Conflicting signals of climatic change in the Upper Indus Basin

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Abstract
Temperature data for seven instrumental records in the Karakoram and Hindu Kush Mountains of the Upper Indus Basin (UIB) have been analysed for seasonal and annual trends over the period 1961-2000 and compared with neighbouring mountain regions and the Indian sub-continent. Strong contrasts are found between the behaviour of winter and summer temperatures and between maximum and minimum temperatures. Winter mean and maximum temperature show significant increases whilst mean and minimum summer temperatures show consistent decline. Increase in diurnal temperature range (DTR) is consistently observed in all seasons and the annual data set, a pattern shared by much of the Indian sub-continent but in direct contrast to both GCM projections and the narrowing of DTR seen worldwide. This divergence commenced around the middle of the 20th century and is thought to result from changes in large-scale circulation patterns and feedback processes associated with the Indian monsoon.

The impact of observed seasonal temperature trend on runoff is explored using derived regression relationships. Decreases of ~20 % in summer runoff in the Rivers Hunza and Shyok are estimated to have resulted from the observed 1 ^{o}C fall in mean summer temperature since 1961, with even greater reductions in spring months. The observed downward trend in summer temperature and runoff is consistent with the observed thickening and expansion of Karakoram glaciers, in contrast to widespread decay and retreat in the Eastern Himalaya. This suggests that the Western Himalaya is showing a different response to global warming than other parts of the globe.

1. Introduction

Instrumental records show a systematic increase in global mean temperature (Folland et al. 2001a, b), with global mean temperature increasing at a rate of 0.07 °C decade^{-1} over the last century (Jones and Moberg 2003). In addition, the 1990s were the warmest decade, and 1998 the warmest year since the start of the global mean temperature record in 1856 (Jones and Moberg 2003). However, the warming has not been globally uniform. High northern latitudes have been particularly affected, with reconstructions of mean surface temperature over the past two millennia suggesting that the late 20th century warmth is unprecedented (Mann and Jones 2003) and attributed to the anthropogenic forcing of climate (Thorne et al. 2003).

In most parts of the world there have been differential changes in daily maximum and minimum temperatures, resulting in both a narrowing of diurnal temperature range (DTR) and an increase in mean temperature (Karl et al. 1993; Easterling et al. 1997; Jones et al. 1999). However, there are suggestions that the western Himalayan region is showing a different response to global warming (Kumar et al. 1994; Yadav et al. 2004), with an increase in DTR and a cooling of mean temperature in some seasons, possibly as a result of local forcing factors.

Meteorological data from the western Himalaya and trans-Himalayan mountains have been little studied. However, here we report on instrumental records of temperature data obtained for the Upper Indus Basin for 1900-2000. The Upper Indus Basin (UIB) (Fig. 1) is a key area for climatological studies as it is located at the boundary between tropical and continental climatic influences and has quite different climatic controls from the eastern or Greater Himalaya. Indian monsoon air masses, which bring significant rainfall on the southern margin, penetrate infrequently across the Himalayan mountain divide to the trans-Himalayan regions of the Karakoram and Hindu Kush Mountains. Precipitation in these northern ranges is concentrated in winter and spring months (Archer and Fowler 2004) and carried on westerly disturbances originating predominantly from Mediterranean and Caspian Sea regions. Winter precipitation provides the principal source for accumulation on UIB glaciers in the greatest area of perennial ice outside the Polar Regions (22 000 km^2) and upon melting, for runoff in the River Indus.

In the UIB, summer runoff (July to September) is highly correlated with winter precipitation in middle-elevation basins dependent on an ephemeral snowpack (Fowler and Archer 2005). However, summer temperature and runoff are negatively correlated on these snow-fed catchments where increased temperature results in increased evaporative loss and, since snow cover volume is limiting, in reduced runoff (Singh and Bengtsson 2005).

In contrast, summer runoff volume in major high-elevation Indus tributaries such as the Rivers Hunza and Shyok (Fig. 1), fed by glaciers and permanent snowpack, is uncorrelated with winter precipitation but highly correlated with summer mean temperature (Archer 2003). Linear regression analysis by Archer (2003) suggests that a 1 °C rise in mean summer temperature arising from climate change would result in an increase of 17 % in summer runoff for the River Shyok (basin area 65 025 km^2) and a 16

% increase for the River Hunza (basin area 13 925 km^2) respectively. Together these two catchments contribute more than 25 % of the inflow to Tarbela Dam which is the main controlling structure for the Indus Basin Irrigation System, one of the world's largest integrated irrigation networks.

These significant, but opposing, responses to variations in summer temperatures indicate potentially important practical impacts of temperature trend on river runoff and consequently upon local rural livelihoods in the mountains and the productivity of agriculture on the Indus plains. In addition, temperature trend (along with precipitation) will influence glacier mass balance. Whilst most of the world's mountain glaciers have been shrinking for at least the last thirty years (WGMS 2000), including the neighbouring Greater Himalaya (Hasnain 1999; Mastny 2000; Shrestha and Shrestha 2004), Hewitt (1998) reports the widespread expansion of larger glaciers in the Central Karakoram, accompanied by an exceptional number of glacier surges. This contrast in the behaviour of glaciers suggests a pattern of climatic change in Karakoram and Hindu Kush different from that in the Greater Himalaya.

The main focus of this study is the investigation of trend in seasonal temperature series over the period from 1900-2000 for seven instrumental records in the UIB. However, before this is addressed, the extent to which individual temperature records or groups of records are representative of the region in which they are located has been investigated.

2. Data

The analysis investigates temperature data for six trans-Himalayan stations. Srinagar, in the neighbouring Jhelum Basin south of the Himalayan divide, has also been included for comparison. The stations cover a distance east to west of about 300 km and from north to south of over 200 km. Several data sources have been used, with measurement locations shown in Figure 1 and listed in Table 1.

INSERT FIGURE 1 AND TABLE 1

1. For the pre-partition period up to 1947, climate statistics for Gilgit and Skardu were derived from daily maximum and minimum temperature records in six-monthly daily weather reports published by the India Meteorological Department (IMD). Records commence in 1900 and 1903 for Skardu for Gilgit respectively and are essentially complete from 1905-35 and thereafter intermittent until 1947.

2. For the post-partition period, the Pakistan Meteorological Department (PMD) provided daily records of maximum and minimum temperature for Gilgit and Skardu to 1999, and monthly records for Astore, Dir, Drosh and Bunji to 1997. With the exception of Dir (1967) these records commence in the 1950s.

3. Monthly mean temperature records for Srinagar (1894-2000) were obtained from the Climate Research Unit at the University of East Anglia, UK.

Validation was based primarily on the inspection of records as station history files were not available. Interviews with PMD staff indicated that standard meteorological practice, established by IMD in 1891, had been adopted by PMD. However site changes had occurred during partition at Gilgit and Skardu. The principal validation check of records was through double mass curve analysis which was carried out by season and separately for maximum and minimum temperatures. No evidence of heterogeneity was found in any of the maximum temperature series. However, two years of inconsistent winter minima were identified at Drosh and these were substituted by regression with other stations. In addition, evidence of heterogeneity was identified in the winter minimum record at Bunji with a break of slope in the double mass curve in 1977; the pre-1977 record was adjusted to be consistent with the later record. Whilst Bunji has been included in time series analysis, results must be considered of lower reliability than other stations.

The network has recently been supplemented by more than 20 automatic observations stations at elevations up to 4700 m (Hewitt and Young 1993) and although record lengths are still insufficient for time series analysis, they provide a basis for assessing altitudinal variations in temperature (Archer 2004).

3. Spatial correlation of temperature

Mountainous regions are said to require a much greater density of climate stations than neighbouring flat lands to achieve the same reliability of areal estimates. In this most rugged of world landscapes, the existing network of weather stations falls short by more than an order of magnitude of the minimum requirement recommended by WMO (1970). The climatological network in the UIB is also biased by the predominant location of stations on valley floors. In order to assess regionally representative trends, it is necessary to establish the extent to which seasonal variations are correlated across the region.

Archer (2004) investigated such spatial relationships by drawing up correlation matrices of annual, seasonal and monthly temperature between weather stations in the region. A summary of the results for the key period of ablation, divided between the 3-month periods April-June and July-September is shown in Table 2.

INSERT TABLE 2

Conclusions drawn from this table and from the related analysis are:

1. In spite of the separation of stations in distance and by major topographic barriers, seasonal temperatures and their anomalies are remarkably conservative. Positive correlation coefficients occur between all stations, with better correlation over shorter distances. The high correlation between temperature at Srinagar, south of the main Himalaya, and the northern valley stations of Gilgit and Skardu is particularly notable.

2. Correlation of mean temperature is much higher during spring and summer than during winter months. It is suspected that in winter, under the prevailing influence of the Tibetan anticyclone, more local conditions such as valley temperature inversions prevail.

3. On average, the station with the best seasonal correlation with other stations is Astore. For the spring and summer season the correlation coefficient for mean temperature is greater than 0.75 with all stations except Drosh (July-September).

4. Correlation of monthly mean temperature at high altitude stations with short records at Kunjerab (4733 m) and Shandur (3719 m) with Gilgit and Skardu yielded on average eight months with correlation coefficients greater than 0.8.

The high correlation in mean temperature between valley stations separated by considerable horizontal distances and intervening mountain barriers strongly implies that these correlations will be reflected over much shorter vertical distances. This view is supported by the high correlation coefficients demonstrated between valley stations and short records from high altitude stations (Archer 2004) and also by the strong observed correlation between valley temperature stations and catchment runoff which is generated at higher elevations (Archer 2003). It is concluded that the temperature time series at valley stations can be used to provide a regional picture of seasonal and year-to-year variation in temperature.

