CLIMATE CHANGE AND RIVER FLOW IN PARTIALLY-GLACIERISED HIMALAYAN CATCHMENTS

Submitted in partial fulfilment of the requirements of the Degree of Master of Science

by

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Declaration

This is to certify that the copy of my thesis, which I have presented for consideration for my postgraduate degree embodies the results of my own course of research, has been composed by myself and has been seen by my supervisor before presentation.

The thesis contains 21,456 words (excluding references).

Signature Date

Abstract

Climatic and hydrological data for stations in the Himalayan headwaters of the Ganges and Indus rivers were obtained where available from national meteorological institutes and bodies responsible for river flow gauging. The aim was to develop a database of climatic variables with long homogenous series at monthly resolution for periods of up to 150 years, and of river flow data for as long as available for rivers with natural flow, i.e. upstream of dams which regulate discharge at seasonal timescale. Glaciologically-relevant climatic variables which influence glacier mass balance were taken from records for stations at high elevations generally in the range 1000 to 2000 m a.s.l. and up to 4531 m a.s.l. along the Himalayan arc. A north-west to south-east transect was combined with two stations on the Tibetan plateau. Hydrological observations with long term records cover rivers tributary to the Indus and Ganges, which drain partially-glacierised basins. The stations are located are at elevations between 191 and 3688 m a.s.l. and have records of between 8 and 45 years in length. Mean annual and summer (May through September) temperatures across Himalayan arc increased with initial warming between 1900 and 1950, followed by dips until around 1980, before recovering to levels just above 1950 maxima. Total winter precipitation between November and May in north-west regions of the Himalaya was stable at many stations but declined 18% and 41% at Srinagar and Shimla respectively. Total monsoon (summer) precipitation (June through September) declined considerably, by around 50% in the western and central Himalaya but 35-40% in far eastern Himalaya.

River flows declined also in this period. In the upper Indus tributaries, flows reduced by up to 24 % between the 1960s and 1990s, whilst the upper Sutlej River declined by about 30% in the same period by comparison with about 10 % in central Himalaya of Nepal. Summer temperatures have been insufficient to enhance snow and icemelt to offset falling flows resulting from reducing precipitation. Precipitation is the main driver of river flow in the Himalaya as the monsoon covers wide swaths of -discharge fail to offset declining monsoon precipitation.

1 Introduction

High mountain regions of the world contribute large quantities of runoff, derived from seasonal snow pack and glacier ice, to the headwaters of the world's major rivers. Such mountainous areas are seen as water towers for surrounding regions (Viviroli, 2007), as their contributions to runoff are much greater than those from lowland areas. Snow- and ice-melt provide disproportionately large quantities of downstream river flow in both varying proportions and absolute amounts through time. Along the Himalayan arc, runoff from snow- and ice-melt heads downstream into the major rivers – the Indus, Ganges, and Brahmaputra. As precipitation is generally enhanced at higher elevations, large amounts of river flow are provided from mountain basins by comparison with those at lower elevations. Should Himalayan glaciers continue to retreat, water shortages might become widespread within a few decades (Immerzeel, 2008). Climate change may greatly influence discharge from glacierised catchments, with river flow varying depending on global and regional warming. As temperature reduces with elevation, decreasing 0.6°C for every 100 metres (Barry, 1992), most winter and some summer precipitation will fall as snow. Glaciers are present where annual snowfall is in excess of the amount that summer heat energy is available to melt. If annual snowfall experiences a downward trend and available heat energy increases then glaciers recede. Relative changes in meteorological parameters are important in understanding whether glaciers are receding or advancing over a particular period of time as they are indicative of the extent of the ice loss.

Climate change is thought to have an accelerated impact on air temperature and precipitation in high elevations basins where much of the flow is derived from the melting of seasonal snow pack and glacier ice. Beniston (2010) recognised that mountains are particularly sensitive physical environments, and climate change may affect the quantity and quality of Himalayan freshwater resources as Earth's atmosphere warms (Singh and Kumar, 1997). Along the Himalayan arc total discharge does not rely only on meltwater from winter snowpack and glacier ice, but is also heavily dependent on the amount of monsoonal precipitation in summer. As precipitation increases with elevation, the Himalayan reaches of the Indus, Ganges and Brahmaputra rivers contribute disproportionately to flow downstream in summer months. The IPCC report (2007) highlighted concern that glaciers in the Himalayas were receding faster than in any other parts of the world. If the present rate of retreat was to continue, the IPCC considered that glaciers might disappear altogether by 2035. However, this statement arose from an 'informal' publication by the World Wildlife Fund. The WWF later retracted the statement saying that this statement was used in good faith but it is now clear that this was erroneous and should be disregarded (WWF, 2010). Across the Himalayas, future meteorological changes will impact on runoff levels of the three major rivers as increasing melt from snow pack and glacier ice will greatly alter as precipitation and temperature trends also change. The Himalaya can be divided into four main regions; A. North-west, B. Tibetan Plateau, C. Eastern catchments and D. Central as shown in the map by Collins and Hasnain (1995) in figure 1.3.

In high mountain areas, winter precipitation accumulates to form a stable winter snow pack. In spring snow starts to melt, contributing to runoff as the transient snow lines rises up basins. In partially-glacierised catchments, the rising transient snowline exposes glacier ice to melt, runoff continuing to increase as the bare ice area expands until energy levels fall towards the end of summer. Maximum discharge levels depend on glacier area and winter precipitation, with air temperatures rising through spring into summer, as snow and ice reservoirs decline reducing the quantity of melt but increasing the contributing area for rainfall (Collins and Young, 1981). Higher runoff levels are generally experienced in warm summers following winters with little snow cover, with river flows reduced greatly in cool summers which follow much winter snowfall. There is an increase in the thickness of ice melting when the transient snowline rises sooner and higher in early summer leaving a larger area of ice exposed to melt for longer (Collins, 2009).

