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Spatial analysis of precipitation time series over the Upper Indus Basin

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Abstract The upper Indus basin (UIB) holds one of the most substantial river systems in the world, contributing roughly half of the available surface water in Pakistan. This water provides necessary support for agriculture, domestic consumption, and hydropower generation; all critical for a stable economy in Pakistan. This study has identified trends, analyzed variability, and assessed changes in both annual and seasonal precipitation during four time series, identified herein as: (first) 1961–2013, (second) 1971–2013, (third) 1981–2013, and (fourth) 1991–2013, over the UIB. This study investigated spatial characteristics of the precipitation time series over 15 weather stations and provides strong evidence of annual precipitation by determining significant trends at 6 stations (Astore, Chilas, Dir, Drosh, Gupis, and Kakul) out of the 15 studied stations, revealing a significant negative trend during the fourth time series. Our study also showed significantly increased precipitation at Bunji, Chitral, and Skardu, whereas such trends at the rest of the stations appear insignificant. Moreover, our study found that seasonal precipitation decreased at some locations (at a high level of significance), as well as periods of scarce precipitation during all four seasons. The observed decreases in precipitation appear stronger and more significant in autumn; having 10 stations exhibiting decreasing precipitation during the fourth time series, with respect to time and space. Furthermore, the observed decreases in precipitation appear robust and more significant for regions at high elevation (>1300 m). This analysis concludes that decreasing precipitation dominated the UIB, both temporally and spatially including in the higher areas.

1 Introduction

The Indus River passes through an enormous region of mountains in Asia and provides one of the foremost water reserves of Pakistan as it emerges from the Tibetan Plateau and the Himalayas. More than 60 % of this area lies within Pakistan and encompasses a drainage area of nearly 966,000 km^2. This includes a basin contributing its six major tributaries: the Sutlej, Ravi, Beas, Jhelum, Chenab, and Kabul rivers, from the four bordering countries (Afghanistan, China, India, and Pakistan). Sharing of the Indus River water has sometimes been quarrelsome between India and Pakistan, since Pakistan declared its independence in 1947. Irrigation systems of both countries depend on the water from the Indus and its tributaries, both for domestic consumption and hydropower. Irrigated lands within the Indus basin produce wheat, rice, and sugarcane, completely dependent on the river water, because of scarce precipitation during the sowing and harvesting periods. Hydropower generation from the Indus at Tarbela supplies nearly 20 % of Pakistan's national electricity demand, exemplifying the river's importance, both locally and countrywide. An assessment regarding the potential impacts of climate change on water resources over the Upper Indus Basin (UIB) would be helpful to understand (and potentially resolve) water issues in the coming decades, with particular

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importance for the downstream areas (Khan et al. 2014; Zentner 2012; Archer et al. 2010; Barnett et al. 2005).

Human-induced releases of greenhouse gases to the atmosphere directly affect the hydrological cycle and cause global warming (Arnell 1999; Allen and Ingram 2002; Alan et al. 2003). The spatial and temporal distributions of precipitation can respond to this accelerated cycle in the form of frequent or extreme precipitation events, which can result in floods (and droughts) globally and regionally (including areas within China: Mirza 2002; Easterling et al. 2000; Zhang et al. 2010, 2011). Warming of the global climate system is likely and it can affect sea level both by the melting of perennial snow and through thermal expansion of the oceans. Furthermore, this will raise the global mean annual sea surface temperatures (IPCC 2007, 2013). Moreover, high northern latitude glaciers will retreat and much of the land will suffer from frequent heat waves and heavy precipitation events. Several causes exist for recent and future climate changes, on both temporal and spatial scales. Firstly, snow-covered mountains are particularly sensitive to climate change, due to high troposphere and spatial variation. Secondly almost one sixth of the world's population depends on water flowing downstream from mountain glaciers and their associated precipitation (Bradley et al. 2006; Beniston et al. 1997; Barnett et al. 2005; Viviroli et al. 2007). Scientific evidence strongly supports that the predicted retreat of snow and ice will have a remarkable impact on the downstream areas, as warming proceeds at the rate of 0.01 °C/year in the eastern Hindukush–Karakoram–Himalaya (HKH). Climate is changing globally due to increased levels of greenhouse gases in the air (which diminishes the energy reflected back into space), resulting in increasing temperatures and changes in the circulation of precipitation (Shrestha and Devkota 2010; Houghton et al. 1995; Yue and Hashino 2003).

The Tibetan Plateau and Himalayas provide the sources of many Asian Rivers, such as the Indus (Pakistan), the Brahmaputra and Ganges (India), the Salween and Irrawaddy (Burma), and the Mekong, Yellow, and Yangtze Rivers (China); all of which originate from the Tibetan Plateau and the Himalayas. Preservation of these water resources is also essential because millions of people living downstream depend on these water resources (Immerzeel 2008a). The Tibetan Plateau affects the water accessibility downstream in the major river basins of Asia, largely because precipitation in the upper portions of the drainage falls largely as snow, which then discharges the liquid runoff to the rivers throughout the year. The melt season definitely affects the snow cover dynamics and the availability of water in downstream areas (Barnett et al. 2005; Immerzeel 2008b).

Fowler and Archer (2005) and Tahir et al. (2011) enlightened the Himalaya and Karakoram region by reporting that its monsoon dynamics also determine the pattern of precipitation over the Himalayas and that the westerly circulation transports its maximum precipitation in winter, as snow. Akhtar et al. (2008) used a regional climate model and observed increases of 4.8 °C in the annual mean temperature, and up to 16 % in annual mean precipitation at the completion of the twenty-first century (2071–2100). Bocchiola and Diolaiuti (2013) found decreasing precipitation over the UIB and increasing trends over Chitral and the northwest Karakoram, although most of the precipitation changes were statistically insignificant.

Archer and Fowler (2004, 2008) explored the climatic impact on hydrological regimes in the UIB and divided into three hydrological regimes. First of these (which consists of the Hunza and Shyok watersheds) has a high-elevation catchment, which includes considerable glaciated areas, which produce summer runoff controlled by air temperatures providing the energy to melt these glaciers. They characterize the southern Karakoram (Astore watershed), with mid-elevation catchments, where the summer flows depend on the previous winter's precipitation. Then, rains during the monsoon control the runoff in the foothill catchment areas (Khan et al. 2002).

Spatial distribution of precipitation is affected by complex interactions between the geography and atmosphere of the catchment at smaller scale (Bookhagen and Burbank 2006; Immerzeel 2008b). Generally, stations installed at low altitudes remain dry and amount of precipitation increases up to a definite maximum altitude. According to these authors, HMAX is different for the entire HKH region, i.e., 4500, 5000, and 5500 m for the Hindukush, Karakoram, and Himalayas regions, respectively. After gaining such height, the precipitation decreases again due to removal of moisture from the air (Immerzeel et al. 2015). Due to the higher tropospheric altitude of westerly air flow, HMAX is generally higher over Westerly precipitation-dominating areas (Harper 2003; Hewitt 2005, 2007; Scherler et al. 2011; Winiger et al. 2005). Due to the inaccessibility and restricted available data (≥3000–4500 m) of high-altitude stations over UIB, we were restricted to carry out trends identification with respect to altitude, only at low-elevation stations.

We used the Mann–Kendall (MK) test, which applies broadly to time series analysis, to identify trends. Our trend identification was based on two steps, i.e., determining the statistical significance and directions of the trend with respect to time. Statistical significance was calculated using the MK test, while the shape of the trend was defined by Sen's slope method. The MK test is simple and controls values up to a certain detection limit, but this test has certain defects regarding serial correlations. Various approaches have been proposed to handle the effects of serial correlation, such as prewhitening and variance correction (Mann 1945; Kendall 1975, Yue et al. 2002a; Shao and Li 2011; Khaliq et al. 2009; Rivard and Vigneault 2009).