4. Analysis of temporal temperature change

a. Recent trends in the UIB, 1961-2000

Trends in annual, seasonal and monthly mean, maximum and minimum temperature were investigated for stations at Dir, Drosh, Bunji, Gilgit, Astore, and Skardu for the period 1961-1999. For Srinagar, only mean temperature for the period 1961-2000 was available for investigation. The records were analysed by fitting a linear least squares trend line to the annual deviation from the mean and assessing the significance of trend using Student's t-test.

A summary of the results of the analysis of annual and seasonal trend in mean, maximum and minimum temperature is provided in Table 3 as a basis for illustrating common patterns and regional differences. Increases (+) and decreases (–) are shown as change in ^{o}C decade^{-1}.

INSERT TABLE 3

1) MEAN TEMPERATURE

In contrast to many studies indicating unprecedented warming in the 20th Century (e.g. Jones and Moberg 2003), about half the records in the UIB show a reduction in mean annual temperature since the 1960s and more strongly since the mid-1970s. The trend is statistically significant ($p<0.05$) at Bunji and Dir, with cooling rates of –0.52 and –0.26 ^{o}C decade^{-1} respectively since 1961 (Table 3a) and is related to significant cooling of mean temperature during spring (March to May and summer (June to August).

All records show summer cooling. This is large and statistically significant at some stations, with rates of –0.30, –0.99, –0.35 and –0.38 ^{o}C decade^{-1} at Astore, Bunji, Dir and Gilgit respectively (Table 3a). Similarly, a cooling of spring mean temperature is seen at four out of seven stations, although this is only significant at two stations, and the Skardu record exceptionally shows significant warming (Table 3a). Change in fall (September to November) mean temperature over the same period is comparable to that of spring.

In contrast, in winter (December to February) a warming of mean temperature is seen in most records from 1961-2000. There is evidence of significant winter warming trend at

Skardu and Srinagar ($p<0.05$), Gilgit ($p<0.10$) and Drosh, with insignificant trend seen at other stations.

2) MAXIMUM AND MINIMUM TEMPERATURE

Since 1961 there has been an increase in annual maximum temperature at all stations except Bunji and Astore (Table 3b). Significant increases in maxima occur mainly in the months from November to April. However, there has also been a contrasting and large region-wide reduction in annual minimum temperature (Table 3c). This has resulted in the most striking change in the UIB since 1961; the large increase in diurnal temperature range (DTR) at all stations. Figure 2a shows the large increase in annual DTR at Skardu, Gilgit and Dir during 1961-2000. This increase is strongest at the northern stations, with Skardu, Gilgit, Bunji and Dir showing positive trends in DTR in all months as well as all seasons. The largest change in DTR is found during the period from June to November, summer and fall seasons, when all stations in the UIB show a positive trend (see Figure 2b).

INSERT FIGURES 3 AND 4

Figure 3 clearly shows the region-wide increase in winter maximum temperature over the period 1961-1999 at stations in the UIB. There have been smaller changes in winter minimum temperature, with small increases in minima at some sites and decreases in minima at others (Figure 4). Since temperature maxima have been increasing at a greater rate than temperature minima, even where temperature minima are also increasing, there has been a resulting increase in winter DTR at all stations, and an overall winter warming.

In contrast, in summer months there has been a region-wide reduction in temperature maxima (Figure 5; Table 3c). Perhaps more surprising however, is the concomitant reduction in summer temperature minima (Figure 6). This is statistically significant ($p<0.05$) at all stations except Drosh and there are high rates of cooling, with reductions of –1.11 ^{o}C decade^{-1} at Bunji (Table 3c) and over –0.4 ^{o}C decade^{-1} at all stations. These reductions are at their highest in June. Summer cooling of mean temperature therefore arises as a result of reduction in both minimum and maximum temperature, although the former has by far the larger effect. Since summer minimum temperatures show greater rates of cooling than maximum temperatures, there has also been a significant increase in summer DTR.

INSERT FIGURES 5 AND 6

In spring and fall, a similar cooling trend is also observed in temperature minima (Table 3c). Combination with a significant upward trend in temperature maxima at many stations in fall (Table 3b) results in an increasing DTR in spring and fall.

b. Long term change at Gilgit, Skardu and Srinagar

The duration of temperature records at Gilgit and Skardu is unique for currently operating trans-Himalayan stations. Despite discontinuity in the temperature record arising from site changes, comparison has been made of trend in maximum, minimum and mean monthly temperatures and diurnal temperature range for the periods 1905-35 and 1958-2000 and mean temperature only for Srinagar.

There are significant features in common between the UIB stations and Srinagar during the later period. All records show a warming trend in winter months related to increases in maximum temperature and cooling in summer months related to decreases in minimum temperature.

Both the significant correlation in seasonal temperature between Srinagar and UIB stations (Table 2) and the similarity of the seasonal pattern of change with Gilgit and Skardu suggest that the homogeneous record at Srinagar may provide an analogue of long-term climatic change for the UIB.

INSERT FIGURE 7

Over the longer period from 1894-2000, there has been a small but significant increase in annual mean temperature at Srinagar ($p<0.01$) of 0.07 ^{o}C decade^{-1}. During the winter season (DJF), prior to 1960 the temperature increase was significant ($p<0.05$) at 0.10 ^{o}C decade^{-1} (Figure 7a). However, since 1960 the rate of increase has accelerated to 0.51 ^{o}C decade^{-1}. If individual months are further examined then large and statistically significant increases in mean temperature are seen in the months from December to April ($p<0.05$) (Figure 7b-f), with smaller increases during other months. This suggests that the rate of increase of winter temperatures in the UIB region may also have accelerated in the period since 1960, a possible signal of global warming.

More remarkable however, is the sudden divergence in temperature trend between the high latitudes and the western Himalayan region in the latter half of the 20th century. Climate proxies and instrumental records in the northern high latitudes indicate unprecedented warming (Jones and Moberg 2003) but instrumental records from some UIB stations suggest a cooling of mean temperature, annually and in all seasons except winter. Figure 8 shows the striking increase in DTR at Skardu and Gilgit in all seasons since the middle of the 20th century. This has been caused by a pronounced cooling of minimum temperatures since 1960, particularly in summer and autumn (Figure 9), coupled to a slight increase in maximum temperatures in all seasons since 1960 (Figure 10). This is in contrast to the narrowing of DTR seen in the rest of the globe since 1950 (Karl et al. 1993; Easterling et al. 1997). However, during 1900-1935 DTR in the Western Himalaya remains constant or shows a decreasing trend. This is akin to trends seen in the remainder of the northern high latitudes since 1950 (e.g. Karl et al. 1993).

INSERT FIGURES 8, 9 AND 10

5. Discussion

Whilst there are clearly differences between stations in the amount and even the direction of change, several common patterns emerge, especially when the series are decomposed into seasonal groupings or minimum and maximum series. Trends are less evident in annual and mean temperature series because of differences in behaviour between seasons or between maxima and minima. The observed trends in the UIB may be compared with those in neighbouring parts of the Himalaya, Central Asia and the Indian sub-continent. The most significant common patterns in the UIB with respect to the 1961 to 2000 period are:

1. Decrease in summer mean temperature (7 of 7 sites, 4 significant)
2. Decrease in summer minimum temperature (6 of 6 sites, 5 significant)

3. Decrease in fall minimum temperature (6 of 6 sites, 4 significant)
4. Increase in fall maximum temperature (4 of 6 sites, 3 significant)

5. Increase in winter mean temperature (5 of 7 sites, 3 significant)
6. Increase in winter maximum temperature (5 of 6 sites, 3 significant)

7. Decrease in annual minimum temperature (6 of 6 sites, 4 significant)
8. Increase in annual maximum temperature (4 of 6 sites, 2 significant)

a. Summer temperatures

The consistent decline in summer mean and minimum temperatures is perhaps surprising in view of the frequently observed and reported retreat of neighbouring Himalayan glaciers which has generally been ascribed to an increase in summer temperatures (Yamada et al. 1992: Hasnain 2002). Thus, Shrestha et al. (1999) derived a consistent upward trend in summer temperatures (June to September) for the period 1977-94, ranging from 0.14 ^{o}C for low elevations stations to significant increases of 0.62 ^{o}C for the Nepal Himalaya and 1.09 ^{o}C for trans-Himalayan stations. However their analysis was based on maximum temperatures, whilst minimum temperatures, graphed but not analysed, show steady or falling trends. The decline in minimum temperatures was confirmed by Sharma et al. (2000) for several stations in Nepal. Only two of six stations in the UIB show an upward trend in maximum summer temperatures, but the rates of decrease in the remaining four stations is much less than for minimum temperatures.

Cook et al. (2003) re-examined a longer Kathmandu mean temperature record and compared it with a 0.5^{o} latitude-longitude gridded data set based on records from neighbouring northern India; both showed a cooling trend in the monsoon (June to September) as well as in pre-monsoon (February to May) for the period 1901-95. In addition, Hingane et al. (1985) found that northwest India shows a significant decreasing trend in mean temperature in summer (–0.05 ^{o}C decade^{-1}), confirmed by Kumar et al. (1994) for minimum temperature at the same decreasing rate whilst maximum temperatures have been steady. Summer cooling trends during the last few decades of the 20th century have also been documented for the Tibetan Plateau (Liu and Chen 2000), central China (Hu et al. 2003) and central Asia (Briffa et al. 2001).

Cook et al. (2003) attributed this summer cooling to a coincident increase in monsoonal rainfall in Nepal after 1950, which would tend to suppress temperatures through increased daytime cloud cover. In the UIB, increasing summer rainfall since 1961 (Archer and Fowler 2004) and a coincident increase in daytime cloud cover may also provide an explanation for the summer cooling of maximum temperature. It does not, however, provide an explanation for the large decreases in minimum temperature seen in the summer. The decrease in summer temperature minima could be explained by a coincident decrease in cloud cover during the night, although currently no data are available to assess this.