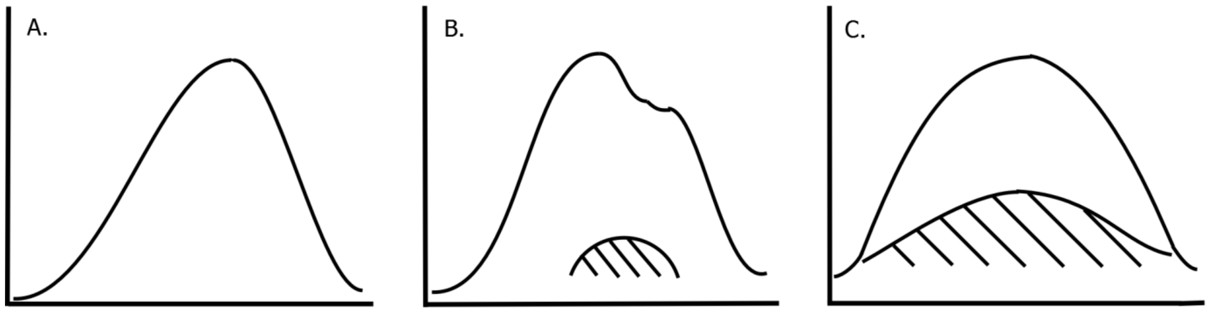

Figure 1.1 Hydrological regimes with precipitation (shaded) for three Himalayan catchment types (A. North-West B. Central and C. Eastern).

Hydrological regimes across the Himalaya are determined by glacial melt and monsoon precipitation. Thayyen and Gergan (2010) define three different types of catchments the Himalayas: cold arid, summer monsoon and winter snow regime. Monsoon precipitation reduces in strength and quantity from east to west, with runoff regimes in central and eastern Himalaya dominated by summer precipitation. Hydrological regimes are dominated by south-west Indian monsoon in summer and mid latitude westerlies known as western disturbances during winter months (Upadyaya 1995; Mani 1981). In the north-west Himalaya and Karakorum, runoff increases with snow and ice melt as available heat energy rises, until peaking in late July and early August (Graph A, Figure 1.1). In glacierised Alpine catchments such as Gornera and Findelenbach in Switzerland, discharge peaks in late July or early August (Collins, 1998), which is similar to the winter snow regime of the north-west Himalaya. Monsoon precipitation rarely impacts upon the north-west region, with winter westerlies providing large amounts of snow accumulation contributing to runoff through summer as temperatures warm. Rees and Collins (2006) found that melt water in the western Himalayas remained a major component of runoff downstream at low elevations as a result of otherwise arid conditions.

The central Himalaya form a region where summer monsoon starts to dominate runoff, as summer precipitation gradually increases from west to east. River flow is enhanced in summer as the monsoon cuts in but then dips into August (Graph B, Figure 1.1). Runoff decreases during August because the transient snow line is lowered due to summer snow fall, which leads to a decline in discharge from glacier melt upstream due to the higher albedo of snow and reduced radiation coupled with increased cloud cover.

A typical hydrograph for the eastern Himalaya is a dome shape (Graph C, Figure 1.1) with no sharp peak during summer due to 3-4 months of consistently high monsoonal downpour. From June through to September large quantities of precipitation from the Bay of Bengal contribute heavily to runoff across the eastern Himalaya. Bookhagen and Burbank (2010) estimated that the snowmelt contribution to total runoff in the western Himalaya was 50% by comparison with 20% in the eastern and central Himalaya. Runoff contributions in central and eastern Himalaya are driven by summer monsoon, while in the north-west depend on snow and glacier ice melt.

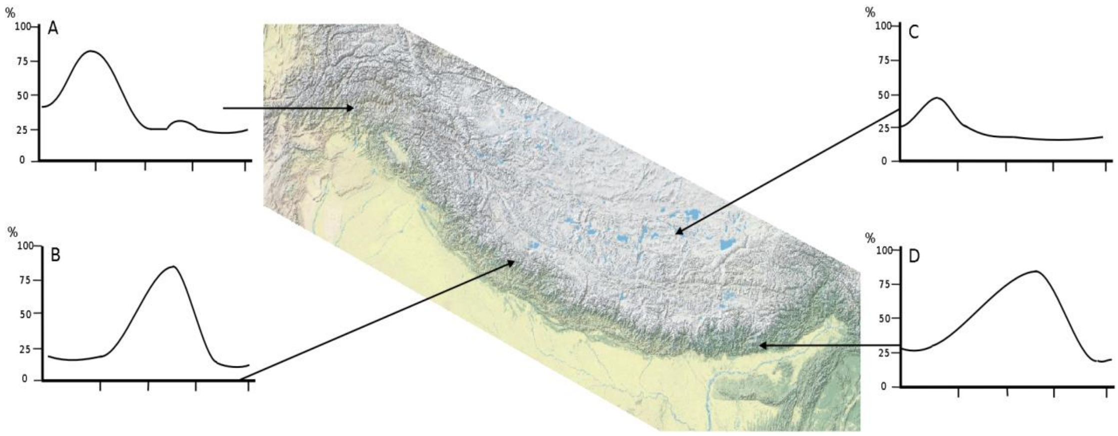

Figure 1.2 Comparative seasonal distribution of precipitation amounts in four precipitation regimes of the Himalaya (A. North-west B. Central C. Tibetan Plateau and D. South East).

Across the Himalaya precipitation differs with westerly winds providing a large amount of winter precipitation and summer monsoon dominating central and eastern catchments. Along the Himalayan arc, the length and quantity of the monsoon season generally declines from south east to north-west (Collins and Hasnain, 1995). The north-west Himalaya receives huge quantities of winter precipitation, reaching maxima in February and March before decreasing and levelling out until rising again in December (Graph A, Figure 1.2). Seasonal precipitation regimes in the north-western Himalaya are similar to those of Alpine catchments, as both receive large amounts of winter precipitation, followed by the melting of the winter snowpack and glacier ice in summer. Precipitation does slightly rise during summer in July and August but not enough greatly to affect runoff (Graph A, Figure 1.2).

In the eastern Himalaya monsoon precipitation starts to rise from June staying high through to September, but quickly declines with precipitation low from October until May next year (Graph D, Figure 1.2). In the Central Himalaya, summer precipitation quantity is reduced, peaks being lower than in the eastern Himalaya (Graph B, Figure 1.2), as the monsoon season starts later and shortens from east to west.