This study made an assessment of existing precipitation variations related to climate change over the UIB (and its sub-catchments) based on recent data collected over the period

from 1961 to 2013. Additionally, we observed climatic trends over annual and seasonal (3 months) time scales, using four individual data sets and the full data period (1961–2013). We also plotted these trends against elevation to measure variability over stations having high as well as low altitude according to available earliest record at our study area. The first objective was to present a comprehensive understanding, without any loss of annual tendencies caused by the averaging process. Seasonal precipitation was subdivided into 3-month time periods to investigate any inconsistencies. Moreover, we conducted this research to assess spatial effects of climate change, as well as temporal effects, over the UIB. In addition, the trend-free pre-whitening (TFPW) approach was used to eliminate the influences of significant lag-1 serial correlation trend tests. This provided better management and planning, suitable for studies related to climate change for this region. Hopefully, these results can help facilitate local and regional planning of water resources and help governors select optimum strategies related to water management.

2 Study area

We have selected the UIB to investigate the regional climate trends. The catchment of this basin extends from 33° 40′ to 37° 12′ N latitude and 70° 30′ to 77° 30′ E, longitude. The UIB is a sole region including a complex Hindukush–Karakoram–Himalaya (HKH) territory, diverse physio-geographical features, conflicting signals of climate change, and consequently, conflicting hydrological regimes. This region generally crosses national borders, but shares borders with Afghanistan in the west, with China in the north, and with India in the east. Figure 1 show the UIB watershed boundary derived from the digital elevation model (DEM) at the convergence of the Kabul and Indus Rivers, at the head of the Khairabad and Attock districts. The elevations of the watershed vary from 254 to 8570 m. According to SWHP, WAPDA the estimated area of UIB at Besham Qila is around 162,393 km². The Indus River flows from the north side of the Himalayas at Kaillas Parbat in Tibet at an elevation of 5500 m and is fed by a number of small rivers, which contribute water to
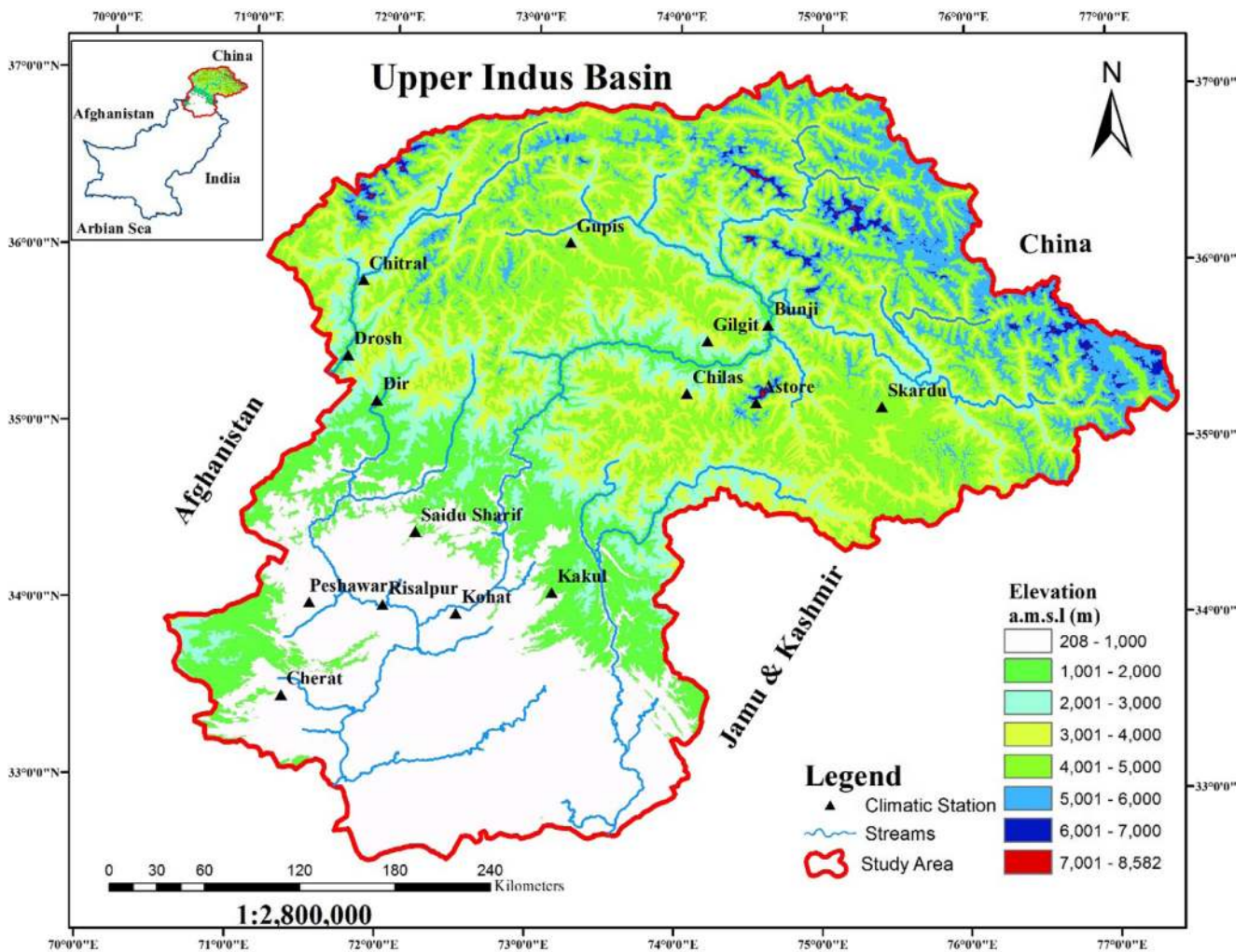

Fig. 1 Upper Indus Basin (UIB) and meteorological stations network

the main Indus River. The sub basins include the Chitral, Swat, Kabul, Hunza, Gilgit, Astore, Shigar, Shyok, Kunhar, Neelum, Kanshi, Poonch, Soan, Siran, Sil, and Haro. Traversing about 8000 m toward the north west (NW), the Indus is joined by the Shyok River near Skardu (elevation 2700 m). After traveling about 1600 m in the same direction, it reaches the Nanga Parbat area and joins by the Gilgit River at an elevation of 1524 m. Flowing about 3200 m further in a South West (SW) direction, the river enters into the plains of the Punjab province (Pakistan) at Kalabagh (800 ft). The westerly disturbances, which control the precipitation regime instigated at Atlantic Ocean and Caspian seas, transported over the UIB by the Mediterranean storm tracks (Bengtsson et al. 2006). Moreover UIB also accumulates summer monsoonal precipitation in the form of solid at higher altitudes while in the form of liquid at lower altitudes crossing the Greater Himalayas. (Wake 1989; Ali et al. 2009; Hasson et al. 2015; Archer and Fowler 2004). According to Randolph Glacier Inventory version 4.0 (RGI4.0 – Pfeffer et al. 2014) the total area covering glaciers and perennial ice cover is around 18,500 km^2 (11 %) of the aggregate surface area of the basin. Pakistan covering almost 46 % of the UIB and around 60 % of this area is cryosphere.

3 Data

This study chose the 15 weather stations having the earliest and most precise records of precipitation. These stations range in elevation from 2317 m (Skardu) to 320 m (Peshawer), and cover most of the key areas of the UIB. These meteorological stations

installed within UIB cover an east–west extent of around 200 km from Skardu to Gupis and north–south range of around 100 km from Gupis to Astore (Hasson et al. 2015). For the present study, all precipitation data is acquired from the Pakistan Meteorological Department (PMD). This organization is national level government agency which is solely responsible for recording the synoptic meteorological variables. PMD is the member of World Meteorological Organization (WMO) and implement all the standard operating procedures (SOPs) related to installation of instruments and acquisition and dissemination of data to end users. The data used in this study is the only available data for the region over a 53-year period. Further, the topography of UIB is mostly mountainous, and as such, very little part of it is accessible for human inhabitance. In recent years, a score of automatic weather stations have been installed to fill the gaps but the data time series is for a short period of time. Table 1 provides the specifications for these stations. Figure 1 shows the locations of these weather stations. Monthly precipitation amounts were computed to investigate trends, and any missing data were approximated by linear interpolation. Our study calculated trend analyses from daily precipitation observations for four different periods: 1961–2013, 1971–2013, 1981–2013, and 1991–2013.