Other research has argued for an aerosol induced cooling over much of South Asia and the Indian sub-continent. Ramanathan et al. (2005) suggest that during the "Indian dry season", from October to May, the haze from the Asian Brown Cloud reduces the surface receipt of solar radiation and precipitation efficiency, causing the observed cooling of sea surface temperatures over the North Indian Ocean. This, in turn, has reduced monsoonal summer precipitation by ~5 % in India. However, it is unlikely that aerosol induced cooling offers an explanation for the observed summer cooling in the UIB as during the monsoon the effect of the Asian Brown Cloud is diminished in India. Additionally, the UIB region has seen a recent increase in summer precipitation (Archer and Fowler 2004).

Finally, there is the possibility that upward trends in summer temperature are more marked at higher elevations as shown by the analysis of Shrestha et al. (1999) for Nepal; this would imply an increased lapse rate. A tendency for warming trends to increase with elevation has been found in both the Tien Shan of central Asia (Aizen et al. 1997) and the Tibetan Plateau (Liu and Chen 2000). There is no evidence of elevation dependency of warming (cooling) in our dataset although this would not necessarily be expected as the range in station elevation is only 1000 m.

b. Winter temperatures

The UIB shows a consistent increase in winter mean temperatures, with increasing trends since 1961of 0.17 and 0.38 ^{o}C decade^{-1} at Gilgit and Skardu respectively. The increase in winter maximum temperature is large and statistically significant ($p<0.05$) at Gilgit, Skardu and Dir, with increases of 0.27, 0.55 and 0.51 ^{o}C decade^{-1} respectively. However, winter minima only show positive trend at three of six stations and none are significant. The greatest warming in neighbouring regions has also been in winter temperature. For India, Kumar et al. (1994) derived winter warming trends of 0.09 ^{o}C decade^{-1} for maximum temperature in an all-India temperature series and 0.08 ^{o}C decade^{-1} for the Northwest region for 1901-1987. Shrestha et al. (1999) found increases in winter maxima of 0.61, 0.90 and 1.24 ^{o}C decade^{-1} for all-Nepal, Himalaya and trans-Himalayan stations respectively. Cook et al. (2003) found indications of warming since 1960 at Kathmandu only in the period from October to January. Additionally, decadal winter increases of 0.32 ^{o}C in mean temperature have been suggested for the Tibetan Plateau by Liu and Chen (2000).

Change in winter temperature is less hydrologically significant than change during the summer as the most hydrologically active areas above an elevation of 4000 m remain above the freezing level even for maximum temperatures from November to March (Archer 2004). However, average winter slope lapse rates of 0.70 ^{o}C (100 m)$^{-1}$ (Archer 2004) suggest a shift in the freezing level of up to 300 metres and a decrease in the ratio of solid to total precipitation at lower elevations. The perception of decreased winter precipitation and increased duration of opening of passes in the Karakoram and Hindu

Kush through the winter may therefore be explained by an increase in winter temperature rather than by the decrease in winter precipitation suggested by Miehe (1992).

c. Asymmetry of temperature trends

The most striking temperature change in the UIB is the recent large increase in diurnal temperature range (DTR) observed in all seasons and in the annual data set. Contrasting changes in maximum and minimum temperature for Gilgit and Skardu shown in Figures 9 and 10 suggest that this divergence commenced around the middle of the 20th century. This change is not unique to the high elevation UIB but has also been observed over much of the Indian sub-continent (Kumar et al. 1994; Yadav et al. 2004). In contrast, in many other parts of the world and especially at higher latitudes, minimum temperatures have increased at a higher rate than maximum temperatures resulting in a narrowing of DTR since 1950 (Karl et al. 1993; Easterling et al. 1997; Jones et al. 1999). Several explanations have been put forward for the increase in DTR on the Indian sub-continent.

Firstly, it has been proposed by Cook et al. (2003) that change in DTR may result from change in precipitation regime. Cloud cover associated with precipitation reduces incoming solar radiation and traps more outgoing long wave radiation, hence altering surface temperature, and up to 80 % of DTR variance can be explained by cloud and precipitation records (Dai et al. 1999). Given the observed increase in DTR, one would expect a decrease in precipitation, with warmer days and cooler nights. However, Archer and Fowler (2004) found an increase in annual precipitation in the UIB from 1961-2000 with a statistically significant ($p<0.05$) upward trend in precipitation during both winter and summer seasons.

Simple monthly correlation of various indices of temperature with precipitation is shown in Table 4. There is a significant ($p<0.05$) negative relationship between maximum temperature and precipitation in summer and autumn months, suggesting that increased precipitation may induce a cooling effect on maximum temperature. A negative relationship is found between DTR and precipitation in all months, suggesting that an increase in precipitation should cause a decrease in DTR. Therefore, although change in summer precipitation regime provides a possible explanation for reduction in summer maximum temperature in the UIB, it does not explain the increase in DTR.

INSERT TABLE 4

Secondly, change in snow covered area and thus regional albedo may cause change in DTR. Snow covered area is mainly dependent on winter precipitation. Low winter precipitation can therefore lower the regional albedo and cause more solar radiation to be absorbed at the surface, causing increased surface melting. This positive feedback may result in increased daytime maxima (Giorgi et al. 1997). Archer (2001) illustrated this feedback for the UIB by examining correlation between winter half-year precipitation and spring temperature. Significant negative correlation was found only for the months of March and April and did not persist into the summer. Additionally, as noted above, winter precipitation has shown an increasing trend in the UIB, significant ($p<0.05$) at three stations (Archer and Fowler 2004). Snow albedo feedback is therefore unlikely to have caused the observed change in DTR. Besides, Kumar et al. (1994) report that the temperature asymmetry is shared with northwest India where there is no question of snow albedo effects.

Thirdly, in the lowlands of the Indian subcontinent, land use changes including urbanisation (Kumar and Hingane 1988), deforestation (Kothyari and Singh 1996; Yadav et al. 2004), intensification of agriculture (Singh and Sontakke 2002) and widespread irrigation (Adel 2002) have been associated with change in temperature. Urbanised areas often show a narrower DTR than nearby rural areas (Gallo et al. 1996) as urbanisation increases night time temperature minima through the urban heat island effect. Land use change such as deforestation may reduce temperature minima. Yadav et al. (2004) suggest that land use change, particularly deforestation, has contributed to a widening of the DTR in northern India. The removal of tree canopies causes direct heating of the soil during the daytime to increase. This is then followed by the intense cooling of these soils by terrestrial radiation loss at night, enhanced by the fact that they are moisture deficient due to surface runoff increases associated with deforestation.

Development activities leading to widespread deforestation do appear to offer a plausible explanation for the increase in DTR in northern India with, for example, 50 % of the areal expansion of agricultural land in the Garhwal region during 1963-93 coming from previously forested area (Sen et al. 2002). However, the same trend of widening DTR is also seen in the neighbouring region of the UIB where none of the same developmental pressures are relevant. Here, natural vegetation is sparse and limited either by low rainfall at lower elevations or by low temperature at higher elevations. The UIB is also sparsely populated and neither the limited urbanisation nor land use change on valley floors is considered sufficient to have initiated significant changes in thermal regime. The upward trend in DTR in the UIB is thus difficult to explain but the region-wide increase in DTR in the western Himalaya suggests that changes in atmospheric circulation patterns are likely to be responsible.

GCM future climate simulations for the region from the IPCC suggest more pronounced increases in minimum rather than maximum temperatures over Asia related to increased monsoon cloudiness; hence an increase in mean temperature annually and during all seasons and a decrease in annual and winter DTR (Lal et al. 2001; IPCC 2001). During the summer an increase in DTR is simulated, but this results from a more pronounced increase in maximum temperature relative to the minimum temperature (Lal and Harasawa 2001) rather than the reduction in both summer maximum and minimum temperatures seen since 1961 in the UIB. GCM integrations also suggest that variations in DTR are much more sensitive to changes in feedbacks than in direct forcings, with variations arising mostly from changes in clouds and in soil moisture (Stone and Weaver 2003). It is possible that global warming has induced process coupling in this region that occurs on scales too small for current GCMs to resolve well. Using a regional climate model, Pan et al. (2004) were able to simulate a local minimum of warming in the central U.S. (a "warming hole") associated with changes in low-level circulations caused by regional-scale feedback processes under global warming. This was absent from GCM simulations of the same region, but this region is known to have experienced a cooling trend in the last 25 years of the 20th century. A relationship between circulation changes during the northern Hemisphere winter and winter DTR has also been examined by Easterling et al. (1997) for the European region. They found a bipolar pattern resulting from stronger westerly flows, with Iberia experiencing increased DTR and decreased DTR over northern Europe.

Whilst explanation remains problematic; the shared response with the Indian sub-continent suggests a regional or latitudinal effect associated with large scale climatic processes and their feedbacks, and related to changes in the monsoon circulation. Recent modelling work by Ramanathan et al. (2005) suggests that aerosol cooling brought about by the Asian Brown Cloud in the non-monsoonal months from October to May has caused the observed cooling of sea surface temperatures over the North Indian Ocean. This, in turn, weakens the monsoon circulation and causes a southward shift, reducing rainfall in India by ~5%. In the UIB, summer precipitation has increased but this may be due to the incursion of more westerly weather rather than monsoonal incursions. The relationship with circulation patterns will be investigated in further work.

e. Impacts on hydrological regime and glaciological response

The overall impact of climate change on glacier mass balance and hydrological regime require the joint consideration of change in both precipitation and the energy budget. The emphasis here is on the energy budget as represented by temperature. Trend in temperature will have both short and long term effects on glaciers and runoff.

