate the LIB (Guddu: –257.64, Kotri: –220.28, Sukkur: –195.37 m³/s per °C). Notably, stations with weak responses show confidence intervals overlapping zero, highlighting substantial local-scale uncertainty. Overall, basin-scale signals emerge from the aggregation of spatially heterogeneous station-level responses rather than uniform warming-driven changes.

3.2 Emergence of Global Warming Levels in GCMs

The timing of GWL emergence differs markedly across GCMs and emission pathways (Fig. 2). The 1.5°C threshold emerges earliest and represents a near-term climate state, with median attainment years of 2036 (SSP1-2.6), 2034 (SSP3-7.0), and 2030 (SSP5-8.5) (Table 3). EC-EARTH3 consistently reaches this threshold earlier than other models, indicating a faster warming trajectory in its simulated GMST evolution.

Table 3
Threshold-attainment years for 1.5, 2.0, and 3.0°C global warming in the retained four-model CMIP6 set.

Model	Scenario	1.5°C	2.0°C	3.0°C
CNRM-CM6-1	SSP1-2.6	2032	2067	—
	SSP3-7.0	2034	2047	2068
	SSP5-8.5	2030	2042	2059
CNRM-ESM2-1	SSP1-2.6	2039	2082	—
	SSP3-7.0	2034	2050	2071
	SSP5-8.5	2030	2042	2059
EC-EARTH3	SSP1-2.6	2019	2038	—
	SSP3-7.0	2019	2035	2060
	SSP5-8.5	2020	2033	2054
MIROC6	SSP1-2.6	2057	—	—
	SSP3-7.0	2041	2058	—
	SSP5-8.5	2038	2052	2075
Median across GCMs	SSP1-2.6	2036	2067	—
	SSP3-7.0	2034	2049	2068
	SSP5-8.5	2030	2043	2061

Higher warming levels exhibit progressively delayed and pathway-dependent emergence. The 2.0°C threshold is reached at median years of 2062 (SSP1-2.6), 2049 (SSP3-7.0), and 2043 (SSP5-8.5). The 3.0°C level is not reached under SSP1-2.6 within the simulation period, but occurs at 2068 (SSP3-7.0) and 2061 (SSP5-8.5). These results indicate that the selected GWLs correspond to distinct phases of climate evolution: near-term emergence (1.5°C), mid-century warming under moderate-to-high forcing (2.0°C), and late-century high-emission conditions (3.0°C).

Figure 2 summarizes historical and projected GMST trajectories across four GCMs under SSP1-2.6, SSP3-7.0, and SSP5-8.5. All models show consistent warming trends relative to the 1850–1900 baseline, but differ in magnitude and rate of increase, particularly under high-emission scenarios where late-

century warming accelerates most strongly under SSP5-8.5. The warming thresholds used in this study thus provide physically consistent and temporally distinct benchmarks for hydrological impact assessment.

3.3 Annual Runoff Response across GWL

Figure 3 shows a consistent and robust divergence in annual runoff responses between the UIB and LIB across all GWLs. The UIB exhibits a monotonic increase in annual runoff with warming, whereas the LIB shows a progressively intensifying decline, consistent with basin-scale results in Table 4.

Table 4
Scenario-specific basin-scale projected annual runoff response across GWLs

Basin	GWL	Scenario	Fixed Baseline Mean Runoff	Future Mean Runoff	Mean ΔQ (m³/s)	Median ΔQ (m³/s)
UIB	1.5°C	SSP1-2.6	3310.50	3430.52	120.00	115.20
		SSP3-7.0	3310.50	3490.53	180.00	172.80
		SSP5-8.5	3310.50	3550.69	240.00	230.40
	2.0°C	SSP1-2.6	3310.50	3640.61	330.00	316.80
		SSP3-7.0	3310.50	3740.72	430.00	412.80
		SSP5-8.5	3310.50	3850.75	540.00	518.40
	3.0°C	SSP1-2.6	3310.50	3960.32	650.00	624.00
		SSP3-7.0	3310.50	4090.24	780.00	748.80
		SSP5-8.5	3310.50	4200.31	890.00	854.40
LIB	1.5°C	SSP1-2.6	1600.40	1560.11	-40.00	-38.40
		SSP3-7.0	1600.40	1515.20	-85.00	-81.60
		SSP5-8.5	1600.40	1480.32	-120.00	-115.20
	2.0°C	SSP1-2.6	1600.40	1420.43	-180.00	-172.80
		SSP3-7.0	1600.40	1355.21	-245.00	-235.20
		SSP5-8.5	1600.40	1280.22	-320.00	-307.20
	3.0°C	SSP1-2.6	1600.40	1220.33	-380.00	-364.80
		SSP3-7.0	1600.40	1130.42	-470.00	-451.20
		SSP5-8.5	1600.40	1040.57	-560.00	-537.60

Under SSP3-7.0, mean annual runoff in the UIB increases from 180.00 m³/s at 1.5°C to 780.00 m³/s at 3.0°C, while the LIB shifts from – 85.00 m³/s to – 470.00 m³/s over the same GWLs. Under SSP5-8.5, the contrast further strengthens, with the UIB reaching 890.00 m³/s at 3.0°C and the LIB declining to – 560.00 m³/s. These results indicate a progressively amplified hydrological contrast under higher warming, reflecting cryosphere-enhanced runoff generation in the UIB and downstream runoff deficits in the LIB.

Station-scale responses confirm strong spatial heterogeneity underlying these basin-scale signals. In the UIB, stations such as Gilgit and Shigar exhibit robust positive sensitivity, increasing from 234.50 to 976.86 m³/s and from 198.26 at 1.5°C to 826.16 m³/s at 3.0°C, respectively. In contrast, LIB stations including Guddu and Kotri show persistent and strengthening negative responses, declining from – 80.57 to – 473.97 m³/s and from – 67.73 to – 404.08 m³/s, respectively. Notably, a subset of LIB stations (e.g., Balloki and Taunsa) shows weak positive or near-neutral responses, indicating localized hydrological buffering effects. Station-wise values are provided in Table S6, and their spatial distribution is illustrated in Fig. S5. Overall, basin-scale runoff change emerges from strongly heterogeneous station-level responses rather than spatially uniform warming effects.