In winter, precipitation falls as snow at higher elevations in the central and western Himalaya (Graph B, Figure 1.2). Central Himalayan catchments are transitional zones as the summer monsoon weakens towards the west, and winter precipitation extends east from in the western regions of the Himalaya.   The Tibetan Plateau receives little annual precipitation by comparison with the south slope of the Himalaya, with only a slight increase in February and March (Graph C, Figure 1.2).

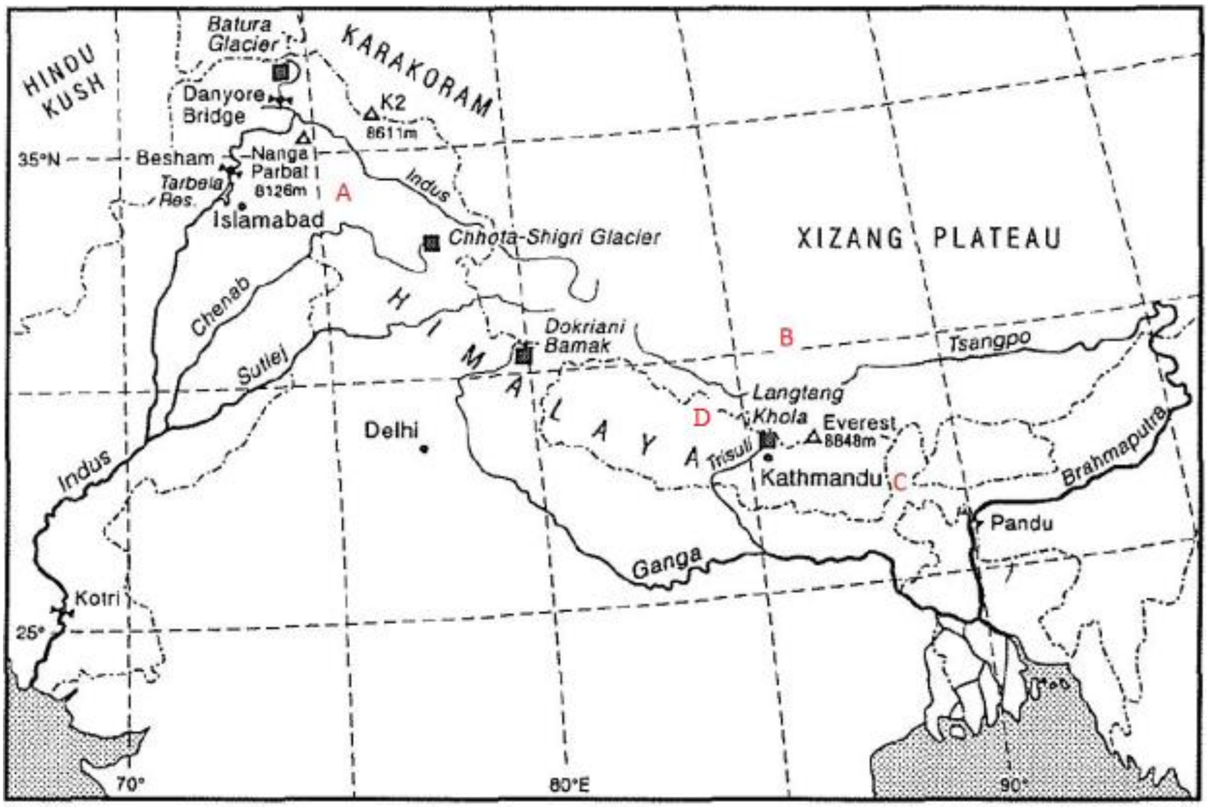

Figure 1.3 Map of Himalayan arc with locations of precipitation regimes A. North-west B. Tibetan Plateau C. South-east and D. Central Himalaya (Source: Collins and Hasnain, 1995).

Air temperature is an important parameter in determining hydrological changes in partially and highly glacierised basins. Across the Himalaya, July and August are generally the warmest months (Thayyen & Gergan, 2010), when heat energy will be greatest for melting remaining winter snowpack and glacier ice, mean monthly air temperatures ranging from 11.4-9.5 °C at 3,760 m a.s.l., 13.4-11.2°C at 3,400 m a.s.l. and 18.5-16.0°C at 2,540 m a.s.l. (Thayyen & Gergan, 2010). As temperatures decline with elevation, precipitation generally falls as snow at higher elevations especially during winter, and the possibility of snow falling during the summer monsoon. As temperatures warm during summer, runoff increases as available energy is enhanced leaving a larger area of ice exposed to melt for longer in the summer months. However, cooler summer temperatures will expose less ice and therefore result in less runoff. This is typical of north-west Himalayan basins in which the majority of runoff is contributed from snow and ice melt. In monsoon-fed catchments runoff may stay fairly high during cool summers with a large amount still contributed from summer precipitation. However, summer monsoon falling as snow can greatly reduce discharge as cloud cover increases, leading to a dip in the runoff as heat energy available for melt declines.

The heating effect is much greater on the south side slopes of the Himalaya than on the north side. The high Himalayan mountains divide moist warm air of south Asia from the cooler drier air to the north on the Tibetan Plateau (Figure 1.4: Cane, 2010). The Himalayan mountain range acts as barrier between the south and north sides, with warmer temperatures in south Asia driving the monsoon, and the maritime air mass giving favourable conditions for moisture driven convection (Figure 1.4: Cane, 2010). Temperatures increase with distance away from the Indian Ocean. Overall temperatures in the eastern Himalaya tend to be warmer by comparison with the north-west Himalaya which is situated much further north.