Short term effects of seasonal temperature are revealed by the analysis of current variability. A seasonal mean temperature range for both spring (April to June) and summer (July to September) of 4 to 5 °C since 1961 provides a basis for the assessment of impact on runoff. Observed summer runoff on high altitude glacier-fed catchments shows significant positive correlation with summer temperatures. For the River Hunza a 1 °C fall below the current mean temperature – the approximate fall in summer mean temperature since 1961 – gives a predicted 16 % decrease in summer runoff ($r = 0.69$); the equivalent decrease for the River Shyok is 17 % ($r = 0.73$) using derived regression equations (Figure 11). An equivalent figure of 16 to 18 % increase for a 1 °C temperature rise was identified from modelling studies on Western Himalayan catchments by Singh and Kumar (1997). Predicted runoff decreases for a 1 °C fall in the spring months are more striking, 48 % and 62 % for the Rivers Hunza ($r = 0.73$) and Shyok ($r = 0.53$) respectively, albeit with much smaller runoff totals.

INSERT FIGURE 11

On average, climate stations in the UIB show a reduction in mean summer temperature of 1.2 °C over the period 1961 to 2000 (Figure 11a); representing a fall in average summer runoff of ~20 % on the Hunza and Shyok (Figure 11b). Similarly, during the more sensitive spring months a 40 year mean temperature reduction of 0.3 °C yields a fall in spring runoff of 15 to 20 %. These results are confirmed for the River Hunza at Dainyor where there is a significant downward trend in observed runoff for spring and summer of 46 % and 35 % respectively over the record period (Figure 11c). The absence of significant trend for the River Shyok (Figure 11c) may be explained by regional variation in climatic trends. The nearest climate station to the Shyok catchment at Skardu has an upward trend in spring and a less steep downward trend in summer than other UIB stations (Table 3a).

In contrast, temperature and runoff are negatively correlated on middle altitude (snow-fed) catchments although the regression relationships are non-significant and confounded by the much stronger influence of winter precipitation (Archer 2003). Singh and Bengtsson (2005) suggest through modelling studies that increased temperature results in increased evaporative loss and – since snow cover volume is limiting – in reduced runoff, with an estimated decrease of 9 % for a 1 ^{o}C rise in temperature for the Satluj basin in the Western Himalaya.

These varied responses to changes in the driving variables mean that caution must be exercised in the prediction of runoff response to climate change, especially in complex catchments such as the main stem of the River Indus where a mix of climatic controls impact upon hydrological regimes.

In the longer term, a reduction in summer temperature implies reduced ablation and a positive mass balance of glaciers, given constant precipitation. In addition, climate stations in the Karakoram show consistent positive trends in winter precipitation; averaging a 7 % increase per decade for the period since 1961 (Archer and Fowler 2004). In contrast, the observed higher temperatures in winter and associated decrease in the ratio of solid to total precipitation is expected to have an effect only on the lower reaches of glaciers. Whilst the lag time of large Karakoram glaciers may be many decades, the reported expansion and thickening of many glaciers (Hewitt 1998 and *pers. comm.)* is consistent with the observed climatic changes. Karakoram glaciers offer a significant contrast with those in the Eastern Himalaya (Hasnain 1999).

6. Conclusions

Although there are variations between the stations analysed, in terms of the amount and direction of change in temperature, common patterns emerge which can be compared with neighbouring mountain regions and the Indian sub-continent.

1. Mean and minimum summer temperatures show a consistent cooling in the period since 1961, a characteristic shared with northwest India and, at least, for lower level stations in Nepal. Summer maxima show little clear trend.
2. Winter mean and maximum temperatures show statistically significant increases. Although winter minima show no significant trend, the large positive changes in the maxima are sufficient to tip the balance of change in the mean annual temperature to significantly positive at the most easterly stations of Skardu and Srinagar.
3. The most striking recent change is the large increase in diurnal temperature range, observed at all seasons and in the annual data set. Long records at Gilgit and Skardu suggest that the trend of divergence commenced in the middle of the 20th century. This change is in common with much of the Indian subcontinent but in direct contrast to most other parts of the globe where a narrowing of DTR has been observed. Explanations are problematic but appear to be associated with the

coupling of regional feedbacks such as aerosol cooling to large scale climatic processes.

4. Using linear relationships between spring and summer temperatures and runoff in the Rivers Hunza and Shyok and the cooling regional temperature trend, produced a predicted reduction of 20% in runoff over the record period. This predicted fall was exceeded by actual runoff decreases on the River Hunza. The absence of equivalent decline in the River Shyok can be explained by regional variations in temperature trend, with easterly stations in the UIB showing more positive temperature trend.
5. Summer temperature reductions and positive trend in winter precipitation imply reduced ablation and increased accumulation of Karakoram glaciers. These climatic changes are consistent with the observed thickening and expansion of glaciers in the UIB region, in contrast to widespread retreat and decay in the Eastern Himalaya.
6. The hydrological and glaciological impact of change in temperature is highly dependent upon the season in which the change occurs. Summer cooling has a widespread impact at all elevations whereas the effects of winter warming are restricted to lower elevations only.
7. The viability of temperature measurements at valley stations as indices of change at higher elevations is confirmed both by the strong links between temperature and runoff and by the strong spatial correlation in seasonal temperature across the region.

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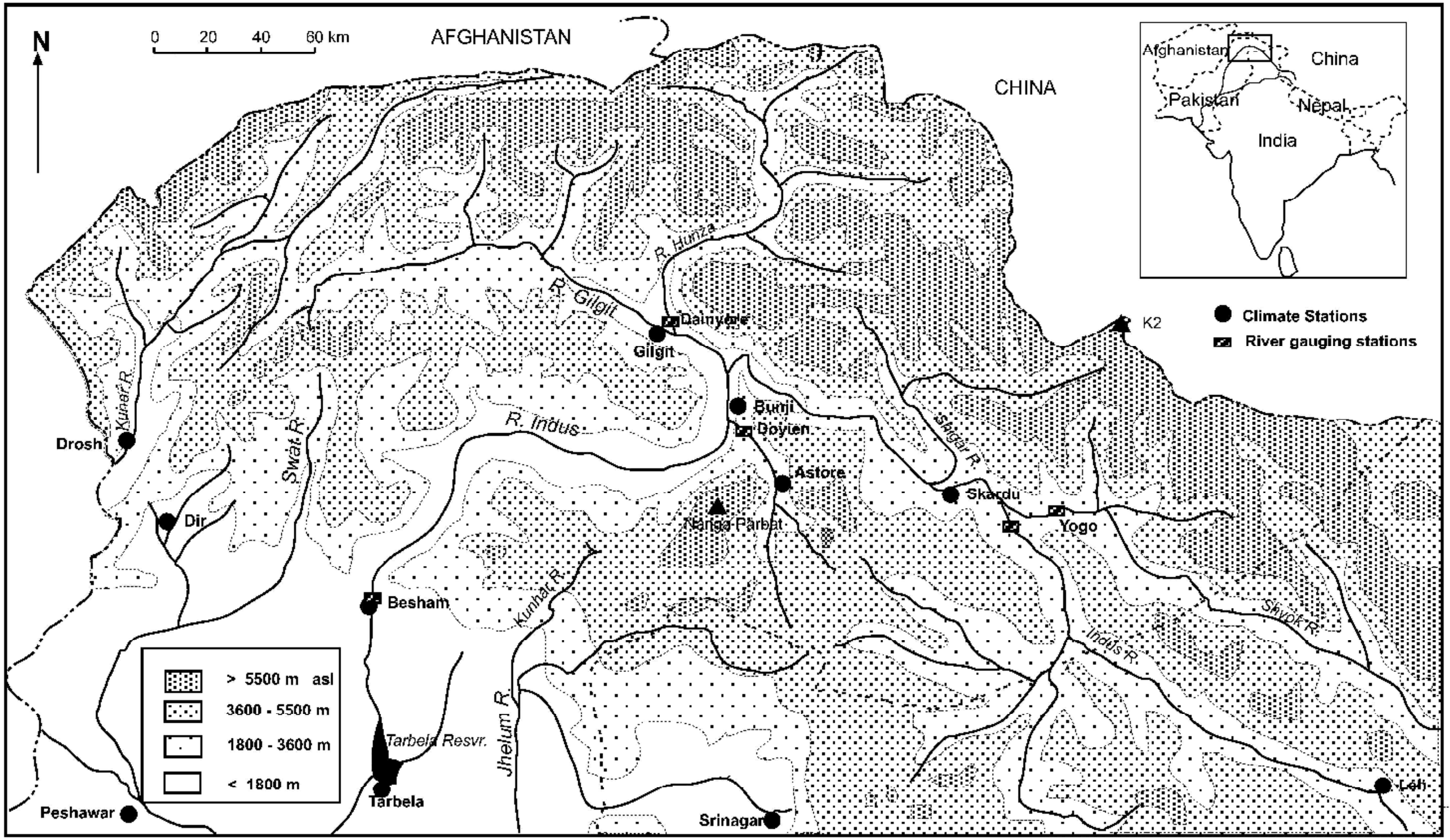