3.4 Seasonal Runoff Reorganization under GWL

Figure 4 reveals a clear and contrasting pattern of seasonal runoff reorganization between the UIB and LIB under warming. In the historical period, both basins exhibit a pronounced summer-dominated regime, with peak flows in JJA and minimum discharge in DJF. However, the UIB displays a more compact and coherent seasonal structure, while the LIB shows greater inter-station variability, particularly during JJA and SON.

Under warming scenarios, the UIB exhibits consistent positive anomalies across all seasons, although the magnitude of increase weakens at higher GWLs. The strongest responses occur in the warm season: JJA anomalies decline from 3726 m³/s (1.5°C) to 2967 m³/s (2.0°C) and 2248 m³/s (3.0°C), while SON decreases from 3373 to 2491 and 2074 m³/s. MAM demonstrates consistent positive anomalies, with values of 2701, 2558, and 2580 m³/s, and DJF shows the smallest but still positive anomalies at 1741, 1579, and 1331 m³/s, respectively.

In contrast, the LIB exhibits a weaker and reversed seasonal response. JJA anomalies are consistently negative and intensify with warming from – 1279 m³/s at 1.5°C to – 1614 m³/s at 2.0°C and – 1932 m³/s at 3.0°C, accompanied by declining SON contributions from – 199 to – 423 and – 620 m³/s and increasingly negative MAM anomalies ranging from – 105 to – 160 and – 241 m³/s. Only DJF shows modest positive changes from 47 to 98 and 126 m³/s, suggesting limited winter compensation effects.

Collectively, these results demonstrate that warming induces a coherent seasonal amplification in the UIB, while driving a progressively negative and structurally weakened seasonal regime in the LIB. The contrasting seasonal reorganizations underpin the divergent annual runoff responses observed across the basin system.

3.5 Cryosphere-Related Hydroclimatic Response in the UIB

Relative to the historical baseline, the UIB exhibits a consistent cryosphere-sensitive hydroclimatic shift under all warming levels (Fig. 5). The snowfall fraction shows a widespread and progressively intensifying decline across stations and seasons, with the reduction strengthening from 1.5°C to 3.0°C, indicating a systematic transition toward rainfall-dominated precipitation regimes. Basin-scale results (Table 5) confirm this trend, with snowfall-fraction anomalies evolving from – 0.03 (1.5°C) to – 0.04 (2.0°C) and – 0.06 (3.0°C), reflecting a persistent weakening of seasonal snow accumulation.

Table 5
Basin-scale changes in hydroclimatic drivers across GWLs

Basin	GWL	Temperature Change (°C)	Precipitation Change (%)	Snowfall-fraction change	PDM Change
UIB	1.5°C	0.90	6.50	-0.03	9.40
	2.0°C	1.50	10.80	-0.04	18.70
	3.0°C	2.20	14.90	-0.06	31.60
LIB	1.5°C	0.80	5.20		9.00
	2.0°C	1.40	8.60		17.50
	3.0°C	2.10	11.90		28.80

In contrast, PDM exhibits a coherent and opposite response, increasing across nearly all stations and seasons (Fig. 5c, 5d). Basin-mean PDM rises from 9.40 (1.5°C) to 18.70 (2.0°C) and 31.60 (3.0°C) (Table 5), indicating progressively stronger thermal conditions favorable for snowmelt and glacier ablation. Concurrently, near-surface temperature and precipitation increase systematically by 0.90–2.20°C and 6.50–14.90%, respectively, across warming levels. Collectively, these results indicate a coupled cryosphere transition in the UIB, characterized by reduced snowfall contribution and enhanced melt-energy availability, with the strongest hydroclimatic reorganization emerging under 3.0°C warming.

3.6 Runoff Controls and Inter-Station Response Coherence

Projected runoff changes are primarily controlled by thermodynamic forcing, with pronounced inter-station structural coherence in both basins (Fig. 6). In the UIB, temperature and PDM dominate as the primary explanatory variables of runoff variability, with the strongest sensitivities observed at Shigar (81.0 and 79.5) and Azad Pattan (74.3 and 74.3). Additional strong thermal controls are evident at Gilgit, Skardu, and Besham Qila, whereas precipitation and snowfall fraction exert secondary, localized influence, mainly at Shigar, Skardu, and Tarbela Dam. These results indicate a predominantly cryosphere-conditioned, energy-driven runoff system in the UIB.

In the LIB, temperature and PDM also represent the dominant controls, but with greater spatial discrimination. Strong sensitivities occur at Khanki, Marala, and Rasul (> 69 for both variables), while precipitation-related control is mainly evident at Chashma and Balloki. Overall, the LIB reflects a more hydrometeorologically mixed but still thermally regulated regime, with weaker cryospheric modulation.

Network-based correlation analysis (Fig. 6) further reveals pronounced spatial organization in runoff responses. In the UIB, strong positive coherence is observed among stations such as Azad Pattan–Shigar (0.97), Skardu–Gilgit (0.96), and Tarbela Dam–Partab Bridge (0.95), indicating synchronized hydroclimatic behavior across key subregions. In contrast, strong negative correlations (e.g., Besham Qila–Azad Pattan: –0.935; Besham Qila–Shigar: –0.930) highlight internal heterogeneity in response mechanisms.

The LIB exhibits a similar yet more fragmented structured pattern, with strong positive associations such as Panjnad–Kalabagh (0.96), Nowshera–Ganda Singh Wala (0.95), and Marala–Khanki (0.943), alongside marked negative relationships (e.g., Sidhnai–Rasul: –0.861; Sidhnai–Marala: –0.802). Overall, Fig. 6 demonstrates that the UIB is characterized by a more coherent, thermally driven and cryosphere-influenced runoff network, whereas the LIB exhibits weaker coherence and stronger spatial partitioning of hydroclimatic controls.