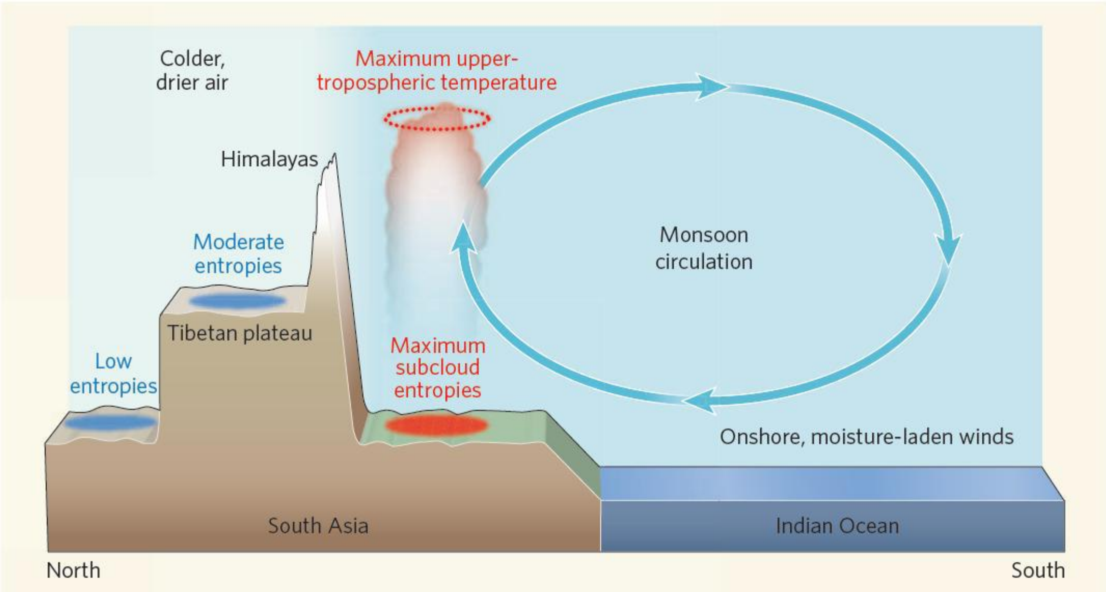

Figure 1.4 A new monsoon model (Source: Cane, 2010)

Future climatic warming is expected to be accelerated across the Himalayan mountain range. The Himalaya experienced rising temperature trends in the 20th century (Shrestha *et al.*, 1999; Immerzeel, 2008), which enhanced the melting of winter snowpack and glacier ice. Mann *et al.*, (1998) identified that northern hemisphere temperatures steadily but not significantly declined from 1000 AD

(Figure 1.5) so that before 1850 AD, glaciers would have experienced growth in both length and area. Lower temperatures and more of the precipitation falling as snow encouraged glacier growth. Then, throughout the 20th century, temperatures warmed rapidly (Figure 1.5: Mann et al., 1998), inevitably leading to changes in river flow. Future hydrological regimes will greatly alter as glacier volume and area reduce, with less dependence on snow and ice contributions. As glaciers shrink initially runoff levels will be enhanced, with increased energy to melt snow and ice. However, in the long term discharge will be reduced over the next few decades as heat energy increases become insufficient to offset much reduced ice areas. As the IPCC (2007) highlighted, enhanced melting of glaciers will first lead to increased runoff, discharge peaks and increased snow melt. However, as the amount of snow and ice reduces, runoff will decline to levels commensurate with future levels of precipitation.

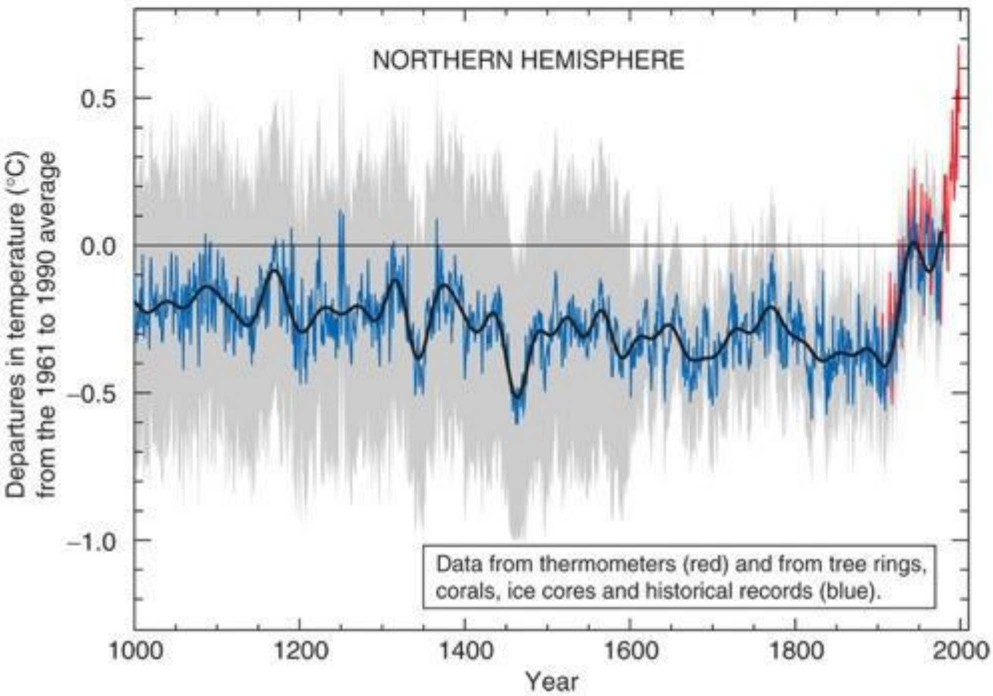

Figure 1.5 Hockey Stick diagram showing northern hemisphere temperature changes from 1000 AD (Source- Mann *et al.*, 1998).

2 Aims and Objectives

The aim of this thesis is to fill existing gaps in current knowledge, as there has been no study examining the relationship between meteorological conditions and glacier runoff across Himalayan arc. Such studies have been carried out in other mountain areas - for example by Collins and Young (1981) in a highly-glacierised alpine basin. To achieve this aim, data from many stations in various different Himalayan river catchments were examined in order to identify impacts of climatic variation on runoff. Snow and ice-related meteorological variables with long homogenous series have been derived in order to indicate the extent to which fluctuations in precipitation and air temperature have influenced river flow, as it is thought that climate change is accelerated with increased elevation, and that Himalayan glaciers are receding more than in other parts of the world (IPCC, 2007). The extents to which air temperature and precipitation have changed differentially over a range of elevations have also been investigated.