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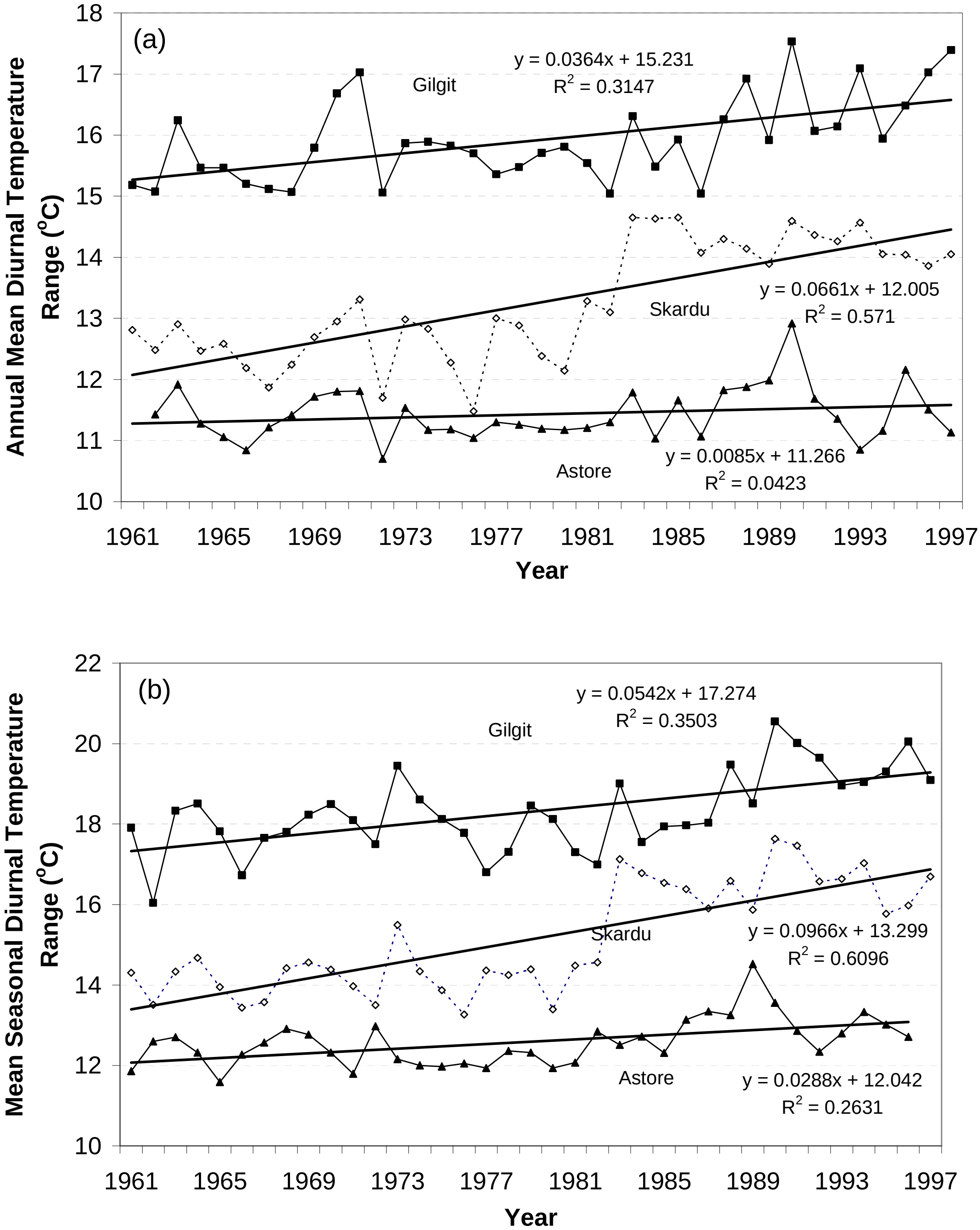

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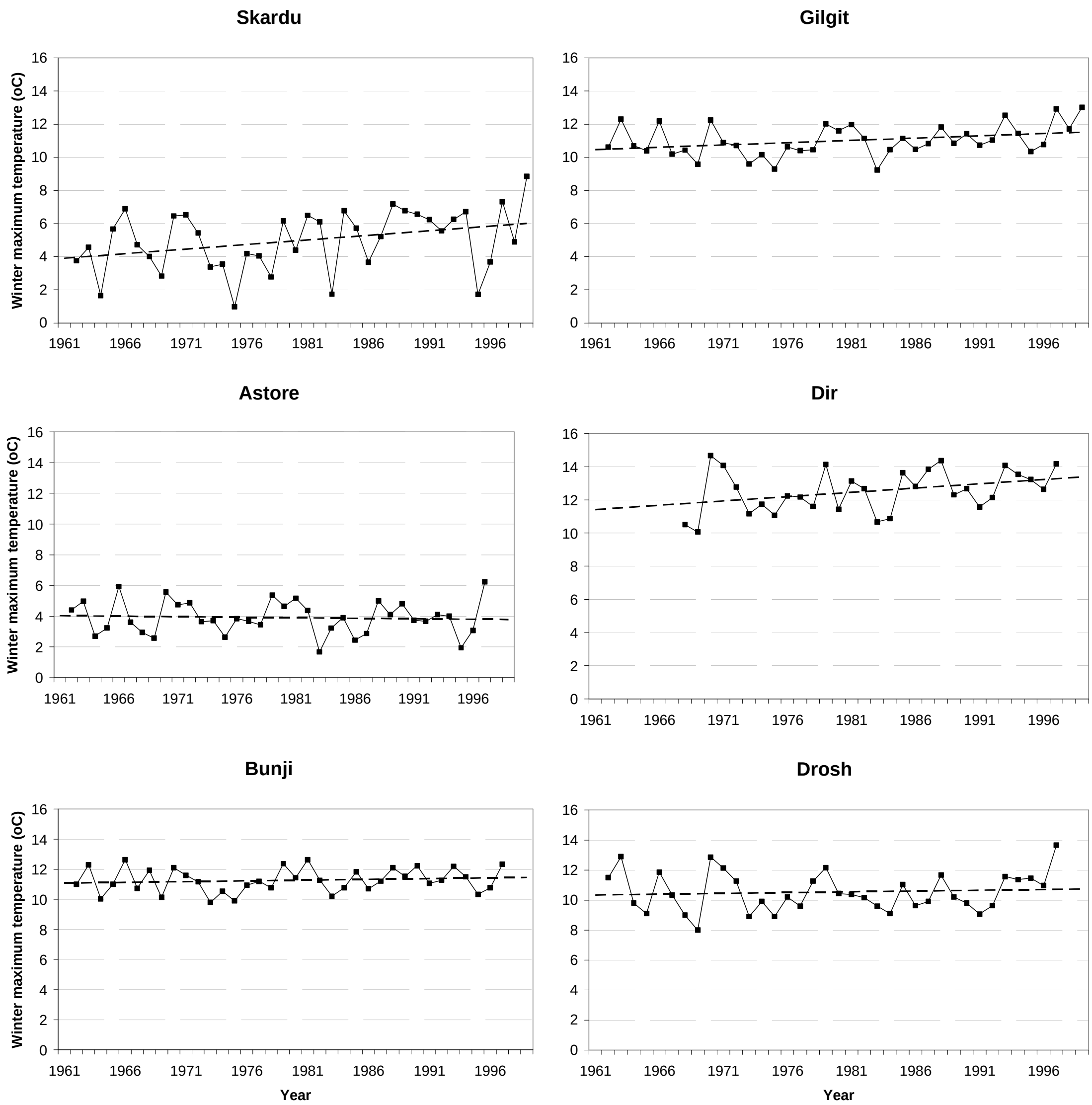

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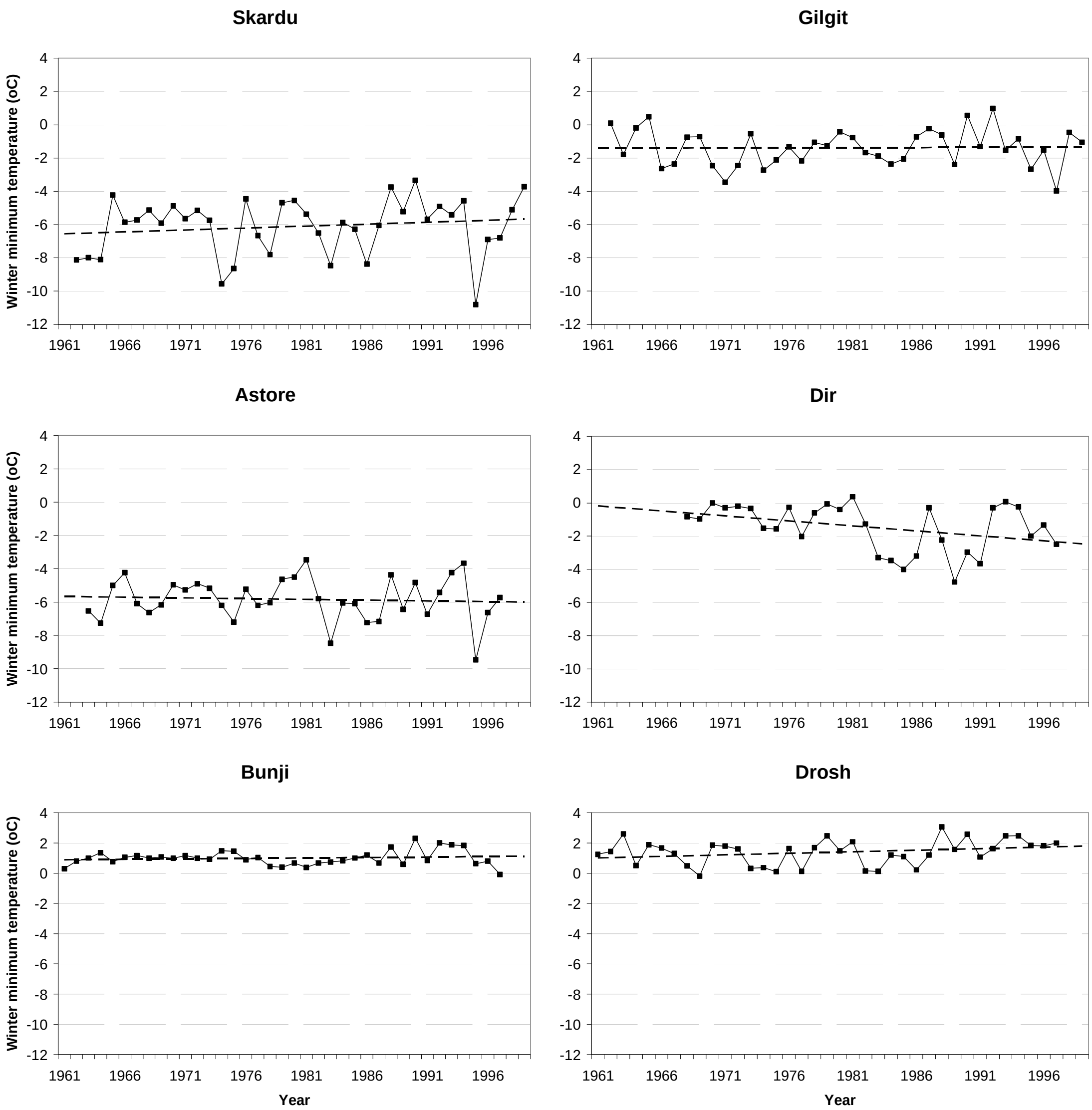