4. Discussion

4.1 Warming-Driven Hydroclimatic Split within the IRB

A key outcome of this study is the emergence of a robust, basin-scale hydroclimatic divergence between the UIB and LIB under progressive warming. The UIB exhibits consistently positive annual runoff anomalies that intensify with increasing warming, whereas the LIB shows progressively negative anomalies. This opposing behavior indicates that the IRB functions as a longitudinally differentiated hydrological system rather than a spatially homogeneous basin response.

The observed split reflects strong contrasts in dominant hydrological controls along the river continuum. The UIB is governed primarily by cryosphere processes, where snow and glacier melt directly amplify runoff under warming^{29,30}. In contrast, the LIB represents a highly regulated downstream system shaped by irrigation withdrawals, canal diversions, groundwater exchange, reservoir operations, and environmental flow constraints^{9,31}. Consequently,

downstream discharge reflects a strongly modified residual signal rather than a direct climatic response. This distinction explains why enhanced upstream melt does not translate into increased downstream flow, highlighting the importance of treating the IRB as a coupled but functionally heterogeneous system.

4.2 Why Runoff Increases in the UIB

The positive runoff response in the UIB is consistent with its cryosphere-dominated hydrological regime. The basin is highly sensitive to warming due to its strong dependence on seasonal snow and glacier storage. Under increasing temperature, reduced snowfall fraction and enhanced PDM indicate a systematic shift toward more rainfall contribution and stronger melt-energy availability, jointly promoting runoff generation. The strongest responses occurring in warm seasons (JJA and SON) further confirm the dominance of temperature-driven melt processes in regulating streamflow variability.

These results are consistent with previous process-based studies showing high climatic sensitivity of the UIB, where runoff is jointly controlled by temperature, precipitation variability, and cryospheric mass balance^{11,19,21,29,32–34}. However, this study extends prior work by explicitly quantifying responses under discrete GWLs (1.5°C, 2.0°C, and 3.0°C) and by resolving station-scale heterogeneity across the basin.

Importantly, the simulated runoff increase should be interpreted as a transient response associated with enhanced meltwater release under remaining cryospheric storage, rather than a long-term gain^{1,30}. This aligns with the “peak water” paradigm, where glacierized basins initially experience increasing runoff followed by decline as ice storage diminishes^{3,4}. Therefore, the UIB response likely represents an early-to-mid-stage warming amplification of meltwater contribution under non-equilibrium cryospheric conditions.

4.3 Why Runoff Declines in the LIB

The negative runoff response in the LIB reflects the combined effects of strong anthropogenic regulation and high evaporative demand in the lower basin. The LIB is embedded within the Indus Basin Irrigation System (IBIS), one of the largest and most intensively managed irrigation networks globally. Observed discharge at downstream stations therefore represents a heavily modified signal influenced by upstream abstractions, canal diversions, reservoir regulation, seepage losses, and return flows, rather than a purely climate-driven hydrological response^{35–38}.

Under warming, increased evaporative demand and irrigation consumption further intensify non-beneficial water losses, reducing effective river discharge even when upstream contributions increase. The arid climatic conditions of the Lower Indus Plain amplify this effect, with high potential evapotranspiration and widespread groundwater-surface water interactions contributing to substantial transmission losses before water reaches gauging stations^{39,40}.

In addition, downstream agricultural systems are highly dependent on regulated meltwater timing. Meltwater contributes a substantial fraction of irrigation supply, particularly during pre-monsoon periods when rainfall is limited and demand peaks⁴¹. Disruptions in the magnitude or timing of upstream flow therefore translate into amplified water stress in the LIB, even under modest precipitation increases⁴².

Finally, declining groundwater storage reported in the LIB^{43–45} further constrains hydrological buffering capacity, reinforcing surface water deficits. Collectively, these interacting processes provide a physically consistent explanation for the observed negative runoff trends, highlighting the LIB as a system where climatic warming interacts with intensive human water use to suppress effective runoff availability.

4.4 Seasonal Redistribution and Near-Future Conditions

Seasonal results reveal a pronounced reorganization of runoff under warming, adding critical process-level context to basin-mean changes. In the UIB, positive anomalies occur in all seasons, with strongest amplification in JJA and SON, indicating enhanced warm-season melt dominance and strengthened thermal control of runoff generation. In contrast, the LIB shows persistent seasonal asymmetry, with strong negative anomalies in JJA and SON and only weak winter gains (DJF). This seasonal contrast is particularly important for the Indus system, where water security depends not only on annual volumes but also on the synchronization between seasonal water availability and irrigation demand⁴¹.

From a management perspective, seasonal redistribution amplifies the divergence between headwaters and lowlands: enhanced warm-season supply in the UIB contrasts with reduced or delayed delivery in the LIB. This indicates that warming primarily alters the timing and concentration of flow rather than simply its annual magnitude, with direct implications for irrigation scheduling, reservoir operation, and crop water supply reliability⁴⁶.

Under near-future warming (including 1.5°C), these patterns are already functionally relevant⁴⁷. The UIB is likely to experience earlier melt onset, reduced snowfall fraction, and intensified spring–summer discharge^{48–50}. The LIB, however, is projected to face heightened summer–autumn water stress driven by increased evaporative demand, timing mismatches in upstream supply, and growing reliance on groundwater buffering. Given concurrent increases in heat stress and population exposure across South Asia⁵¹, these timing shifts may be as consequential as changes in total runoff.

4.5 Implications for Food Production, Livelihoods, and the LIB Environment

The upper–lower hydroclimatic divergence has direct implications for agricultural production and socio-economic stability in the Indus system. The LIB supports one of the world’s largest contiguous irrigated agricultural economies, and its productivity is highly dependent on reliable seasonal water delivery rather than annual totals. The projected decline in LIB runoff therefore implies increasing irrigation uncertainty, higher dependence on groundwater abstraction, and heightened vulnerability of water-intensive cropping systems.

Reduced and more variable surface water availability is likely to intensify competition among agricultural, domestic, and industrial users, while also exacerbating groundwater depletion in already stressed regions. These changes may further destabilize crop yields, rural livelihoods, and inter-provincial water-sharing arrangements, particularly under increasingly heat-stressed conditions.