The Himalaya arc extends between 38° N and 28° N and from 70° E to 95° E, accompanied by considerable regional differences in the quantity and timing of precipitation. Stations were selected across the region to characterise air temperature and precipitation, both of which strongly influence runoff. Gauging station records were examined in order to provide information about runoff at various locations along the Himalayan arc. These stations should allow determination of the extent to which the winter westerlies and summer monsoon influence runoff in basins of different percentage glacierisation of basins. The study aims to examine trends in meteorological variables since 1860, the time of maximum glacier extent in the Little Ice Age.

3 Study Area

The Himalayan region is situated between 38^0 N - 28^0 N and 70^0 E – 95^0 E according to Collins and Hasnain (1995). The study area is located within these coordinates, with the Himalayan arc stretching over more than 2000 km, encompassing mountain foothills that run along south side of the Himalayan arc, and high mountain peaks that separate India from the Tibetan Plateau. The Karakoram is the most north-westerly mountain range, contrasting with the monsoon fed catchments of the Brahmaputra which form the most easterly region of the Himalayan arc. The Himalaya divides into four major regions that are the Tibetan Plateau, and the north-west, central and eastern Himalaya. Miller *et al.,* (2011) divided the Himalaya into three catchments defined by the major rivers Indus, Ganges and Brahmaputra (Figure 3.1). However, these catchments are not necessarily defined by a particular topographic boundary. As the western Himalaya is influenced by both winter and summer precipitation, it is defined as one region in this study, although with basins tributary to both Indus and Ganges. The stations chosen for study should accommodate these regions to highlight parallel and dissimilar trends. Sites must be distributed also over a range of elevations, to see whether climate change is enhanced with altitude.

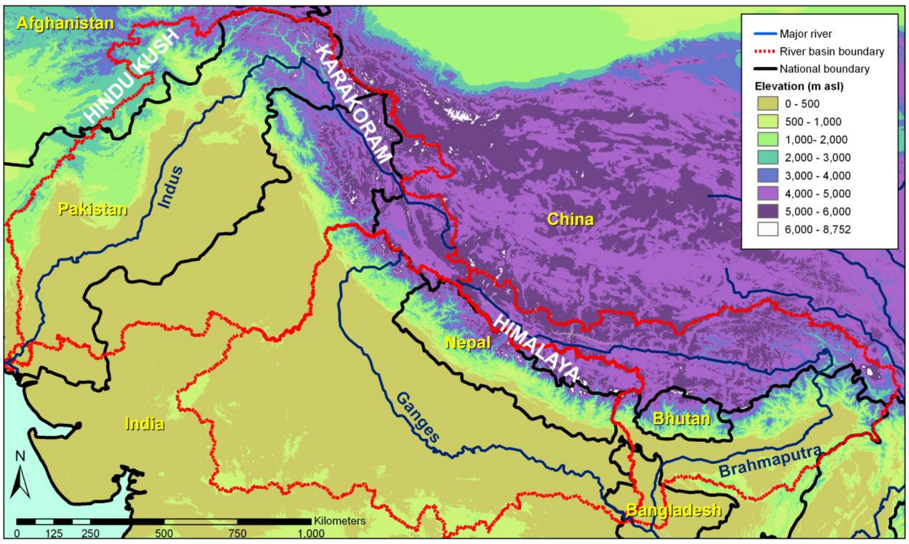

Figure 3.1 Map of the three major rivers of Ganges, Brahmaputra and Indus along with their catchment areas (Source- Miller *et al*., 2011).

3.1 North-west and western Himalaya

The North-west Himalaya is situated between 70^0-78^0E and 30^0-37^0 N, stretching from Karakoram mountain ranges to far north, east to cover Jammu & Kashmir and Himachal Pradesh. This area is at high elevation, with peaks reaching up to 8 000 m in the Karakoram (Bhutiyani *et al*, 2010), compared to the foothills that are only a few hundred metres above sea level. The Indus River is the main conduit for runoff with its main tributaries Sutlej, Chenab, Ravi, Beas and Hunza. Winter westerly winds provide a large amount of precipitation from October to April. Annual hydrological regimes of the north west Himalaya are similar to Alpine catchments, as high winter precipitation is followed by a summer melting season that results in discharge peaks around late July or early August. The Hindu Kush and the Karakoram region receives between 200 and 500 mm of annual rainfall (Akhtar *et al.,* 2008). However, these records were taken at valley based stations at much lower elevations. In high elevation regions above 4000 m, precipitation can range anywhere from 1000 mm to more than 3,000 mm (Akhtar *et al.,* 2008). Higher altitude areas generally receive a large amount of winter precipitation, with snow and ice melt contributing a larger proportion towards runoff in the surrounding areas. Immerzeel *et al*., (2009) highlighted that snowmelt contributes 40% towards the Indus runoff, by comparison to 32% from glacial melt. In the western Himalaya snow and glacier melt provides less contribution with 59% in the Sutlej River (Singh and Jain, 2002). In the Western Himalaya summer precipitation starts to influence runoff levels more, but snow and ice melt still provide large quantities of water towards discharge. The Sutlej river in Himachal Pradesh drains into the Indus further downstream. The Sutlej differs from the other tributaries in the north-west where winter precipitation contributes the majority towards runoff. Nonetheless, the Western Himalaya is considered to be a region, as throughout the annual hydrological regime is influenced by summer monsoon.

3.2 Central Himalaya

The central Himalaya lies between approximately 78^0-85^0E and 30^0-28^0 N, covering the Uttarakhand region of India and Nepal to the east. It is hard to define where the western Himalaya ends and central Himalaya starts, and also which far eastern districts of Nepal can be defined under eastern Himalaya rather than central. High mountains range between 3,000-7,000 m and the foothills measure from 600-1,200 m (Basistha *et al.*, 2009). The Ganges flows through Uttarakhand in the west central Himalaya and then heads south-east towards Bangladesh. The Bhagirathi, Karnali, Gandaki, Kosi and Yamuna Rivers are the main central Himalayan tributaries of the Ganga. The upper Ganges catchment is a transitional zone as summer monsoon strengthens from west to east, but it is still influenced by winter westerlies in the western central Himalaya. Summer monsoon amount increases during August, but discharge dips during this month. Due to the transient snow line lowering as a result of snowfall at high elevation, runoff decreases with the higher albedo of snow coupled with cloud cover. Average precipitation in Nepal is 1768mm (Shrestha *et al*., 2000), but this will greatly differ regionally and with changes in elevation and aspect. The Annapurna range receives around 5 m of annual precipitation compared to 1948 mm at Mukkim at 1900 m a.s.l. (Shrestha *et al*., 2011). Across the central Himalaya, runoff variations occur as summer monsoon declines from east to west, reducing the percentage contribution of precipitation to runoff. Bookhagen & Burbank (2010) highlighted that snowmelt only contributes 20% of river flow in the central Himalaya by comparison to 30% for the Ganga at Devprayag (Singh *et al*., 1997).