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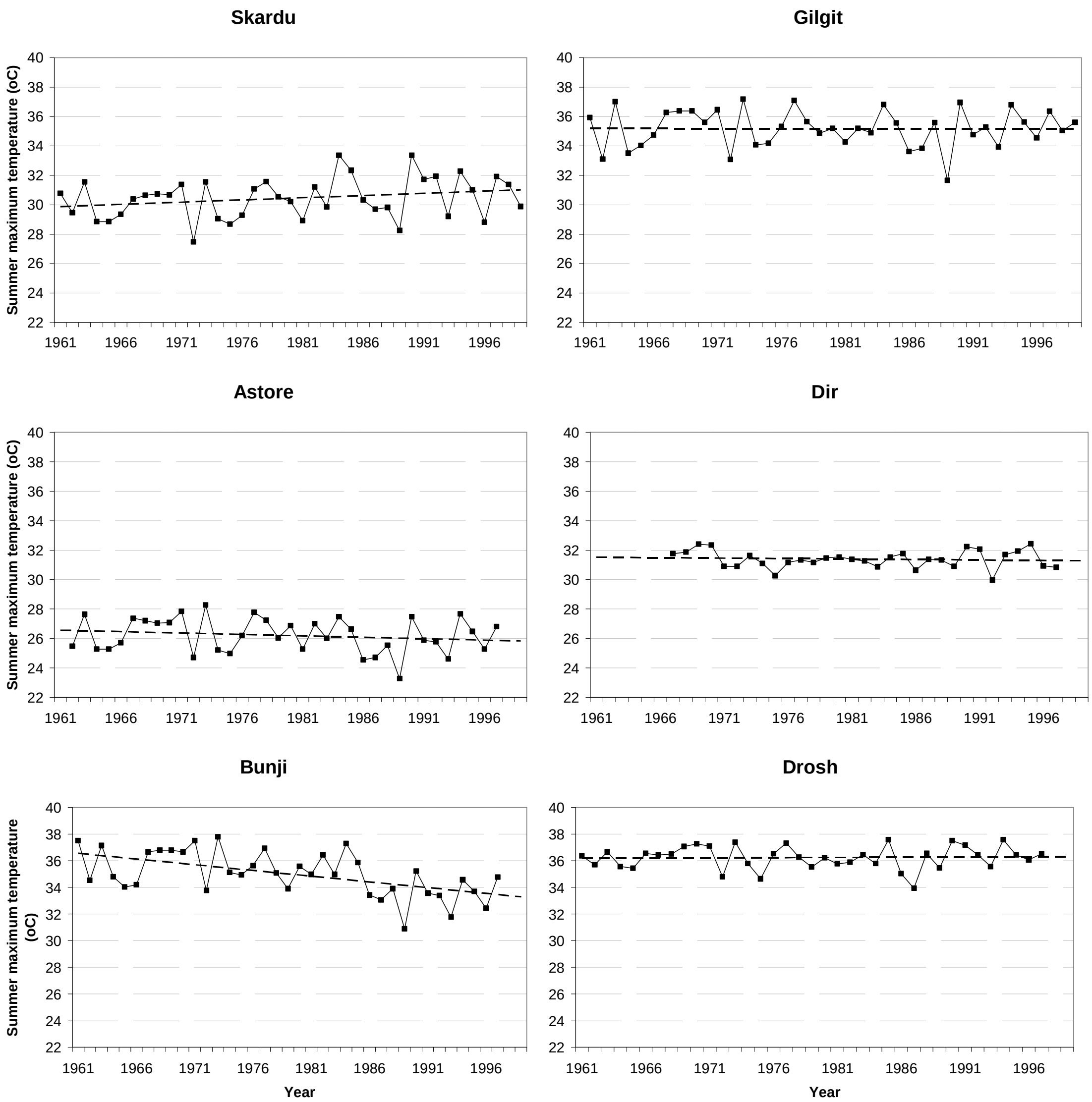

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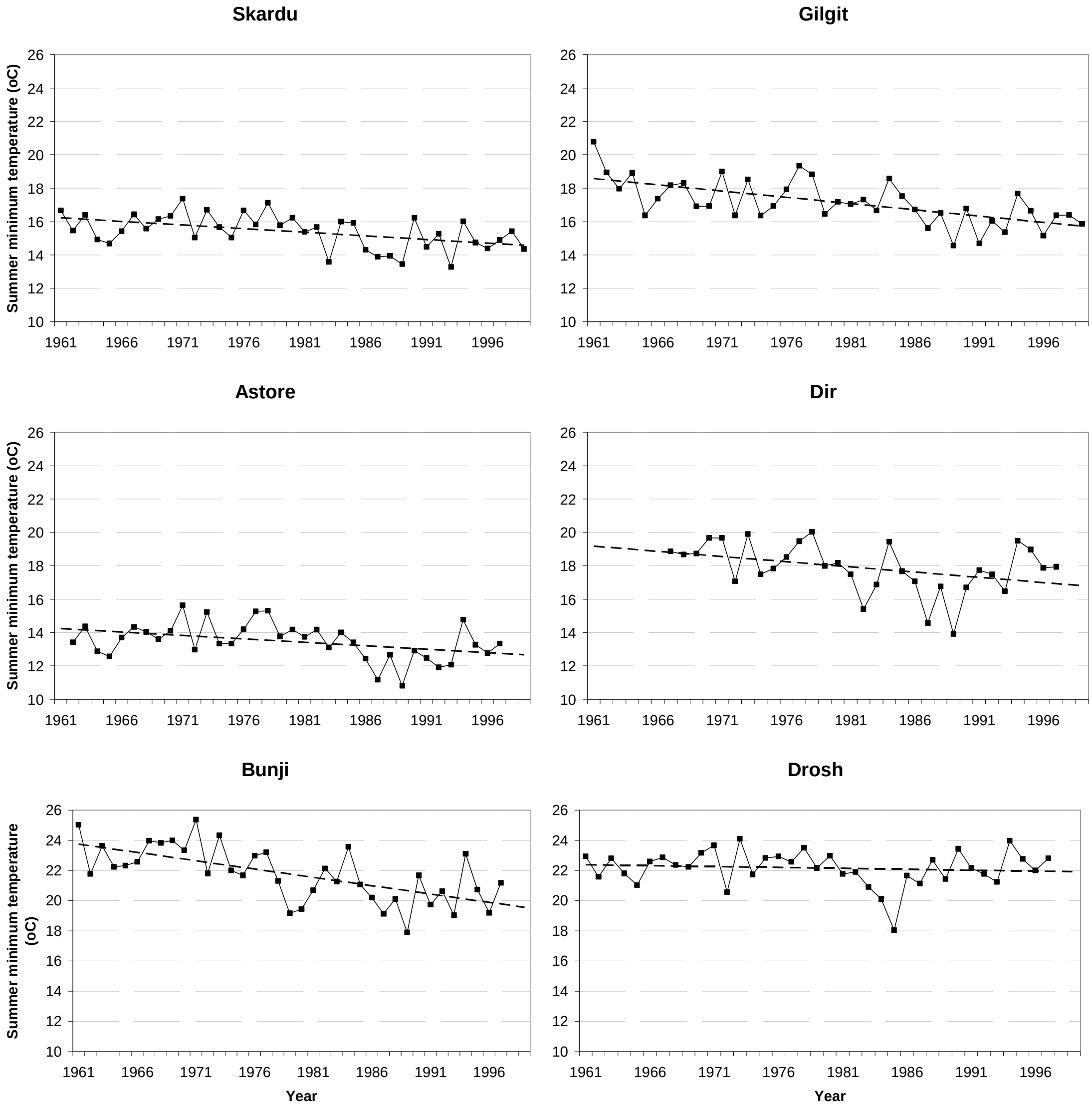