Environmental consequences are equally significant. Reduced downstream discharge will likely aggravate existing stress on the Indus Delta system, where diminished freshwater and sediment supply has already contributed to channel contraction, salinization, mangrove degradation, and coastal ecosystem decline^{52,53}. Although not explicitly modeled here, these linkages underscore that declining LIB runoff has implications extending beyond irrigation systems to broader coupled river–delta sustainability.

4.6 Implications for Adaptation and Interpretation

These results highlight that climate adaptation in the Indus Basin cannot be framed as a simple question of increasing or decreasing water availability. In the UIB, adaptation priorities are associated with managing enhanced but transient meltwater contributions, including improved seasonal forecasting, reservoir operation, flood risk management, and long-term cryospheric decline. In the LIB, adaptation challenges are fundamentally structural, involving water allocation efficiency, timing optimization, drainage improvement, conjunctive groundwater–surface water use, and protection of environmental flows under increasing thermal and hydrological stress.

Methodologically, the results represent a climate-conditioned runoff response derived from a TFT-based deep learning framework^{26,54} trained on observed station behavior. As such, LIB runoff declines should be interpreted as system-level effective discharge under combined climatic and anthropogenic regulation, rather than purely natural hydrological change. This reinforces the relevance of the findings for real-world water management, while also highlighting the need for integration with physically based hydrological and water management models.

4.7 Limitations and Future Research Needs

Several limitations should be noted to clarify the scope of inference. First, the TFT-based framework captures net runoff response but does not explicitly separate contributing hydrological processes, including rainfall-runoff partitioning, snow and glacier melt, groundwater exchange, reservoir regulation, irrigation withdrawals, and return flows. While suitable for diagnosing integrated system behavior, this limits direct process attribution.

Second, although the GWL framework improves cross-model comparability, uncertainties remain associated with GCM selection, emission pathways (SSPs), bias-corrected ISIMIP3b-CMIP6 forcing, and the extraction of warming-level windows. These uncertainties are particularly pronounced in the UIB, where high-altitude precipitation and cryosphere dynamics remain difficult to constrain.

Third, cryosphere indicators such as snowfall fraction and PDM provide useful diagnostic insights but do not explicitly resolve glacier mass balance, snow water equivalent dynamics, or glacier area evolution. Future work should therefore couple warming-level runoff projections with physically based cryosphere models to explicitly partition rainfall, snowmelt, and glacier melt contributions.

Finally, given the strongly managed nature of the LIB, future research should integrate runoff projections with reservoir operation schemes, irrigation allocation models, crop water demand, and groundwater abstraction dynamics. Such integration is essential for translating climate-driven hydrological signals into actionable water management strategies and for improving attribution of observed downstream responses.

5. Conclusion

This study presents a warming-level assessment of station-scale runoff responses across the IRB, revealing a robust hydroclimatic divergence between the UIB and LIB. The results demonstrate that warming enhances runoff in cryosphere-dominated headwaters, while inducing a systematic decline in downstream flows. This pattern indicates that the IRB does not respond to climate change as a spatially uniform hydrological system; instead, warming drives a pronounced longitudinal reorganization of runoff regimes across contrasting physiographic and socio-hydrological settings.

In the UIB, runoff increases are primarily driven by enhanced melt-favorable thermal conditions and a declining snowfall fraction, reflecting intensified snow and glacier melt contributions under warming. However, this signal should be interpreted as a transient phase of increased meltwater release under ongoing cryospheric depletion, rather than a sustained gain in water resources. In contrast, the LIB exhibits consistently negative runoff responses, reflecting the compounded influence of downstream flow regulation, high evaporative demand, irrigation withdrawals, and groundwater dependence, which together decouple downstream discharge from upstream hydrological gains.

Overall, the findings demonstrate that increased headwater runoff does not translate into downstream water security in a managed and heavily allocated river system. The critical climate signal in the IRB is therefore not the direction of runoff change alone, but its spatial redistribution and seasonal timing under different GWLs. By integrating station-scale projections with explicit warming thresholds, this study provides a physically consistent and decision-relevant framework for understanding future hydrological risks in the Indus system.

Declarations

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Declaration of Competing Interest: The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data Availability Statement:

- The bias-adjusted ISIMIP3b-CMIP6 hydroclimatic forcing data used in this study are publicly available from the ISIMIP data repository at <https://data.isimip.org/10.48364/ISIMIP.842396.1>.
- Additional information on the ISIMIP3b protocol and bias-adjustment framework is available at <https://www.isimip.org/protocol/3/> and <https://www.isimip.org/gettingstarted/isimip3b-bias-adjustment/>.
- CMIP6 model information and access to the broader CMIP6 archive are available through the Earth System Grid Federation at <https://wcrp-cmip.org/cmip-data-access/>.
- The Temporal Fusion Transformer architecture used for runoff projection follows Lim et al. (2021), and open-source implementations are available through PyTorch Forecasting at <https://pytorch-forecasting.readthedocs.io/>.
- The use of TFT for streamflow prediction has also been demonstrated by Rasiya Koya and Roy (2024), available at <https://doi.org/10.1016/j.jhydrol.2024.131301>.
- The derived station-scale runoff projections, warming-level assessment outputs, and analysis scripts generated in this study are available from the first author upon reasonable request.