4 Materials and methods

4.1 Selection of variables

Climatic and hydrologic variables provide important climate indicators. These variables indicate climatic changes and help us

Table 1 List of climatic stations used in the present study and their characteristics, normal for the period (1971–2000)

Sr. no.	Station	Lat.	Lon.	Elevation	Maximum, minimum, and mean annual temperatures (°C) and annual precipitation (mm)			
		(dd)	(dd)	(m)	Max.	Min.	Mean	P_{mean}
1	Astore	35.2	74.5	2168.0	15.6	4.1	9.84	480
2	Bunji	35.6	74.6	1372.0	23.8	11.4	17.6	164
3	Cherat	33.5	71.3	1372.0	21.5	13.2	17.3	155
4	Chilas	35.3	74.1	1250.0	26.4	14.1	20.1	190
5	Chitral	35.9	71.8	1497.8	23.3	8.6	15.9	458
6	Dir	35.2	71.9	1375.0	22.9	8	15.4	1356
7	Drosh	35.4	71.7	1463.9	24.1	11.3	17.6	586
8	Gilgit	35.6	74.2	1460.0	23.9	7.6	15.8	136
9	Gupis	36.1	73.2	2156.0	18.7	6.7	12.6	196
10	Kakul	34.1	73.2	1308.0	22.8	10.9	16.9	1329
11	Kohat	34.0	72.5	1440.0	29.3	16.9	23.3	589
12	Peshawar	34.0	71.5	320.0	29.5	16.1	22.8	482
13	Risalpur	34.0	72.0	575.0	29.6	14.5	21.9	670
14	SaiduSharif	34.4	72.2	961.0	26	12	19.2	1068
15	Skardu	35.2	75.4	2317.0	18.6	4.9	11.7	223

understand the relationships between hydrology and climate. This study used annual precipitation as prime indicator, because the total annual precipitation amounts strongly associate with extreme precipitation events that can be evident through different precipitation indices (Gong et al. 2003; Yue et al. 2011; Wang et al. 2013a, b). Moreover, we also observed seasonal variability to measure inter seasonal inconsistency in the amount of precipitation.

4.2 Selection of stations

The selection of stations to be included in a study is one of the most important steps in climate change research. This study's selection of stations was based on several criteria: (1) a station must have a minimum climate record of 20 years; (2) the record accuracy has been assured by unimpaired basin conditions affecting the monthly precipitation; (3) there must have been no overt adjustment of rainfall (e.g., cloud seeding); (4) no interference near measurement points by control structures; and (5) there has been no change in land use that could have significantly affected the amounts of monthly precipitation or runoff.

The rain gauges in most of the observations within the UIB are surrounded by appropriate fencing structure while in newly functional observatories, automatic data recording systems with wind shields were installed at the same height of the orifice of the gauge. The error due to wind field deformation is calculated under the following procedure.

$$Pk = kPc + \text{other losses} \qquad (1)$$

Where Pk is the adjusted precipitation amount, k is the adjustment factor for the effects of wind field deformation, and Pc is the amount of precipitation caught by the gauge collector. The k factor is determined from the ratio of "correct" to measured precipitation for rain plotted on the graph for two unshielded gauges in dependency of wind speed u and intensity i. The k factor is usually defined on a monthly basis or seasonwise, depending upon the local climatology of the station; but for all practical purposes, k factor is provided in a form of catalog for different wind intensities in gauge sites where the technical staff is not competent enough to calculate it for themselves. The same standard is adopted for snow with different scale as the wind error factor almost reduces by 50 % for gauge shielded by wind protectors. Hewitt (2011) reported that most of the ice reserves were found above 2200 m a.s.l. and Karakoram Range. The hydrology of the selected stations for the present study is mostly controlled by liquid precipitation as compared to Karakoram or stations lying above 2500 m a.s.l., which are dominated by solid moisture input.

4.3 Detection of trends

The objective of trend identification is to determine whether the values of a variable of interest mostly increase or decrease over some period of time, statistically (Haan 1977). Evaluation of whether a trend is statistically significant or not can be accomplished with the help of parametric or non-parametric statistical tests. The fields of hydrology and climatology typically adopt non-parametric approaches for determining the magnitude of any trends. Previous studies (Dahmen and Hall 1990) used Mann–Kendall (MK) test in combination with the Sen's slope estimator (SS) (Sen 1968) for the detection and magnitude of the trends, respectively. Yue et al. (2002a) reported that there is no noticeable differences between the results of MK and Spearman rank correlation tests. MK test is affected by autocorrelation of the time series being observed which is considered as a serious issue, to overcome this problem various approaches has been introduced in the literature such as the pre-whitening (PW) and trend-free pre-whitening (TFPW) techniques. We employed MK- and SS-based tests in combination with the (TFPW) for the present study to identify trends in the precipitation. This study analyzed the time series of annual and seasonal precipitation; and these steps essentially involved testing and trend detection of any serial correlation effects by applying the Mann–Kendall test.

4.3.1 Serial correlation

Comprehensive time series analyses require consideration of autocorrelation or serial correlation (defined as the correlation of a variable with itself over successive time intervals), prior to testing for trends. Specifically, if a positive serial correlation exists in the time series, then the non-parametric test will suggest a significant trend in a time series that may actually be more often random than specified by the significance level (Kulkarni and Von Storch 1995). For this, Von Storch and Navarra (1995) suggested that the time series should be "pre-whitened" to eliminate the effect of serial correlations before applying the MK test or any trend detection test. Yue et al. (2002a) showed that removal of serial correlations by pre-whitening can effectively remove the serial correlation and eliminate the influence of the serial correlation on the MK test. Yue and Wang (2004) modified the pre-whitening method to apply to a series in which there was a significant serial correlation. The trend-free pre-whitening method has been applied in many of the recent studies to detect trends in hydrologic and meteorological parameters (e.g., Yue and Hashino 2003; Aziz and Burn 2006; Novotny and Stefan 2007; Kumar et al. 2009; Oguntunde et al. 2011). This study incorporates this approach, which thereby made it possible to identify statistically significant trends in precipitation observations ($x_1, x_2 \dots x_n$), which were examined using the following procedures:

1. For a given time series of interest, the slope of the trend (β) was estimated by using the Sen's robust slope estimator method. This de-trended the time series by assuming a linear trend as:

$$Y_i = x_i - (\beta \times i) \tag{2}$$

where (β) is the slope, x_1 denotes the sequential data, and n denotes the data length.

2. Compute the lag-1 serial correlation coefficient (designated by r_1).

3. If the calculated r_1 is not significant at the 5 % level, then the statistical tests are applied to original values of the time series. If the calculated r_1 is significant, prior to application tests, then the pre-whitened time series were determined as:

$$\overline{Y_i} = Y_i - rY_{1-i} + (\beta \times i) \tag{3}$$

where Y_i is original time series.

4.3.2 The Mann–Kendall test for trend detection

The widely used Mann–Kendall test measures statistical trends in the climatologic analyses (Tabari et al. 2012; Caloiero et al. 2011; Mavromatis and Stathis 2011; Bhutiyani et al. 2007; Rio et al. 2005), and in the hydrologic time series (Yue and Wang 2004). This last test provides two advantages: first, being a nonparametric test, it does not require the data to be normally distributed, and second, the test has low sensitivity to abrupt breaks caused by inhomogeneous time series data (Tabari et al. 2011).