3.3 Eastern Himalaya

The Eastern Himalaya region is located within 85°-95° E and 25°-28° N, including lowland floodplains and high mountain ranges across China, India, Bangladesh and

Bhutan. The Brahmaputra is the main river that starts along Tibetan Plateau, which flows east across southern Tibet until reaching high mountains peaks, then turning southwest towards Assam and Meghalaya, eventually joining the Ganges in Bangladesh. 44.4% of the Brahmaputra basin lies on the Tibetan Plateau by comparison with 28.6% and 27% for the Himalayan mountain belt and floodplain respectively (Immerzeel, 2008). The Ganges is the other main river dominated by summer monsoon with some glaciers in the high mountains east of Nepal, Sikkim and Bihar. Eastern Himalayan climate is characterised by summer monsoon from June to September, which provides 60-70% of its annual rainfall (Immerzeel, 2008). The south-west Indian monsoon during summer leads to consistently high discharges for 3 to 4 months with the greatest levels experienced during July and August. Snowmelt accounts for 20% of the total runoff in eastern Himalaya and 80% from rainfall (Bookhagen and Burbank, 2010). Annual precipitation levels are largely diverse across the eastern Himalaya, floodplains receiving the most with 2354mm by comparison with 1349mm and 734mm in the Himalayan belt and Tibetan Plateau respectively (Immerzeel, 2008).

3.4 Tibetan Plateau

The Tibetan Plateau stretches along the north side of the Himalayan arc from 77^0-91^0 E to 28^0-35^0 N. The plateau is a consistent high elevation region, across the autonomous region of Tibet, now in China. Whilst both the Indus and Brahmaputra rivers start on the plateau, this region provides minor contributions to downstream runoff. The Tibetan Plateau is very dry and arid due to the Himalayan mountain ranges preventing moisture reaching inland. The high mountains act as a barrier Cane (2010). With limited precipitation across Tibetan Plateau, there is insufficient snow to sustain large glaciers.

4 Background

4.1 Little Ice Age

Glaciers in high mountain regions reached their maximum recent extent around 1850. It is difficult to know the coverage of glaciers as there are practically no documentary records relating to length and mass before the nineteenth century (Grove, 1988). Zemu glacier in Sikkim, eastern Himalaya was about 150 metres advanced at 4200 m a.s.l. in 1891 for example (Meyewski and Jeschke (1979). These recordings were basic with only estimates of ice mass making it difficult to identify Little Ice Age maxima. Freshfield (1903) identified how difficult it was to measure glacier length, when moraine lines were all over the place 8 years later. The Indian Geological Survey recognised the need to record glacier termini by sending Touche (1910) to map the position of Zemu glacier. Since 1850, glaciers in the Sikkim district lost length and mass at Zemu (Bose *et al.*, 1971), Jungpu (Meyewski and Jeschke, 1979) and Khumbu glaciers (Miller, 1970). Glaciers have declined since Little Ice Age maxima, but with very simple recording methods and limited accuracy it is hard to identify the true extent of ice loss.

In Uttarakhand, central Himalaya Milam glacier reached Milam village a thousand years ago, but by 1906 the terminus was a mile away according to local tradition (Cotter and Coggin-Brown, 1907). Evidence based upon locals passing down knowledge through many generations is unreliable and gives no accurate measurements. Strachey (1900) identified about 2 miles between ice caves and median moraine for the Pindari glacier compared to Cotter and Coggin-Brown (1907), who recognised that the moraine terminated about a mile from the ice cave. Estimations of glacier recession greatly contradict each other with varied measurement techniques and sources of local information. During the twentieth century, observations in the central Himalaya became more established. Jangpangi and Vohra (1962) pointed out that Milam glacier receded 617m from 1906 to 1957, and further 23 m between 1963 and 1964 (Kumar *et al*., 1975). Gangotri glacier observed even more loss with a 600 metre loss in a short time period from 1936-1967, with some tributaries retreating more than the main glacier (Tewari, 1970). Across central Himalaya records identified a clear reduction in glacier length and mass, but measuring techniques started to improve into the 20th century with more precise data collection.

Bara Shigri glacier declined annually by about 60 m (1890 to 1906), 20 m (1906-1945), 28 m (1955-63), 6 m (1963-1980) and a total retreat of 2.6 km over 90 years (Mackley and McIntyre, 1980). Further west, Jammu and Kashmir region experienced similar patterns of glacier recession after 1850, as Kolohoi glacier retreated quarter of a mile between 1887 and 1912 and more than a mile since 1857 (Neve, 1907). Chungpar glacier lost about half a kilometre from 1856 to 1930s (Kick, 1980), after reaching its maximum in 1850 when a lake was dammed back (Schlagintweit and Schlagintweit, 1856). Kolohoi glacier continued to decline with another 800 metres between 1912 and 1961 after the initial 800 loss from 1857 to 1912 (Meyewski and Jeschke, 1979). Mountain basins in north-west and western Himalaya observed glaciers reducing in size as climatic conditions changed since the Little Ice Age maxima.