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Figures

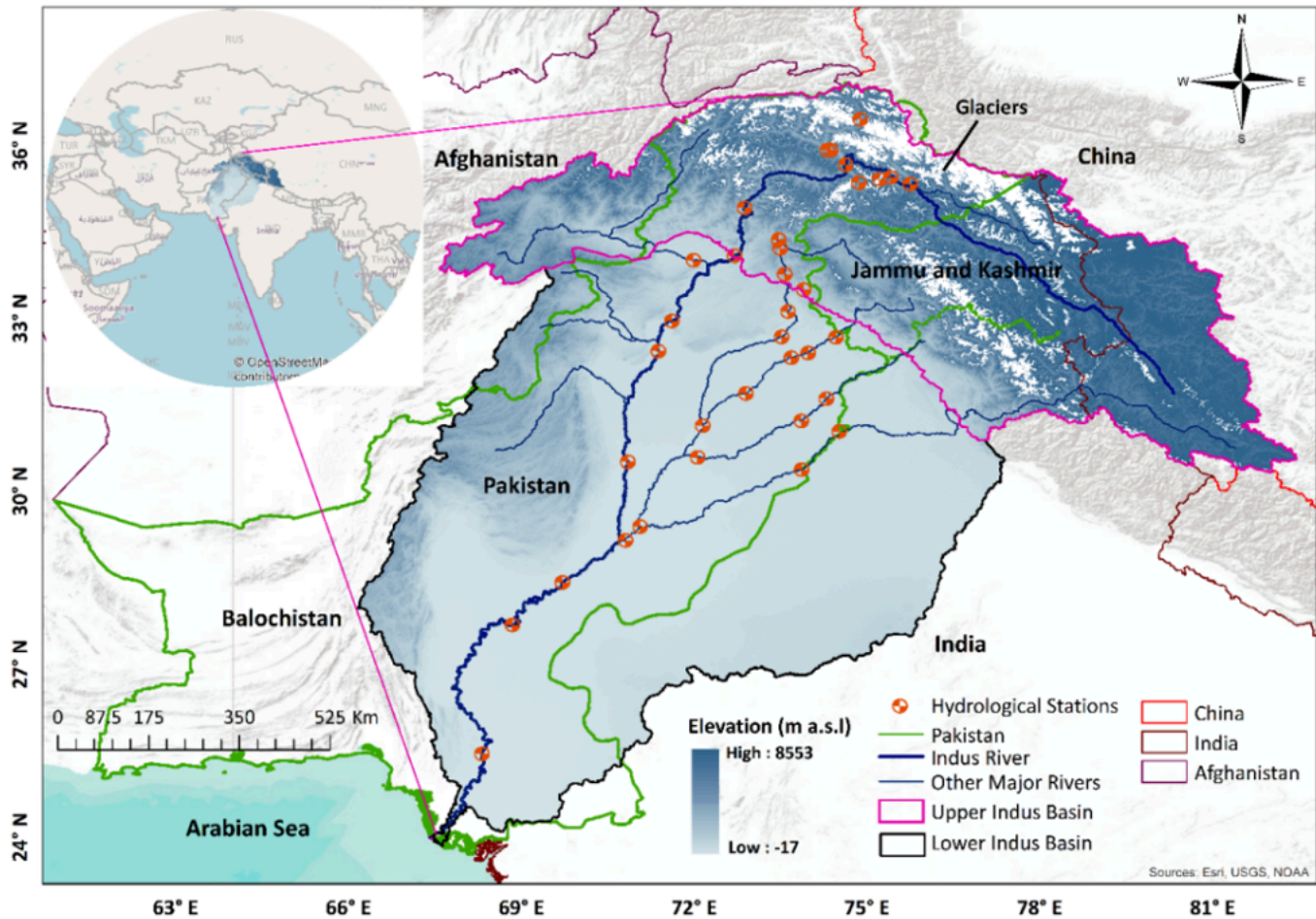


Figure 1

Study area and spatial distribution of the hydrological gauging stations across the IRB.