The Mann–Kendall statistic, Z_{mk}, was evaluated as follows:

$$Z_{mk} = \begin{cases} \dfrac{S-1}{\sigma_s} \text{ if } S > 0 \\ 0 \text{ if } S = 0 \\ \dfrac{S+1}{\sigma_s} \text{ if } S > 0 \end{cases} \tag{4}$$

The MK test statistic S can be computed from the relationship shown below:

$$S = \sum_{k=1}^{n-1} \sum_{j=k+1}^{n} \text{sgn}(X_j - X_k) \tag{5}$$

where n is the number of years, x_j and x_k are the annual values in the years j and k, respectively. The function $\text{sgn}(x_j - x_k)$ provides an indicator function that takes the value 1, 0, or -1 according to the sign of the difference $(x_j - x_k)$; where $j > k$:

$$\text{sgn}(X_j - X_k) = \begin{cases} 1 \text{ if } X_j - X_k > 0 \\ 0 \text{ if } X_j - X_k = 0 \\ -1 \text{ if } X_j - X_k < 0 \end{cases} \tag{6}$$

A positive value of Z_{mk} indicates an upward (warming) trend, while a negative value shows a downward (cooling) trend. The test statistic (S) follows the standard normal distribution. Therefore, if the probability under the null hypothesis (H_o) of observing a value higher than the test statistic Z_{mk} for a chosen significance level α, this indicates a statistically significant trend. The null hypothesis H_o is considered true if there is no trend, and thus uses the standard normal table to decide whether to reject H_o. To test for either an upward or downward trend (a two-tailed test) at α level of significance, H_o is negative if the absolute value of $Z_{mk} > Z_{1-a/2}$ at the α-level of significance.

4.4 Sen's slope estimator

The changes over each period were computed from the slope estimated by Sen's method (Sen 1968). Changes over time were estimated relative to mean daily precipitation over the period. We used relative units, since precipitation differs significantly from one gauge to another, and absolute changes in precipitation are difficult to compare. Daily precipitation values, and especially their changes, often have very low magnitudes, and are less indicative for intercomparison.

The slope estimates of N pairs of data were first computed using the following formula:

$$Q_i = \frac{x_j - x_k}{j - k} \quad \text{if } j > k \tag{7}$$

where x_j and x_k represent the annual values throughout the years j and k, respectively. The Sen's estimator of slope provides the median of these N values of slope (Q). The median of the N slope estimates was obtained in by simple averaging. N values of Q_i were ranked from smallest to largest and the Sen's estimator was computed as follows:

Sen's estimator

$$= \begin{cases} Q_{[\frac{N+1}{2}]} \text{ if } N \text{ was odd} \\ \dfrac{1}{2}\left(Q_{\frac{N}{2}} + Q_{[(N+2)/2]}\right) \text{ if } N \text{ was even} \end{cases} \tag{8}$$

Finally, Q_{med} was tested using a two-sided test at the 100 $(1 - \alpha)$ % confidence interval, and the true slope may be obtained by the non-parametric test.

5 Results

Results of trend analyses depend on the study period chosen. We first expanded our study period to 53 years for the first time series, and then contracted it to 43, 33, and 23 years,

sequentially. This allowed our determinations of the individual characteristics of each time series. If the study periods were extended to longer or different time periods, somewhat different conclusions might be drawn (Bae et al. 2008). Our study determined trend analyses and changes in precipitation among the various UIB stations within different regions, and over different time periods. Our analyses calculated increasing and decreasing trends (as well as significant/insignificant trends) with annual and seasonal time series, with the number of stations being shown in Table 2. To examine the spatial consistency of the observed trends, our study created maps displaying the areas with decreasing and increasing trends. Figures 2, 3, 4, 5, 6, 7 show the spatial distributions of precipitation and changes in annual and seasonal trends.

5.1 Trends in annual and seasonal precipitation

The climatic conditions in the portion of the UIB located within Pakistan varied greatly in terms of precipitation. Figure 2 presents a summary of the results from our trend analyses, by portraying data from 15 climatic stations in the UIB. Most of the stations showed decreasing precipitation during all data series. We note a significant pattern of decreased annual precipitation (Table 2), which shows that most of the weather stations experienced decreasing trends, as six stations showing decreasing trends out of which four stations exhibit significant decreasing trends. Similarly 14 (6**), 5 (5*), and 7 (6*) stations; for the first to fourth data series, respectively. Only during the second data period do 6 out of 14 stations exhibit significant increasing trends. During the last data series (1991–2013), six out of seven stations showed significant negative trends, indicating strong and significant trends of increased dryness. Similarly, the number of stations showing significant decreasing trends for the spring precipitation are observed as 8, 5, 7, and 8, respectively, for the first, second, third, and fourth data period. Similarly autumn

precipitation exhibited significant negative trends, at 8, 6, and 10 stations during the first, second, and third data periods, respectively. We note increased precipitation only in the second data series (1971–2013), during which significant increasing trends occurred at 8 out 15 weather stations. We noted the number of stations exhibiting negative significant trends during summer (JJA) as 3, 3, 6, and 8 stations during the first to fourth data periods. Number of during winter (DJF) we note most of the stations revealed significant increasing trends at 5, 5, and 2 during the first to third data series; only three stations showed significant decreasing trends during (1991–2013). Our study strongly suggests considerable variability in the annual and seasonal precipitation time series, particularly for the period from 1961 to 2013. Annual precipitation records indicate gradual decreases in the number of stations showing significant positive trends at the rate of 47 to 13 % stations during the first three data series. These results provided the basis for the findings of Hasson et al. (2015) (regarding increased drying between 1995 and 2012 of annual precipitation).

5.2 Spatial distribution of annual and seasonal precipitation

Figure 3 shows the spatial distribution of the annual trends for each meteorological index, both in terms of direction (sign) and statistical significance, as were synthetically represented at Astore, Bunji, Gilgit, and Saidu Sharif. These distributions show significantly decreasing precipitation at the rates of 0, 2, 1, and 1 % per decade, respectively, for the first data series. Risalpur showed significantly increased precipitation, at the rate of 1 % per decade. The second data series showed almost 50 % of the stations with significant increases in precipitation at Bunji, Chilas, Dir, Drosh, Kakul, Skardu, and Saidu Sharif at rates of 5, 6, 1, 3, 4, 3, and 2% per decade, respectively. Moreover, Astore, Chilas, Dir, Drosh, and Kakul showed decreased precipitations, at rates of 6, 5, 0, 11, and 5 % per decade, respectively, between 1981 and 2013. Likewise, observations from Astore, Chilas, Dir, Drosh, Gupis, and Kakul showed mostly decreasing trends, at rates of 14, 15, 12, 16, 10, and 5 % per decade, respectively. Chitral, Cherat, and Sskardu experienced increased precipitation, at the rates of 5, 5, and 1 % per decade, respectively, during the last data period of 1991–2013.