4.2 Precipitation

Along the Himalayan arc precipitation trends greatly differ, and the extent to which summer monsoon and western disturbances influence growth and decline of glaciers is still subject to speculation (Benn and Owen, 1998). If the amount of winter precipitation is sufficient, glaciers may remain large enough to provide significant quantities of water downstream. Role of glaciers in precipitation dominated catchments is to enhance stream runoff during years of low summer rainfall in the western Himalaya, e.g. Gad catchment (Thayyen and Gergan, 2010). Periods of glacier growth are closely linked to strong monsoon seasons, which lead to high flows in eastern Himalaya (Thayyen and Gergan, 2010). Bookhagen and Burbank (2010) outlined that precipitation increases by six fold from east to west and tenfold on a south to north gradient across the Himalaya. Central Himalaya receives 80% of the annual rainfall from May to October, whereas in the Tibetan Plateau it only accounts for 50% in western and eastern syntaxes where the Indus and Tsango rivers exit the range (Bookhagen and Burbank, 2010). Bookhagen and Burbank (2010) identified that Sutlej valley catchment receives about 70% of its annual rainfall during summer (June to September), but to the west it accounts for less than 50%. Kumar and Jain (2010) identified that summer monsoon accounted for 21% compared with 43% for winter precipitation in the Kashmir Valley. When looking at long term precipitation trends it is important to look at summer monsoon patterns in central and eastern Himalaya, but winter precipitation variations in the north-west. However, there will be a transitional zone where both summer and winter precipitation influence river flow.

4.3 Precipitation trends

Shekhar *et al*. (2010) observed a decline in winter precipitation across the north-western Himalaya from 1988/89 to 2006/7 at Pir Panjal and Shamshawari. As winter precipitation reduces, glaciers shrink due to insufficient snowfall to maintain glacial mass. Snowfall has not greatly changed from 1988/89 to 2006/07 at sites across Karakorum (Shekhar *et al*., 2010). With little change in winter precipitation, glaciers in north-west Himalaya will have not reduced much with sufficient rates of snowfall. Bhutiyani *et al*. (2010) showed a decline in monsoon and annual precipitation from 1866-2006 in north-west. However, Bhutiyani *et al*. (2010) highlighted an increase in winter precipitation. Increasing winter snowfall in this region is more important because it mean that glaciers could sustain their length and mass. Kumar and Jain (2010) outlined that in the Kashmir valley Srinagar, Kulgam and Handwara experienced a decrease in annual, monsoon and winter rainfall from 1903 to 1982.

However, Srinagar showed an increasing trend in annual rainfall from 1962-2002 (Kumar and Jain, 2010). Precipitation across north-west Himalaya seemed to be declining, but more recently rising, as Bhutiyani *et al*. (2010) observed a much wetter period since 1966.

Over the last century, rainfall in the Uttarakhand region decreased (Basistha *et al.,* 2009) with a similar pattern across north-west Himalaya. However, a wetter period occurred between 1965 and 1980 (Basistha *et al*. 2009). Basistha *et al.,* (2009) and Kumar and Jain (2010) both identified an increasing trend from 1960 onwards, outlining an uncertain picture about precipitation across western and central Himalaya. In the eastern Himalaya, Immerzeel (2008) found no distinct drying or wetting trends over last 100 years in Brahmaputra basin. Gautam *et al.,* (2009) noted that early summer monsoon rainfall over India has risen over 20% from 1950-2004. That Immerzeel (2008) and Gautam *et al.* (2009) observed different trends across eastern Himalaya makes it difficult to decide whether runoff contribution from precipitation is increasing or stable.

4.4 Snow and ice melt contributions towards runoff

Large quantities of ice and snow melt contribute to discharge in the western Himalaya. Snow melt contributes 40% towards Indus runoff compared to 32% for glacial melt according to Immerzeel (2009). Immerzeel outlined larger contributions from ice and snow melt towards discharge in north-west, as expected with high snow coverage in the upper Indus of 33.9%. On average snow and glacier melt contributes 59% towards runoff in the Sutlej in western Himalaya (Singh and Jain, (2002). It is evident that snow and ice melt is the major contributor towards river flow in the north-west and western Himalaya, as Immerzeel (2009) and Singh and Jain (2002) identified. Percentage contributions vary due to different climatic conditions in each basin. Snow cover during winter decreases from west to east across Himalayan arc

(Figure 4.1), with a large percentage contribution towards runoff in north-west and western Himalaya.

Average snow and glacier contribution towards flow in the Chenab River at Akhnoor in western Himalaya is estimated to be 50% compared to 30% for the Ganges at Devprayag in central Himalaya (Singh et al., 1994; Singh et al., 1997). Singh (2006) estimated that glacier melt runoff from Dokriani glacier contributes 87% towards total runoff. This estimate is very high compared to others but Singh (2006) does not account for snow melt. The estimate provided by Singh (2006) may be a lot higher due to the Dokriani glacier having a large ice area, or measurements have been taken at high elevation were contribution from ice melt would be much greater than downstream stations in which rainfall accounts for a much larger percentage of flow.