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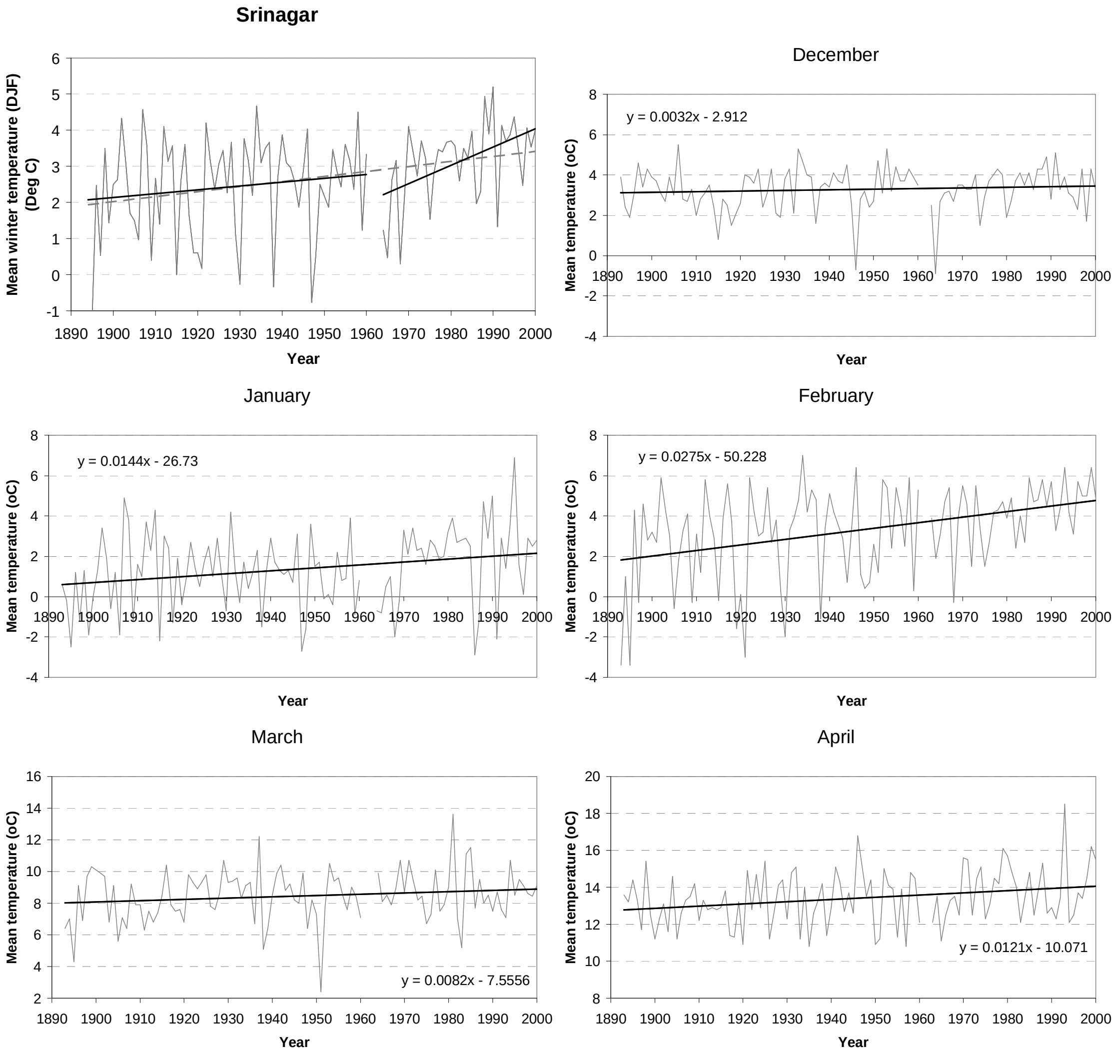

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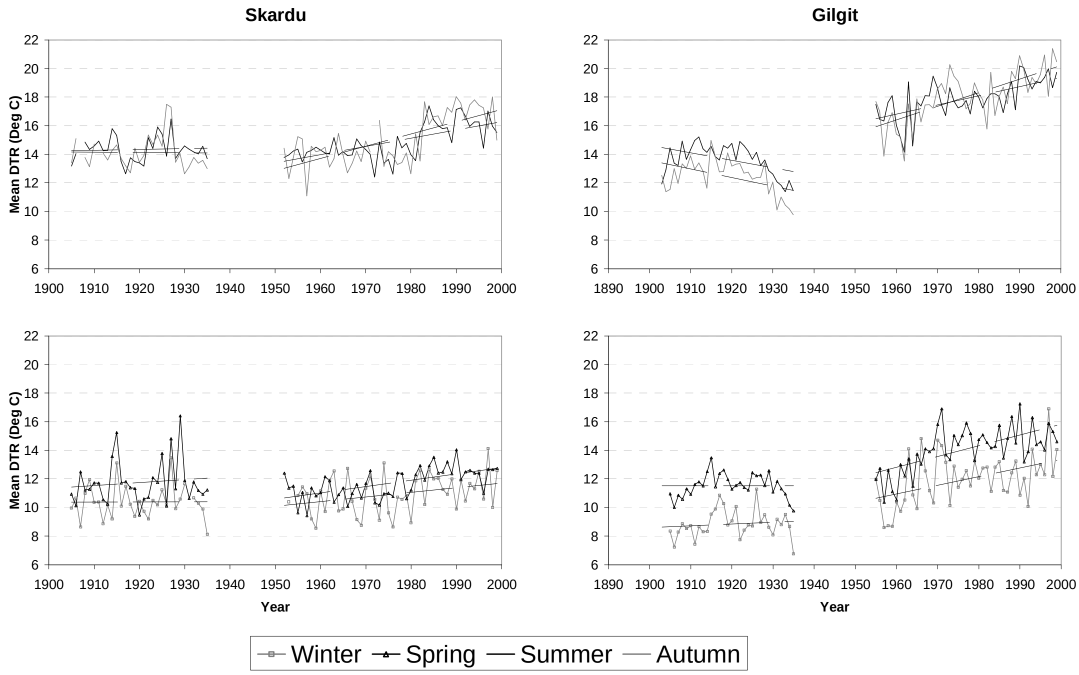

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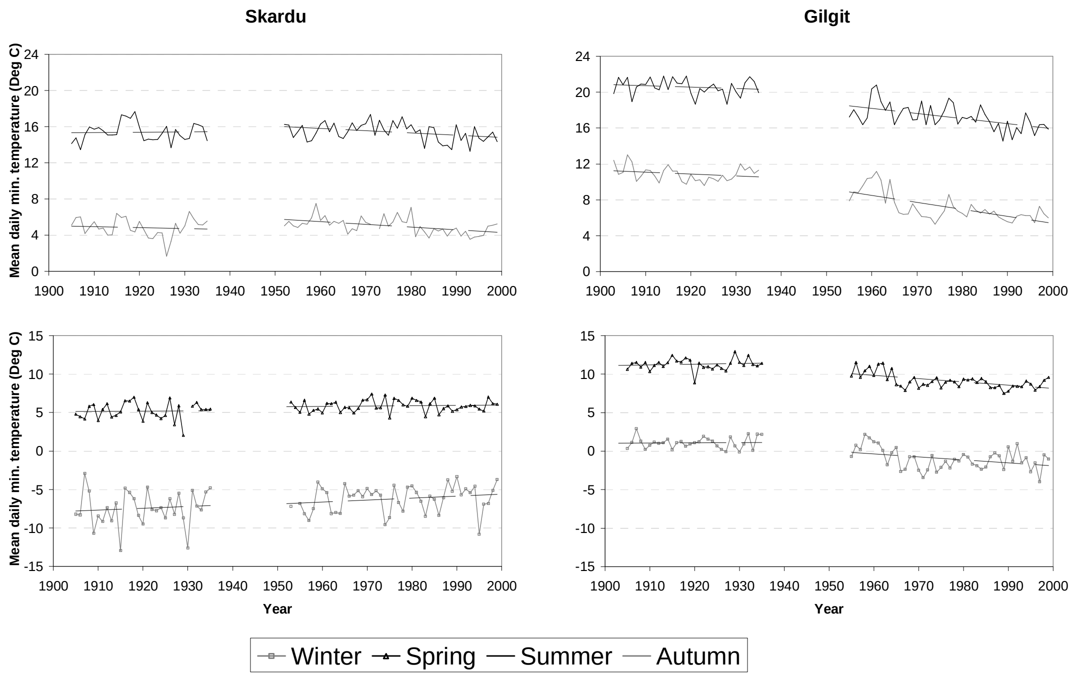

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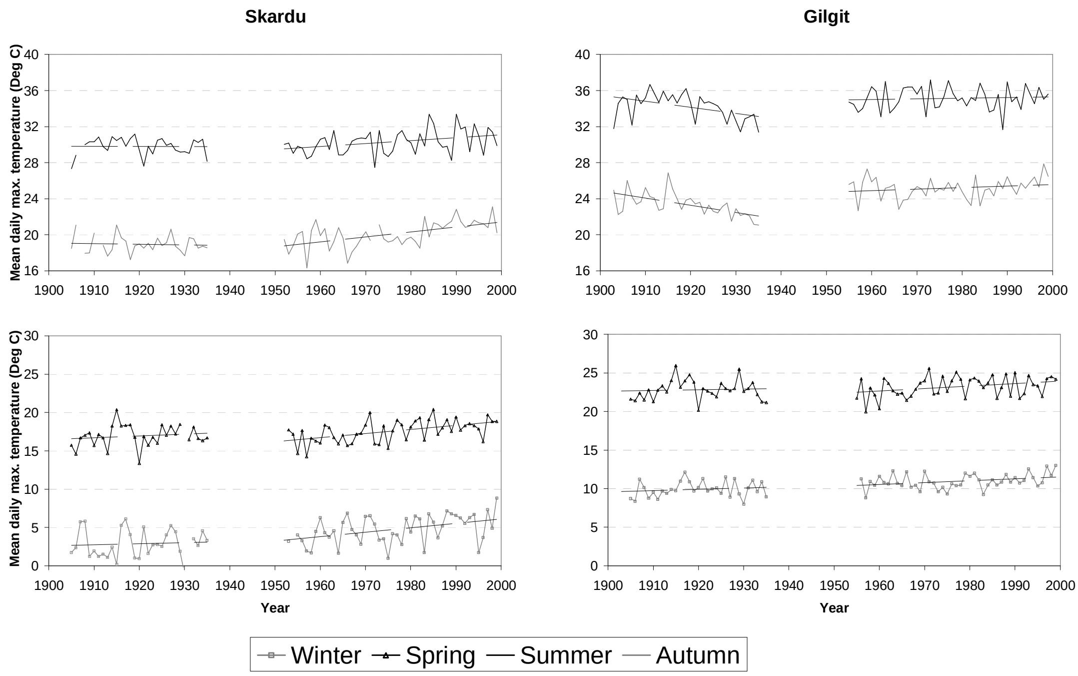