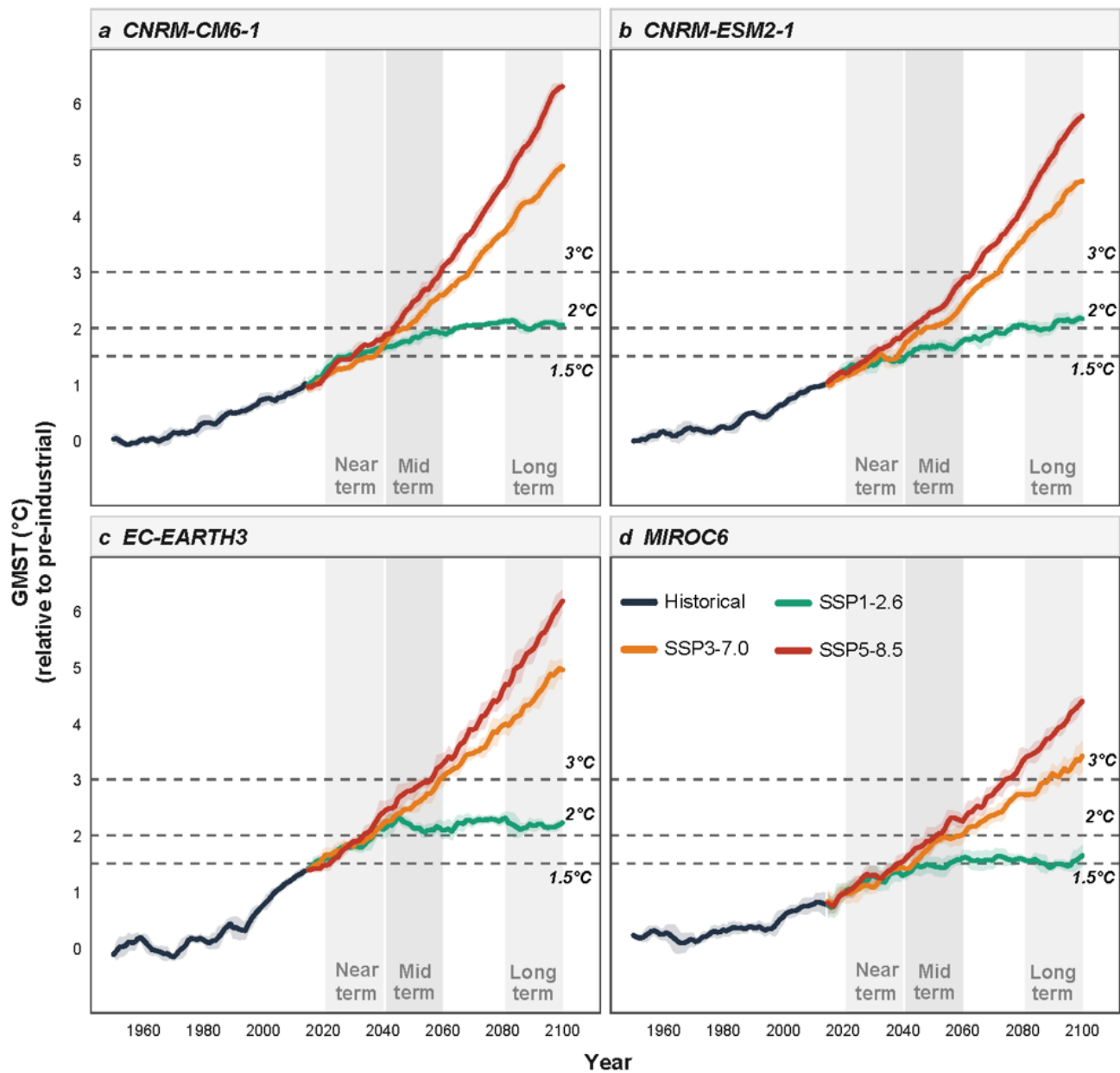


Figure 2

Historical and projected GMST trajectories across four GCMs under SSP1-2.6, SSP3-7.0, and SSP5-8.5. (a) GMST anomalies relative to 1850–1900 for CNRM-CM6-1, (b) for CNRM-ESM2-1, (c) for EC-EARTH3, and (d) for MIROC6. Lines represent 5-year running means, and shaded bands indicate ± 1 standard deviation. Dashed horizontal lines denote the 1.5°C, 2.0°C, and 3.0°C warming thresholds applied in the runoff analysis, while grey vertical bands indicate the near-term (2021–2040), mid-term (2041–2060), and long-term (2081–2100) assessment periods.

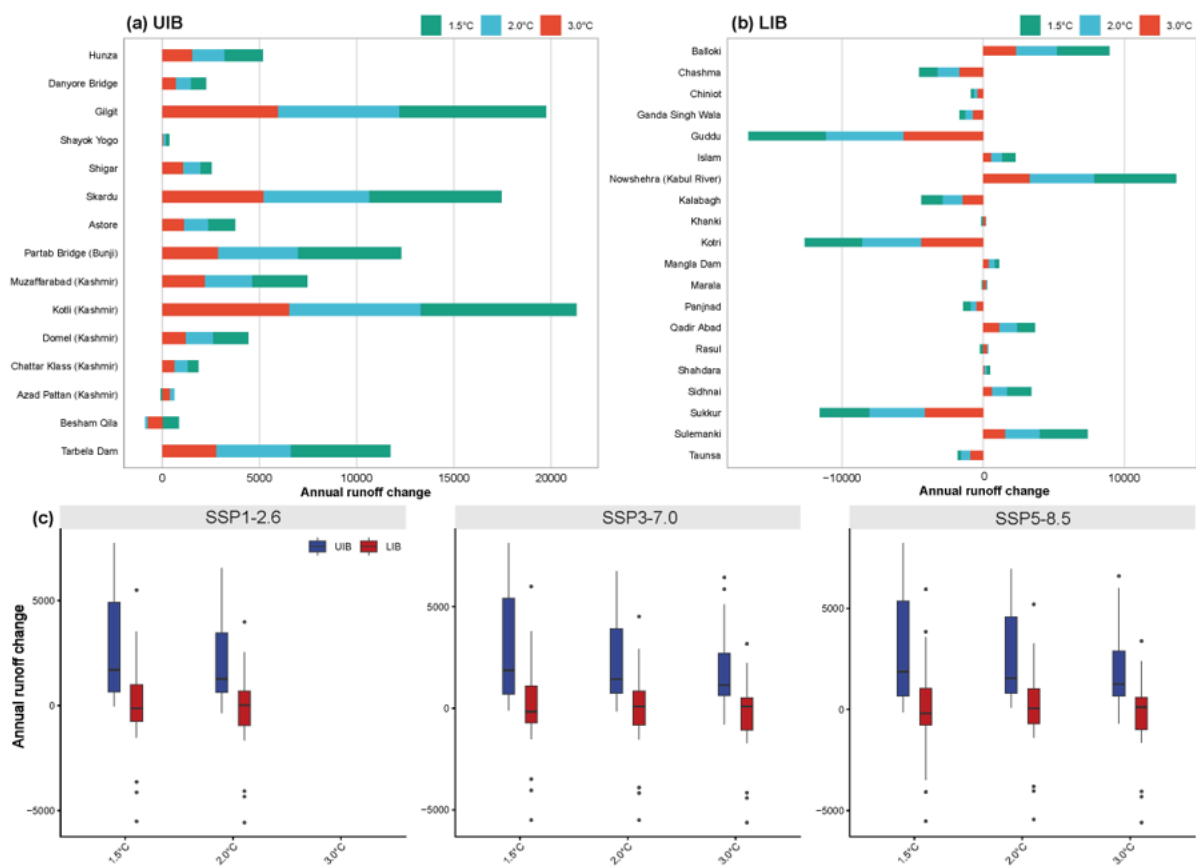


Figure 3

Annual runoff response across GWLs in the UIB and LIB. (a, b) Station-scale annual runoff change (m³/s) in the UIB and LIB under 1.5°C, 2.0°C, and 3.0°C warming levels. (c) Station-wise distribution of annual runoff change under SSP1-2.6, SSP3-7.0, and SSP5-8.5, shown as boxplots for the UIB and LIB. The figure highlights pronounced spatial heterogeneity in runoff response, with consistently positive and strengthening signals in cryosphere-influenced UIB stations, contrasted with mixed to strongly negative responses across most LIB stations.