Our study found large-scale, regionwide, seasonal variations in the amount of precipitation over selected catchments of the UIB. Overall, stronger indications of decreasing trends were found in the seasonal precipitation than in the annual precipitation. In particular, our study found more robust indications of positive, as well as negative, precipitation trends. Every season's distribution of rainfall seemed to differ significantly from the others. The data show significant positive trends during autumn at Skardu, Chilas, and Dir, at rates of 1, 2, and 4 % per decade,
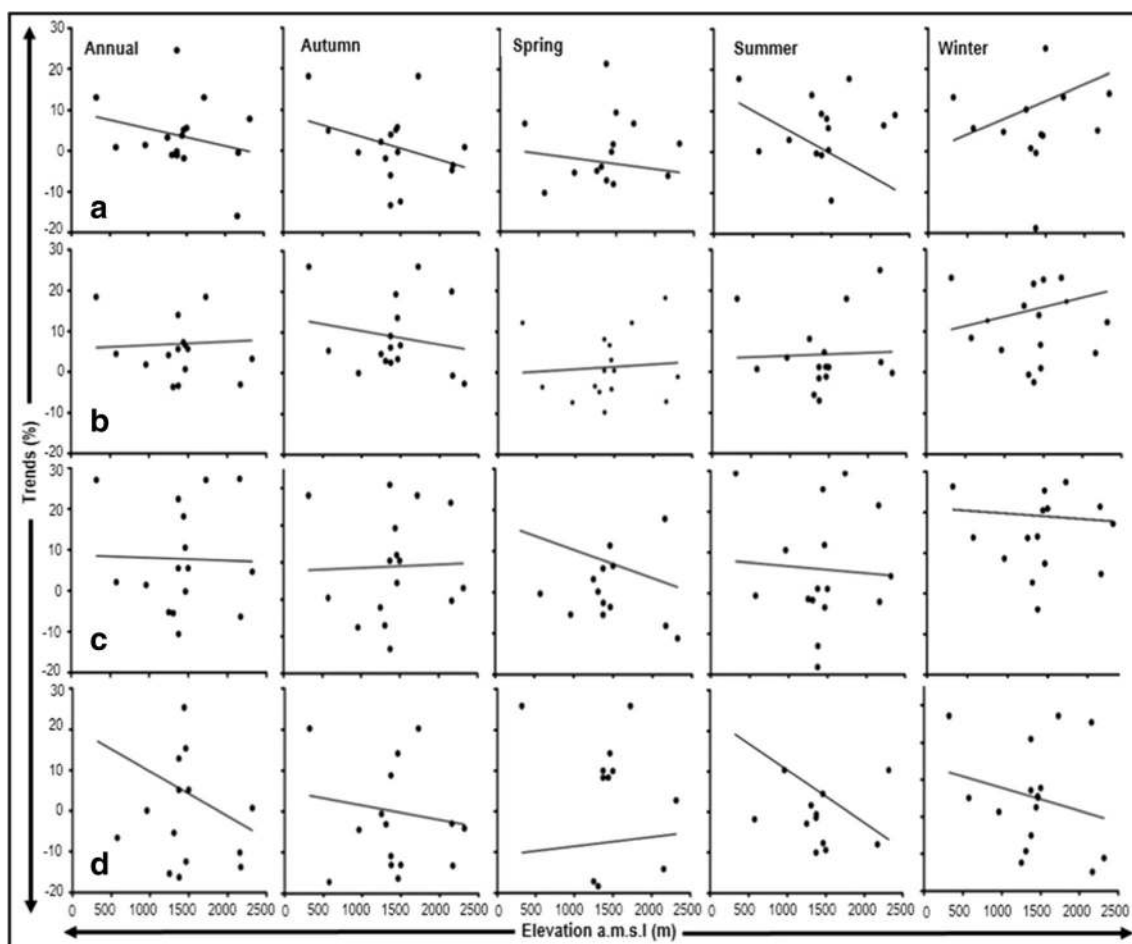
Table 2 Number of stations with positive and negative trends in precipitation time series, for different periods

Variables/months	Periods			
	1961–2013	1971–2013	1981–2013	1991–2013
P_{mean}				
Annual	6 (4*)	14 (6**)	5 (5*)	7 (6*)
Winter (O-M)	5 (5*)	3 (3*)	7 (6*)	8 (7*)
Summer (A-S)	9 (4*)	4 (3*)	8 (6*)	10 (9*)
Winter (DJF)	14 (5**)	15 (5**)	15 (2**)	3 (3*)
Spring (MAM)	10 (8*)	5 (5*)	8 (7*)	10 (8*)
Summer (JJA)	5 (3*)	3 (3*)	7 (6*)	9 (8*)
Autumn (SON)	8 (8*)	15 (8**)	7 (6*)	11 (10*)

*Significant decreasing; **Significant increasing

Fig. 2 Annual and seasonal distributions of precipitation trends plotted against elevation: **a** 1961–2013, **b** 1971–2013, **c** 1981–2013, **d** 1991–2013

respectively. Similarly, most of the stations suffered from significantly decreasing rainfall, notably at Astore, Bunji, Chitral, Cherat, Drosh, Gupis, Kakul, and Saidu Sharif at the rates of 4, 6, 12, 13, 11, 2, and 5 % per decade, respectively, for the first time series. During the second data period, more than 50 % of the locations showed significantly increased precipitation, specifically at Bunji, Cherat, Dir, Drosh, Gilgit, Gupis, Kakul, and Skardu, at rates of 9, 6, 3, 7, 5, 3, and 6 % per decade, respectively. No decreasing trends were found in these data series. The last data series (1991–2013) showed similar increases, with only Bunji and Gilgit showing significant increasing trends, at 9 and 14 % per decade, respectively. Most of the stations exhibited significant decreases during the third data period (including Astore, Bunji, Dir, Drosh, Skardu, and Saidu Sharif, at the rates of 8, 5, 3, 4, 11, and 5 % per decade, respectively, as shown in Fig. 4.

The data collected from these regions exhibited strong evidence of decreasing summer precipitation, as shown in Fig. 5; with most of the stations showing decreasing trends in all the four data series. The amounts of precipitation at Skardu and Astore increased at rates of from 2 to 11 % and from 8 to 12 % per decade in the second and fourth data series, respectively. During

the same period, Astore, Bunji, Chilas, Chitral, Dir, Drosh, and Cherat showed a continuous drying period, with rainfall decreasing from the first to the fourth data period, at the rates of 2, 18, 13, 3, 2, 1, and 1 % per decade, respectively.