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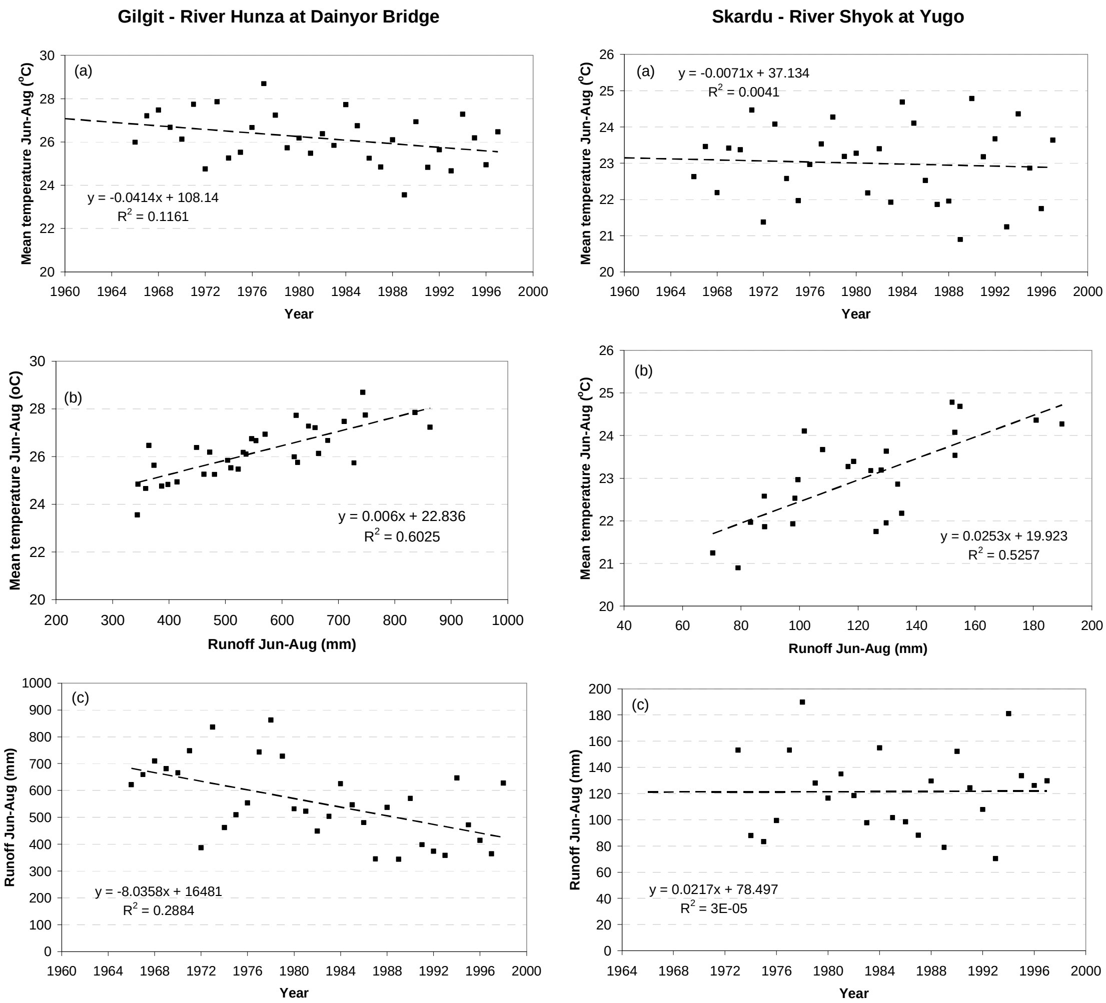

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Table 2 Correlation coefficient (r) of: (a) spring mean temperature (April to June) – upper triangle, and (b) summer mean temperature (July to September) – lower triangle. Summarised from Archer (2004).

Table 3 Trend in annual and seasonal temperature from 1961-1999 for: (a) mean, (b) mean maximum, and (c) mean minimum,, showing change in ^{o}C decade^{-1}.

Table 4 Correlation between monthly precipitation and various indices of monthly temperature at Astore, 1961-1997: (A) maximum temperature, (B) minimum temperature, (C) mean temperature, and (D) diurnal temperature range.

Table 1 Climate stations used in analysis.

No.	Station	Elevation M asl	Record Period	Agency or Source
1	**Dir**	1425	1967-1997	2
2	**Drosh**	1465	1950-1997	2
3	**Gilgit**	1460	1903-1999	1,2
4	**Bunji**	1372	1953-1997	2
5	**Astore**	2394	1954-1997	2
6	**Skardu**	2210	1900-1999	1,2
7	**Srinagar**	1587	1894-2000	3

Agency: 1=IMD 2=PMD 3=CRU

Table 2 Correlation coefficient (r) of: (a) spring mean temperature (April to June) – upper triangle, and (b) summer mean temperature (July to September) – lower triangle. Summarised from Archer (2004).

Station	Astore	Bunji	Drosh	Dir	Gilgit	Skardu	Leh	Srinagar
Astore		**0.77**	**0.82**	**0.77**	**0.93**	**0.80**		**0.84**
Bunji	**0.79**		**0.48**	**0.50**	**0.84**	**0.57**		**0.55**
Drosh	**0.59**	*0.36*		**0.76**	**0.78**	**0.63**		**0.83**
Dir	**0.79**	**0.59**	**0.58**		**0.60**	*0.40*		**0.86**
Gilgit	**0.91**	**0.82**	**0.54**	**0.74**		**0.80**	**0.68**	**0.76**
Skardu	**0.78**	**0.58**	**0.46**	**0.46**	**0.73**		**0.81**	**0.65**
Leh					**0.47**	**0.76**		**0.77**
Srinagar	**0.82**	**0.68**	**0.59**	**0.81**	**0.81**	**0.74**	**0.49**	

Bold figures: significance 0.01. Italic: significance 0.05

Table 3 Trend in annual and seasonal temperature from 1961-1999 for: (a) mean, (b) mean maximum, and (c) mean minimum, showing change in ^{o}C decade^{-1}.

(a)

Station	*Annual*	*Winter (DJF)*	*Spring (MAM)*	*Summer (JJA)*	*Autumn (SON)*
Dir	**–0.26**	–0.03	***–0.41***	**–0.35**	–0.19
Drosh	+0.01	+0.15	+0.02	–0.04	–0.03
Bunji	**–0.52**	+0.07	**–0.32**	**–0.99**	**–0.66**
Gilgit	–0.13	***+0.17***	–0.08	**–0.38**	–0.15
Astore	–0.08	–0.07	–0.06	***–0.30***	+0.01
Skardu	**+0.21**	**+0.38**	***+0.24***	–0.05	**+0.26**
Srinagar	**+0.19**	**+0.51**	+0.07	–0.05	+0.08

(b)

Station	*Annual*	*Winter (DJF)*	*Spring (MAM)*	*Summer (JJA)*	*Autumn (SON)*
Dir	+0.07	**+0.51**	–0.19	–0.06	+0.10
Drosh	+0.05	+0.11	+0.11	+0.03	–0.08
Gilgit	**+0.20**	**+0.27**	+0.16	–0.01	**+0.35**
Bunji	**–0.29**	+0.10	–0.20	**–0.86**	–0.16
Astore	–0.05	–0.07	–0.18	–0.19	***+0.28***
Skardu	**+0.52**	**+0.55**	**+0.46**	+0.30	**+0.75**

(c)

Station	*Annual*	*Winter (DJF)*	*Spring (MAM)*	*Summer (JJA)*	*Autumn (SON)*
Dir	**–0.58**	**–0.60**	**–0.60**	**–0.62**	**–0.48**
Drosh	–0.05	***+0.22***	–0.06	–0.12	–0.04
Gilgit	**–0.45**	+0.02	**–0.32**	**–0.75**	**–0.62**
Bunji	**–0.67**	+0.06	**–0.47**	**–1.11**	**–1.17**
Astore	–0.13	–0.08	+0.06	**–0.41**	–0.08
Skardu	***–0.17***	+0.23	–0.05	**–0.43**	**–0.34**

Notes:

Bold **P < .05**
Bold Italic **P < .10**

Table 4 Correlation between monthly precipitation and various indices of monthly temperature at Astore, 1961-1997: (A) maximum temperature, (B) minimum temperature, (C) mean temperature, and (D) diurnal temperature range.

	Jan	***Feb***	***Mar***	***Apr***	***May***	***Jun***	***Jul***	***Aug***	***Sep***	***Oct***	***Nov***	***Dec***
A	-0.14	-0.09	**-0.56**	-0.32	**-0.67**	**-0.46**	**-0.38**	-0.21	**-0.55**	**-0.64**	**-0.53**	**-0.47**
B	0.22	0.28	-0.29	-0.27	**-0.68**	-0.30	-0.19	-0.25	-0.33	-0.33	-0.13	**-0.38**
C	0.07	0.13	**-0.47**	-0.33	**-0.70**	**-0.40**	-0.31	-0.24	**-0.50**	**-0.58**	**-0.43**	**-0.45**
D	**-0.53**	**-0.40**	**-0.60**	-0.22	**-0.49**	**-0.35**	**-0.40**	-0.02	**-0.41**	**-0.58**	**-0.55**	-0.11

Note: **Bold** **$p < 0.05$**