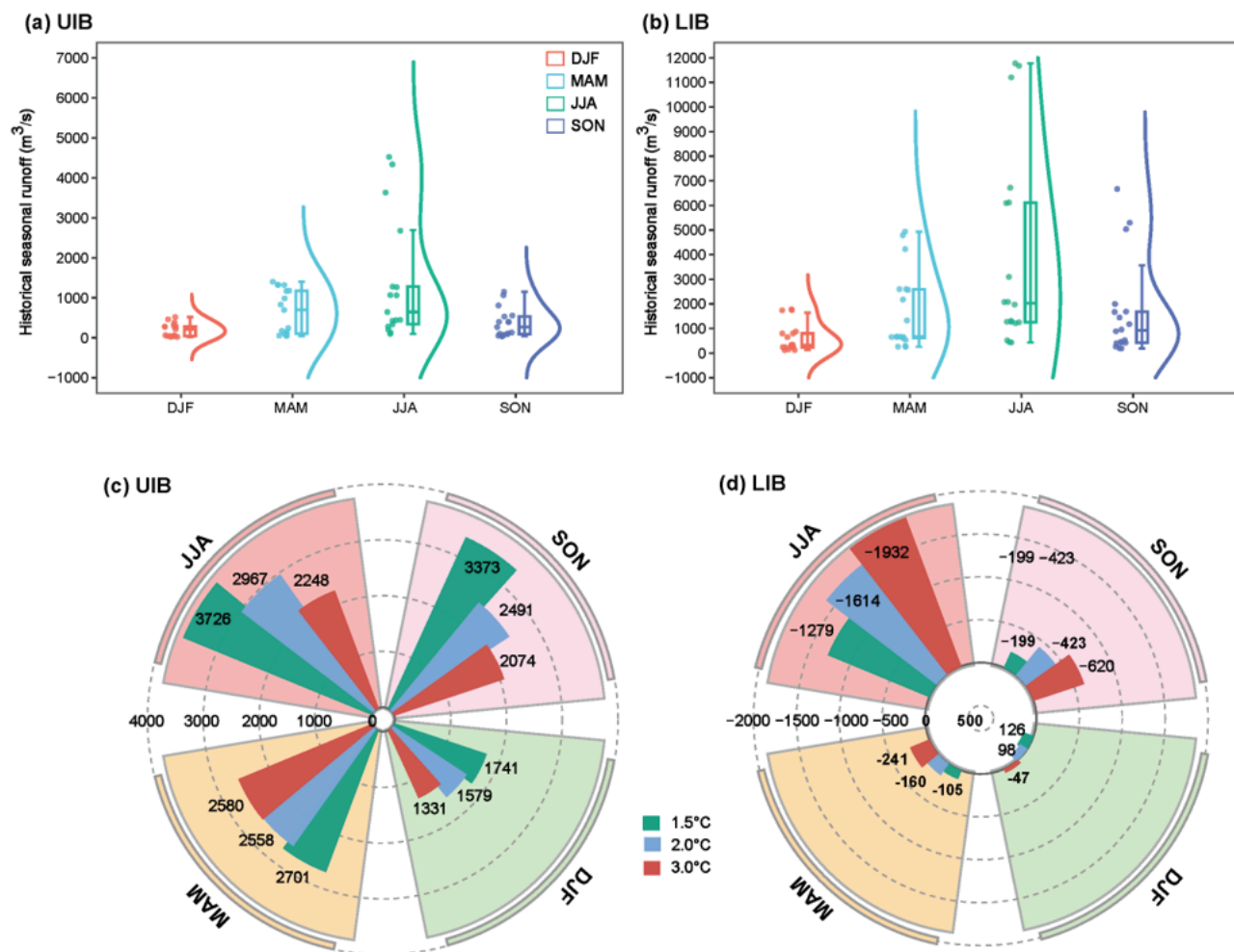


Figure 4

Seasonal runoff baseline and warming-induced seasonal anomalies in the UIB and LIB. (a, b) Historical seasonal runoff distributions across stations in the UIB and LIB for DJF, MAM, JJA, and SON, shown using half-violin density plots, boxplots (median and IQR), and station-level observations. (c, d) Basin-scale seasonal runoff anomalies relative to the historical baseline under 1.5°C, 2.0°C, and 3.0°C warming levels for the UIB and LIB. Results indicate strong warm-season amplification in the UIB, particularly in JJA and SON, while the LIB exhibits weaker and seasonally contrasting responses, including persistent negative anomalies in multiple seasons.