Spring also exhibited continuous drying from 1961 to 2013; as Astore, Bunji, Chilas, Dir, Drosh, Kakul, Risalpur, and Saisu Sharif suffered precipitation decreases at rates of 6 to 26, 8 to 23, 5 to 17, 8 to 34, 4 to 18, 10 to 43, and 5 to 33 % per decade, respectively (as shown in Fig. 6). Winter precipitation showed the reverse situation (compared to summer and spring precipitation) as most of the stations experienced increased precipitation during the entire data series. Astore, Dir, Drosh, Gilgit, Kakul, and Saidu Sharif received significantly increasing rainfall, at rates of 5, 10, 4, 5, and 5 % per decade, respectively, during the first to third data series. During the last time series (1991–2013), there were significant decreases in the amount of precipitation at Astore and Dir, at the rates of 5 to 0 % and 0 to −4 % per decade, respectively. Similarly, Skardu and Chilas also showed decreasing precipitation trends, at 1 and 2 % per decade, respectively, as shown in Fig. 7.
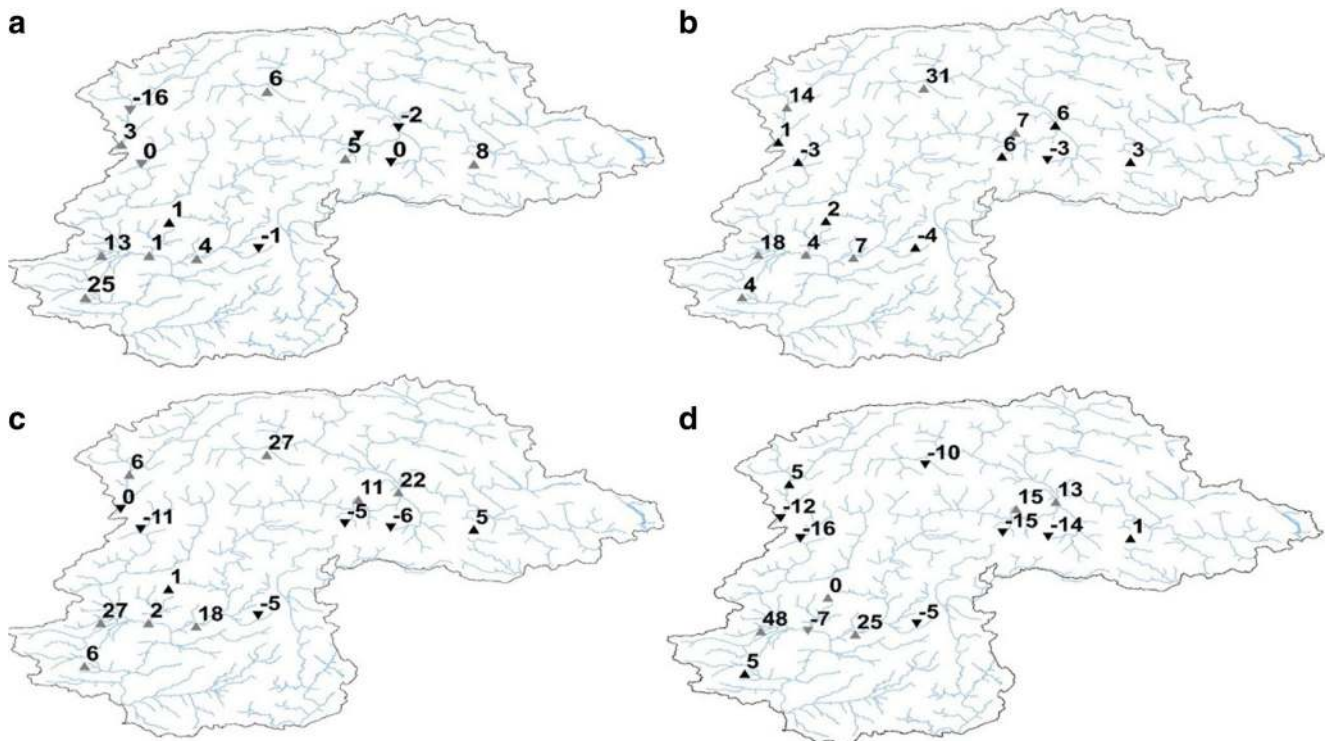
Fig. 3 Spatial distribution of trends detected by Mann–Kendall and trend values estimated by Sen's method of annual precipitation showing change in percentage per decade over time windows in: **a** 1961–2013, **b** 1971– 2013, **c** 1981–2013, and **d** 1991–2013 (*upward and downward arrows* show positive and negative trends, respectively; *bold arrow* shows significant trends at $\alpha = 0.1$)
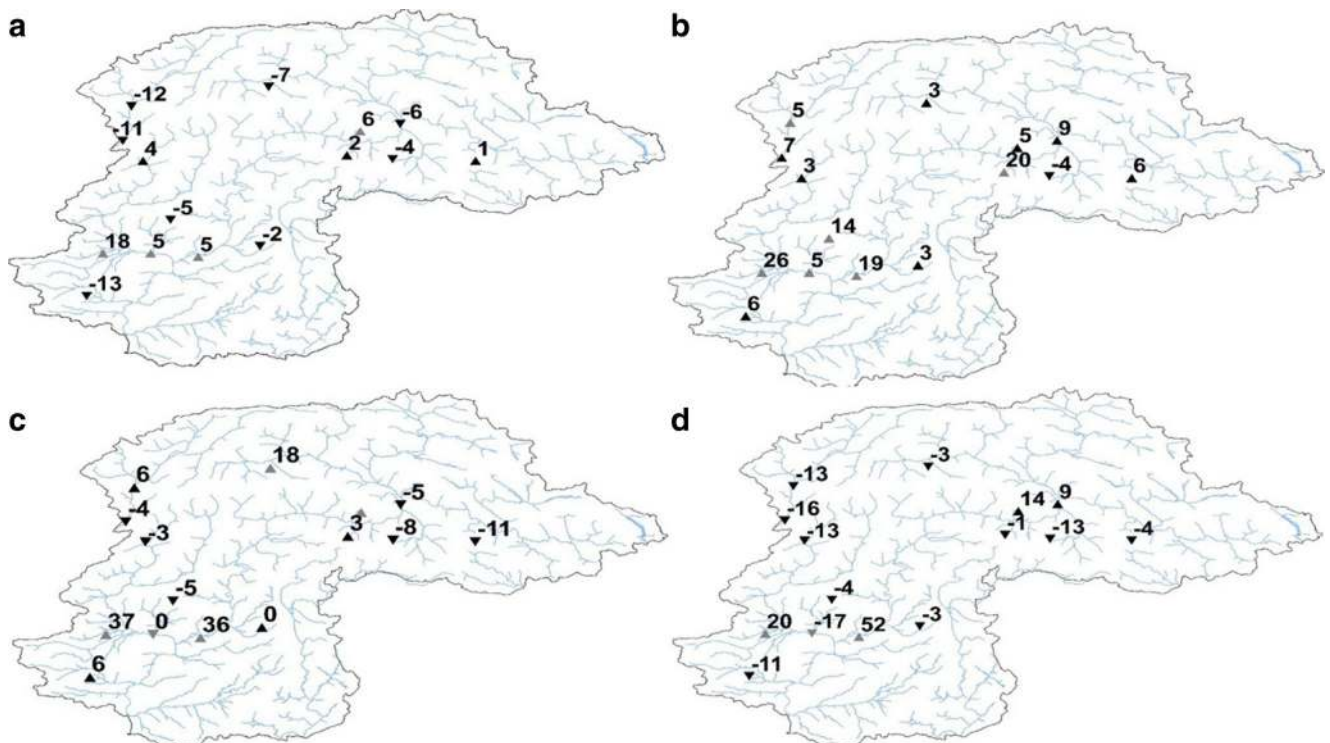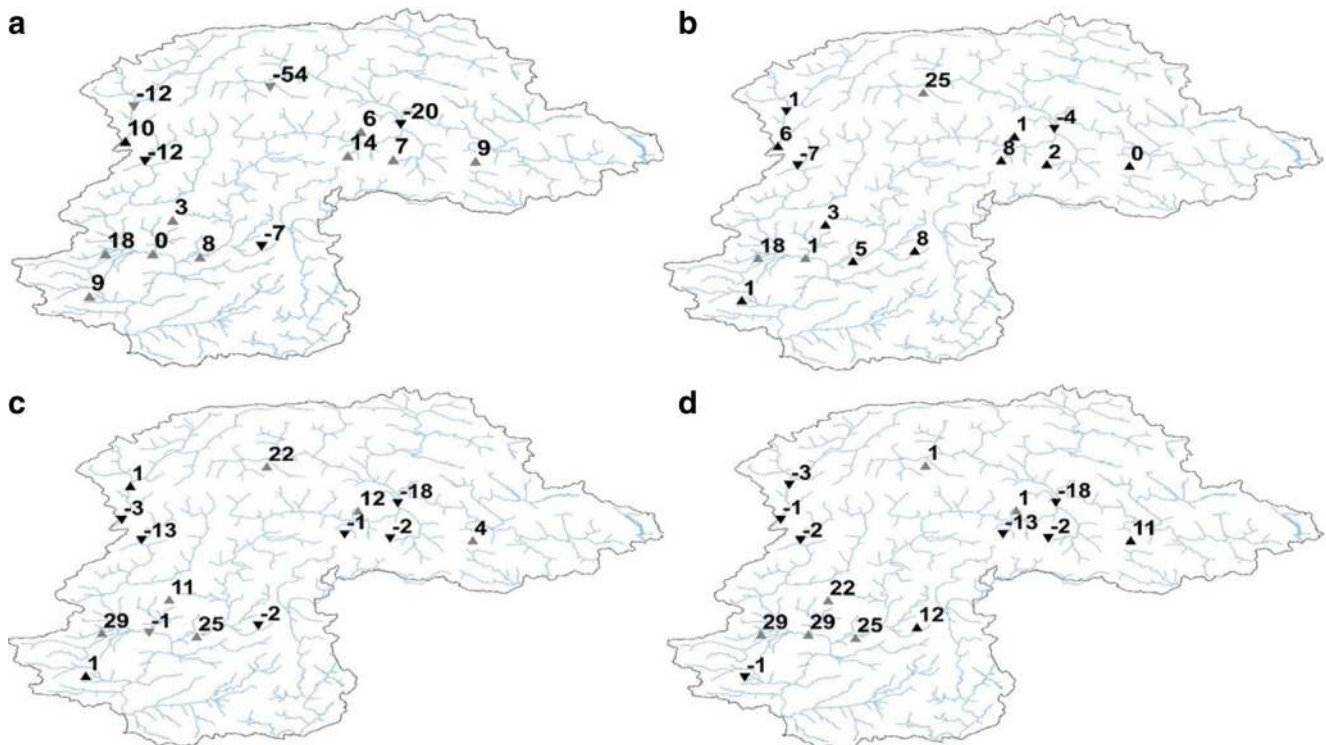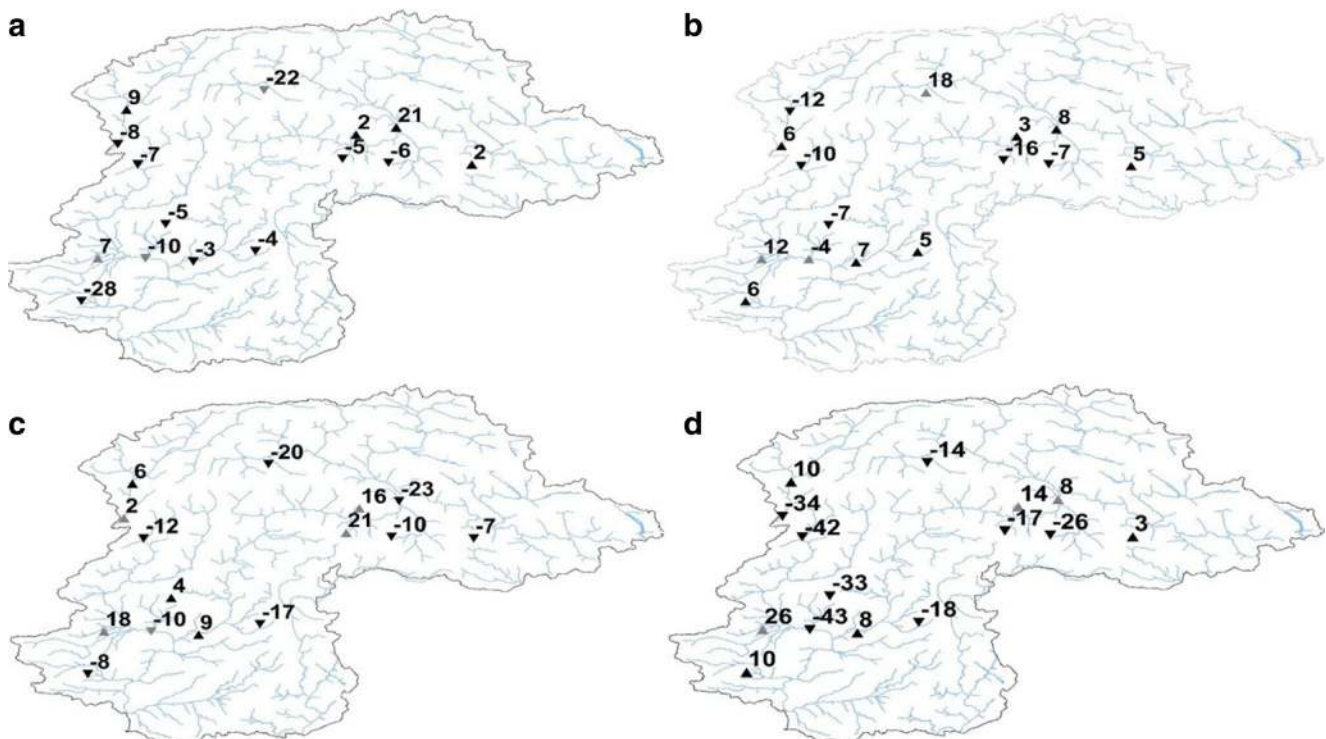