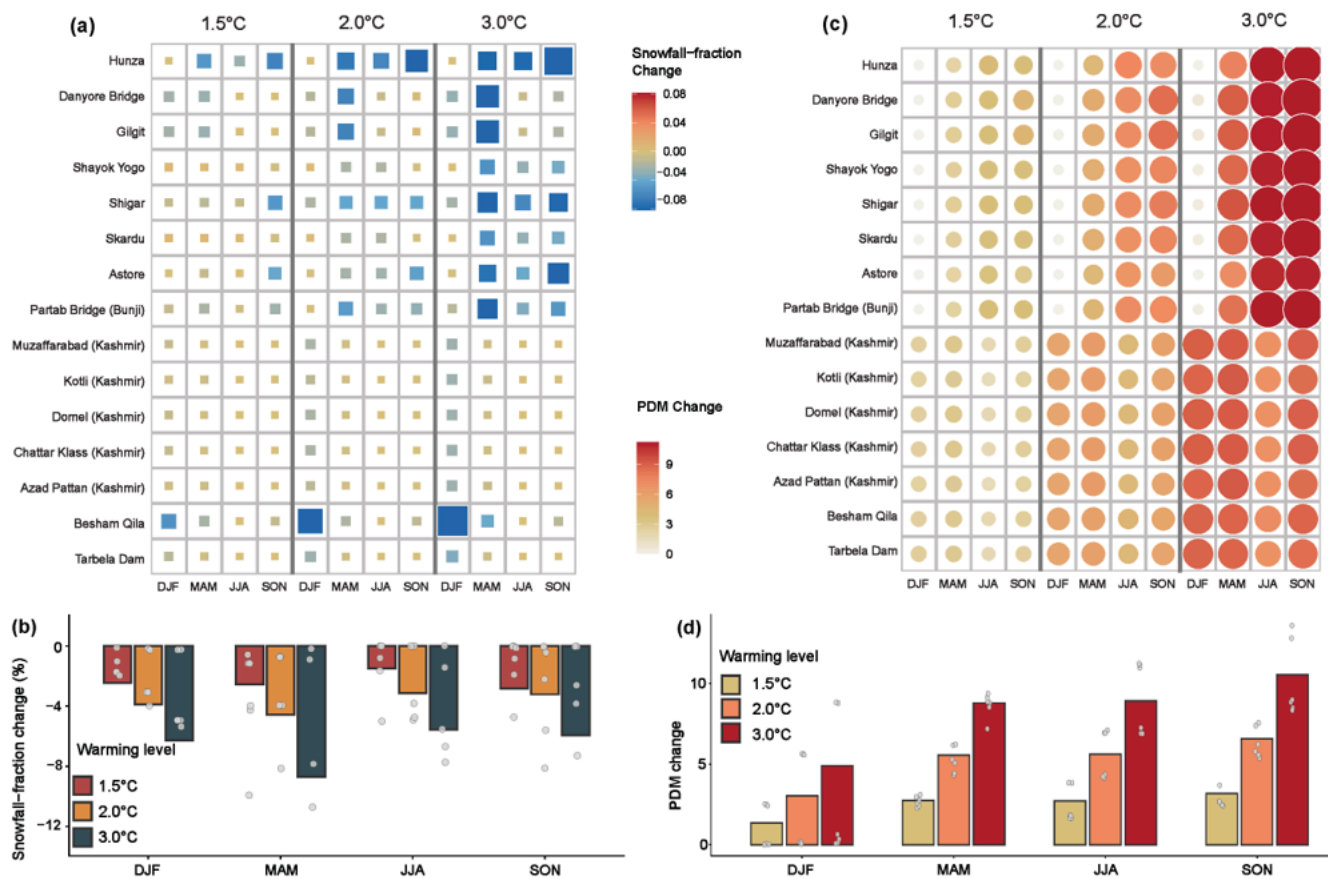


Figure 5

Cryosphere-related hydroclimatic response in the UIB under 1.5°C, 2.0°C, and 3.0°C warming levels. (a, b) Station-level changes in snowfall fraction across UIB stations, aggregated seasonally. (c, d) Corresponding changes in PDM across stations and seasons. In (a) and (c), bubble-matrix plots depict station responses, where symbol size indicates magnitude and color denotes direction and intensity. In (b) and (d), grouped bars summarize seasonal mean responses, with overlaid points showing inter-station variability. Results indicate a consistent decline in snowfall fraction with warming, reflecting reduced solid precipitation contribution, alongside a systematic increase in PDM, indicating enhanced thermal conditions favoring snow and glacier melt.

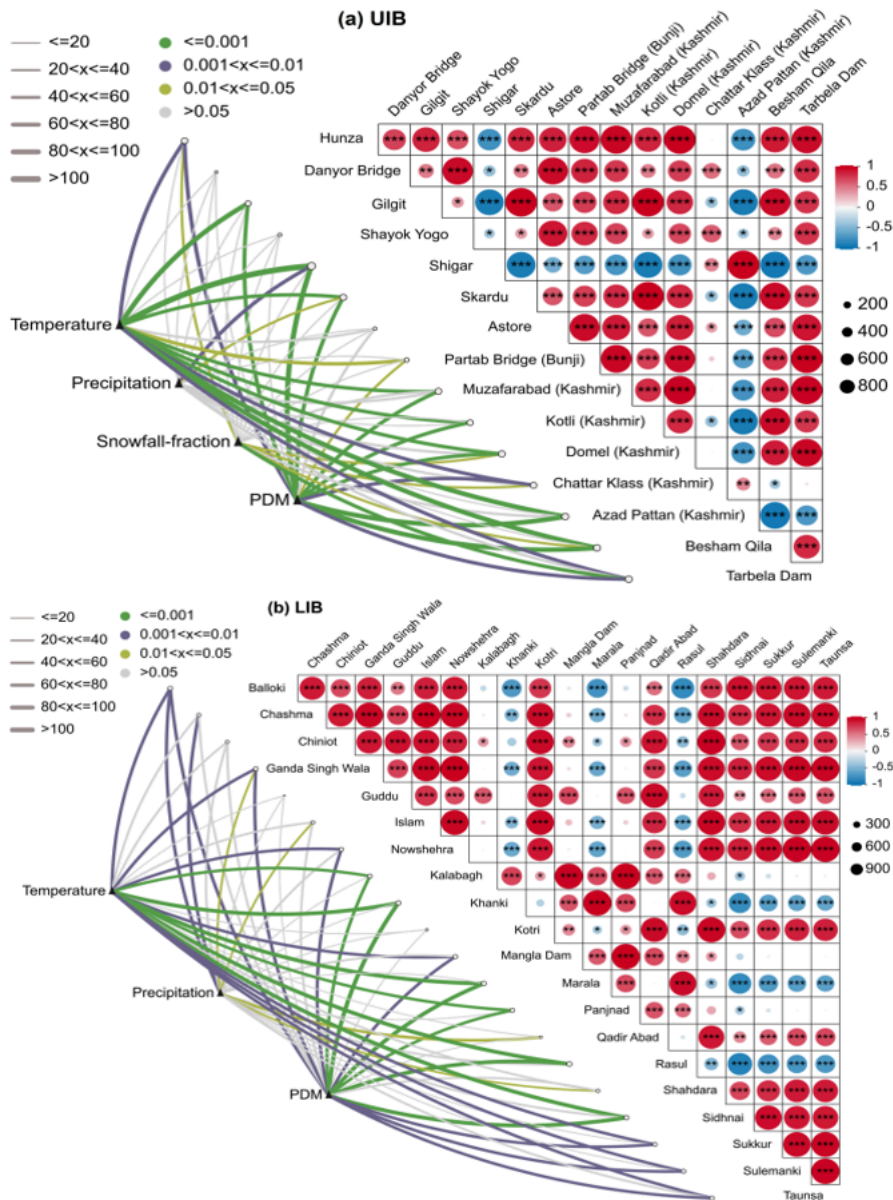


Figure 6

Network-based assessment of runoff controls and inter-station response coherence in the UIB and LIB. (a, b) UIB and LIB results, respectively. In each panel, the left network illustrates station–driver relationships, with line color indicating statistical significance and line thickness representing association strength. The right upper-triangular matrix shows pairwise Spearman correlations of station runoff responses, where bubble size and color denote correlation magnitude and sign, respectively, and asterisks indicate statistical significance. The figure integrates dominant climatic controls and spatial coherence patterns of runoff response within each basin.

Supplementary Files

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