Fig. 4 Spatial distribution of trends detected by Mann–Kendall and trend values estimated by Sen's method of autumn precipitation showing change in percentage per decade over time windows in: **a** 1961–2013, **b** 1971–2013, **c** 1981–2013, and **d** 1991–2013 (*upward and downward arrows* show positive and negative trends, respectively; *bold arrow* shows significant trends at $\alpha = 0.1$)
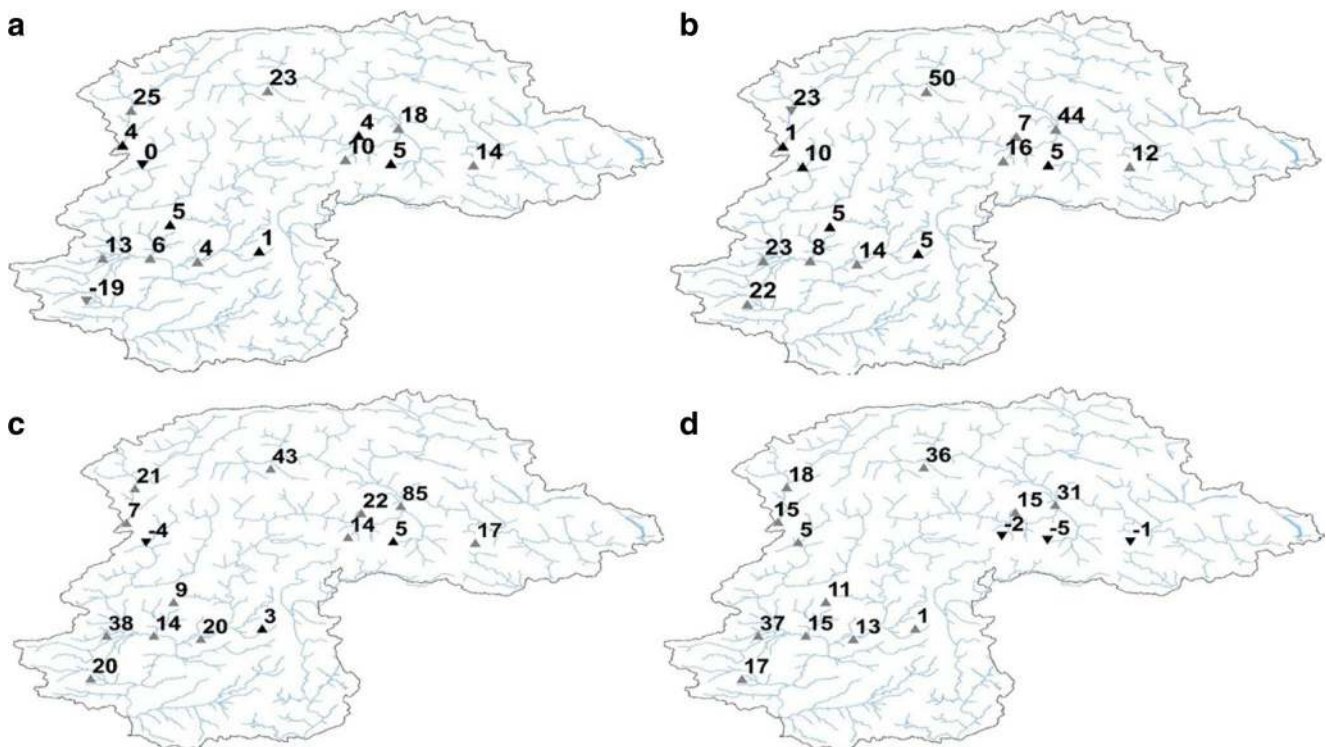
Fig. 5 Spatial distribution of trends detected by Mann–Kendall and trend values estimated by Sen's method of summer precipitation showing change in percentage per decade over time windows in: **a** 1961–2013, **b** 1971–2013, **c** 1981–2013, and **d** 1991–2013 (*upward and downward arrows* show positive and negative trends, respectively; *bold arrow* shows significant trends at $\alpha = 0.1$)

Fig. 6 Spatial distribution of trends detected by Mann–Kendall and trend values estimated by Sen's method of spring precipitation showing change in percentage per decade over time windows in: **a** 1961–2013, **b** 1971–2013, **c** 1981–2013, and **d** 1991–2013 (*upward and downward arrows* show positive and negative trends, respectively; *bold arrow* shows significant trends at $\alpha = 0.1$)

Fig. 7 Spatial distribution of trends detected by Mann–Kendall and trend values estimated by Sen's method of winter precipitation showing change in percentage per decade over time windows in: **a** 1961–2013, **b** 1971– 2013, **c** 1981–2013, and **d** 1991–2013 (*upward and downward arrows* show positive and negative trends, respectively; *bold arrow* shows significant trends at $\alpha = 0.1$)

5.3 Precipitation vs. elevation

The lower elevation regions (<1300 m) of Chilas, Risalpur, Peshawer, and Saidu Sharif showed positive trends in annual precipitation, ranging from 2 to 9 %, whereas the high mountainous regions (>1300 m) experienced decreasing precipitation. Generally, annual precipitation decreased with increasing elevation. Autumn, spring, and summer precipitation, especially during our first data series, decreased with increasing elevation. Whereas winter precipitation distinctly increased, in amounts ranging from 0 to 20 % among the various higher elevation stations. Despite a significant increase of precipitation in the second data period, the changes in the annual, spring, summer, and winter precipitations increased only slightly. Nevertheless, autumn remained dry in both the low- and high-elevation regions. Compared to the first two series, the third data series was rather unchanging in terms of annual and winter precipitations, as there were no deviations in the trend percentages. Whereas autumn and summer received less precipitation with decreasing trends, ranging from 17 to −2 % in autumn. Only spring showed a little increase in precipitation at all stations. While the fourth data series showed a somewhat reduced annual, autumn, summer, and winter trends while the spring exhibited minor rises in precipitation, as shown in Fig. 2.

6 Discussions

Temporal changes over this portion of the UIB followed a definite pattern of decreasing precipitation during our annual time series from 1961 through 2013. Furthermore, this last data series dominated in terms of significant negative trends. Innumerable studies have been conducted to explore climate changes, especially regarding temperature and precipitation. Khattak et al. (2011) studied precipitation data over the UIB from 20 stations from the 1967 to 2005 period, but they could not find a definite pattern.

Our study investigated the impact of the monsoonal system and its precipitation regime over the entire period, annually, and seasonally. Changes in the westerly disturbances and its associated precipitation regime might be expected to drive changes similar to those observed during the annual, autumn, spring, summer, and winter seasons. Particularly for the period from 1961 to 2013, the records show strong evidence of trends (increases and decreases) in the annual and seasonal precipitation time series. Annual precipitation showed a gradual decrease of significant positive trends, as the rates of increase fell from 47 to 13 % between the first and the third of the data series. These results explain the findings of Hasson et al. (2015) (regarding increased drying between 1995 and 2012) of reduced annual precipitation. The fourth data series shows a

drying pattern, by showing 37 % of the stations with significant decreases in precipitation. Similarly, spatial variations in the annual precipitation appear to demonstrate uninterrupted drying patterns at Astore, Bunji, Chilas, Drosh, Dir, and Kakul during all our data series between 1961 and 2013. Astore exhibited constant drops in annual precipitation, at rates of −3, −6, and −14 % per decade during the second, third, and fourth data series. These results do not support the idea of prevailing increases of precipitation at the Astore station discussed by Farhan et al. (2014).

In addition to a temporal seasonal discrepancy, our study observed severe drying patterns in autumn precipitation, with significant negative trends at stations, reaching 50 to 62 % negative significance levels during the first and last data series, respectively. Similar spatial deviations showed significant drying at Astore, Chitral, Dir, Drosh, Kakul, Skardu, and Saidu Sharif during the first, second, and fourth time series. In contrast, the second data period was quite wet at Bunji, Cherat, Gilgit, Gupis, Dir, Drosh, and Kakul. Furthermore, when these trends were plotted against elevation, continuous decreases in precipitation occurred during all data series. The rate of drying was greater at high-elevation areas, such as Astore, Gilgit, and Skardu. Similarly, drier times occurred in the spring season, and the number of stations with negative significance increased at rates of 5 to 44 % throughout the data period. Our study found a dominating drying pattern covering the areas of Astore, Chilas, Dir, Drosh, Kakul, and Saidu Sharif, which exhibited decreasing precipitation during the 1961–2013 period (in all data series). These results reflect similarities with those reported by Hasson et al. (2015) and numerous other authors (e.g. Cannon et al. 2015; Madhura et al. 2015; Ridley et al. 2013). These studies strongly support the finding of drier spring season in these areas. According to these authors, the westerly precipitation regime under changing climate and dry days have also increased the weakening and northward transfer of rainstorm trajectories, and these play major roles in the increase of precipitation during winter and decreased precipitation during spring (Bengtsson et al. 2006, Hasson et al. 2015).

Summer precipitation experienced a significant continuous decreasing trend, at rates from 19 to 50 % per decade during the first data series. Specifically, the last data series (1991–2013) indicates further drying, during which half of the stations showed significant negative trends. Weaker monsoonal influences at lower levels might be one reason for this dryness. Similarly, Astore, Bunji, Chilas, Chitral, Cherat, Dir, and Drosh remained dry during the summer season even though Chilas experienced a continuous decrease in summer precipitation, at rates of 8, −1, and −13 % per decade during the first three data series, respectively. Hasson et al. (2015) found similar results for summer precipitation at low-elevation stations during the 1995–2012 time frame. However, data at Chilas showed decreasing precipitation, contradictory to their reported conclusion of enhanced monsoonal influence at high-elevation stations, including at Chilas.

Our study found a consistent pattern of dryness, but it was not completely consistent with previous data series; in that our first two data series showed that during the winter, the temporal trends increased by 23 % during the second data series. Although negative significant trends increased by 4 and 19 % during the third and fourth data series, respectively. Bhutiyani et al. (2007) presented similar findings, and found mostly weak decreasing trends. Decreasing winter rainfall was only found to be statistically significant at only two stations. Again, these results are not consistent with those of Farhan et al. (2014) or Hasson et al. (2015), as they suggested an increase in winter precipitation. In terms of spatial variations, our study found that winter precipitation at the Astore station resembled their results, with increases of ~5 % per decade. Our results showed a drying pattern at Astore during the 1991–2013 period, at a rate of 8 % per decade. Comparing precipitation vs. elevation, our study found that the first two time series experienced drastic percentage increases regardless of elevation. However, the last two data series indicate much drier conditions, as shown by a decreasing precipitation trend with elevation. Precipitation trends appear highly unpredictable in space and time, as great differences appeared among the studies mentioned above and our findings. Difference in time periods and study areas may be responsible for these dissimilarities.

6.1 Conclusions

This study examined changes in precipitation, on both temporal and spatial scales, using precipitation data from 15 meteorological stations. Our purpose was to determine the influence of precipitation on other aspects of climate change. Our findings follow:

(1) Temporal trends in annual precipitation strongly decreased during the 1961–2013 period (covered by our first through fourth data sets). (2) Our study found robust indications of decreasing precipitation on the seasonal basis, along with a continuous decline in precipitation during all data series (autumn, spring, summer, and winter precipitation measurements. (3) Autumn was found to exhibit the driest weather during all data series. (4) Furthermore, precipitation changes with elevation trends showed that the more northern stations (specifically Astore, Chilas, and Bunji) experienced significant annual decreases and comparable decreases seasonally from the first to fourth data series. (5) The most southwesterly stations (Dir and Drosh) showed significant decreasing trends during all data periods, both annually and seasonally. (6) Meanwhile, stations with lower elevation catchment basins (such as Cherat, Kakul, Risalpur, and Saidu Sharif) showed significantly declining trends during the entire data period. (7) Furthermore, precipitation for these lower elevation stations did not increase with elevation; instead, the high-elevation regions experienced drying patterns during whole data period. For example, Astore,

Chilas, and Skardu remained dry during most of the annual and seasonal time periods. (8) Our data showed autumn to be the driest season during the 1961–2013 period, and that may be a factor in weaker monsoonal precipitation cycles or the temporal shifting of seasonal precipitation.

Climate changes occur most noticeably in terms of temperature and precipitation over the UIB according to various authors. Moreover, this study found the spring season to be quite dry, supporting the idea of declining precipitation (reported by numerous studies carried out earlier in this area). Downstream areas in the lower portions of the drainage basin (where most of the population depends on the agriculture) are being affected by decreasing rainfall and its impacts on crop sowing and harvesting times. There will be more stress on available water resources (which are already scarce) if precipitation does not show any significant upsurge: increased dryness could further stress agricultural production. To avoid this potentially distressing situation from getting worse, water resources management must play an important role to ensure the best utilization of available resources; e.g., flood control, building dams and reservoirs, lining of canals and water courses, and conservative surface irrigation (trickle and sprinkler irrigation). This study's extensive analysis of data from 15 climatic stations cannot support the prevailing suppositions that increasing precipitation results from green house gases and effects of the summer Asian monsoon over UIB related to global warming. Nevertheless, these early interpretations must be validated by additional and updated observations, especially with the inclusion of further hydrological modeling.

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