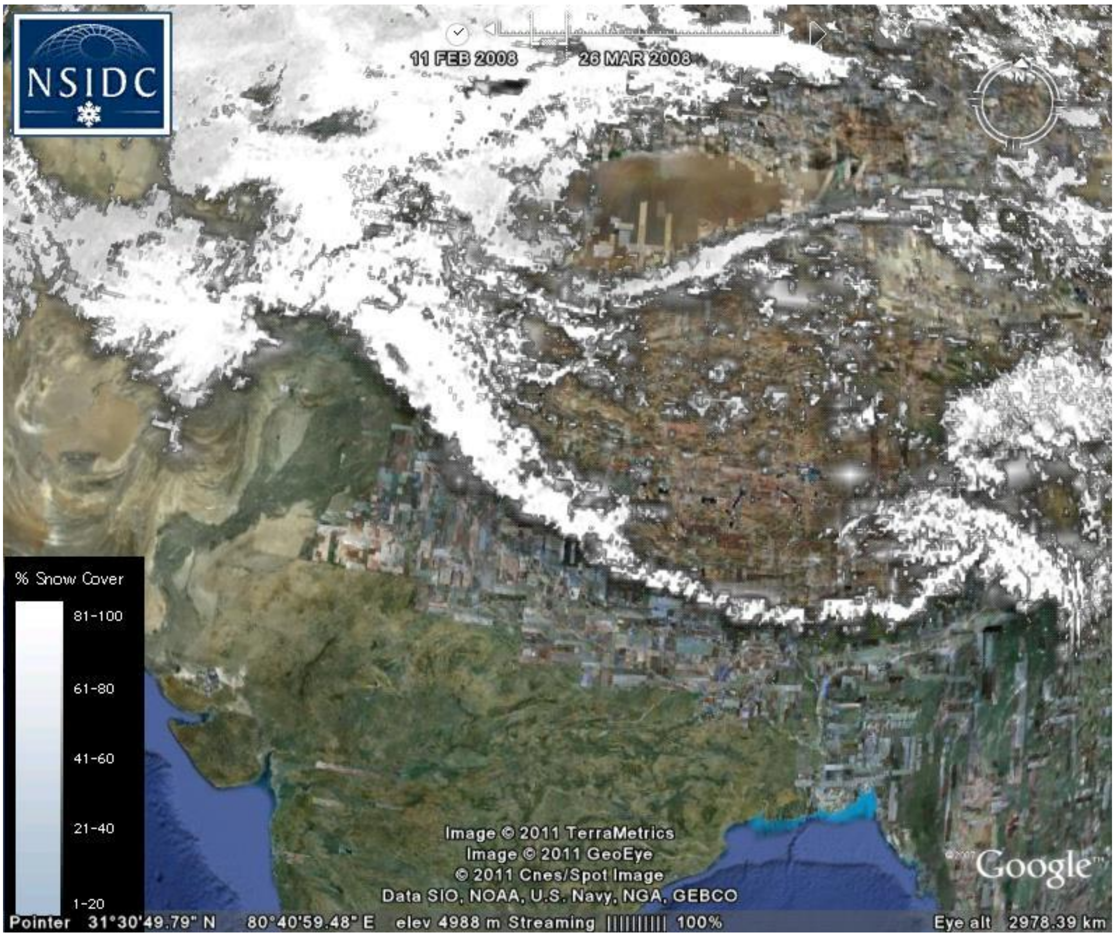

Figure 4.1 Winter snow cover across the Himalayan arc (Modis website, 2011).

4.5 Air temperature trends

Bhutiyani *et al*. (2007) observed a significant warming during last century across all stations in the north-west, i.e. Shimla, Srinagar and Leh. Shekhar *et al*. (2010) similarly identified an increase in temperatures across western Himalaya. Shekhar *et al*. (2010) noticed that seasonal mean, maximum and minimum showed an increase of 2.0, 2.8, 1.0 °C respectively. In general north-west Himalaya experienced warming that must inevitably increase snow and ice melt. Bhutiyani *et al*. (2007) used a standardised temperature index and Shekhar *et al*., (2010) looked at temperature anomalies. Those are not entirely accurate representations of temperature trends because they are not exact temperature values that would give a better indication of climate change. However, Yadav *et al*., (2004) observed cooling during pre-monsoon in western Himalayas but only for a short period of time, indicating that not all basins in western Himalaya are conforming to warming trends.

Over the last century whilst air temperature increased in the western Himalaya, warming may not have occurred across central and eastern Himalaya. Immerzeel (2008) observed an average increase of 0.6°C from 1900 to 2000 across the Brahmaputra basin. Immerzeel (2008) identified the pre monsoon period as having the greatest warming with 1.0 and 1.1°C rise in the Himalayan belt and Tibetan Plateau respectively. Jhajharia and Singh (2010) noted that during winter and pre-monsoon seasons air temperature in eastern Himalaya had no trend, contradicting Immerzeel (2008). However, Jharharia and Singh (2010) agreed with Immerzeel (2008) in observing significant warming in both monsoon and post monsoon seasons. Shrestha *et al.,* (1999) looked at 49 stations in Nepal, which show warming trends from 0.06 to 0.12°C per year. Both Jharharia and Singh (2010) and Shrestha *et al.* (1999) observed post monsoon warming suggesting that this time of year may be especially vulnerable to climate change. Across central and eastern Himalaya temperatures seemed to be warming, which would lead to a change in runoff characteristics by increasing melt during ablation season.

Tibetan Plateau receives minimal precipitation in a cold and arid climate, meaning relatively little melt being contributed towards river flow (Singh and Jain, 2002). Since the mid-1950s, Liu and Chen (2000) monitored a significant warming trend, with an annual increase of 0.16°C and 0.32°C in winter mean over a decade. However, Shi *et al*. (2006) identified a decreasing trend from late 1970s to 1990s. Li and Tang (1986) observed decreasing temperatures from 1950 to 1970, differing from Shi *et al*. (2006) who noted a declining trend from late 1970s. Li and Tang (1986) observed an increasing trend from 1970 onwards supporting Liu and Chen (2000) that identified significant warming since the mid-1950s. Studies of the Tibetan Plateau are inconclusive with trends monitored over short time periods which are therefore inconclusive for long term changes.

4.6 Glacier recession trends

As air temperature and precipitation patterns changed over last century, inevitably the rate of glacier melt has been influenced, but to varying extents. Schmidt and Nusser (2012) outlined that the glaciated area decreased by about 14% (0.3% per year) in Ladakh, north-west India. More recently, 50% of glaciers in the north-west are advancing or stable from 2000 to 2008 (Scherler *et al*., 2011). However, more than 65% of monsoon-influenced glaciers are retreating (Scherler *et al*., 2011). In north-west Himalaya glaciers have experienced less recession and some are even advancing compared to monsoon catchments. Rates of glacier recession in Himalayas differ depending on the region, elevation and meteorological trends. In Himachal Pradesh deglaciation rates are 22% in Parbati basin, 21% Chenab basin, 19% Baspa basin and 21% total from 1962 to 2001 (Kulkarni *et al*., 2005), reinforced by a 19% reduction in ice area between 1962 and 2001 (Kulkarni and Alex, 2003). In Uttarakhand, Dokriani glacier volume reduced by 20% from 1962 to 1995 (Dobhal *et al*., 2004). Dobhal *et al*., (2004); Kulkarni *et al*., (2005) and Kulkarni and Alex (2003) found similar amounts of deglaciation with around 20% in all studies across western central Himalaya. However, Gangotri glacier in Uttarakhand region has only lost 5%

of its length (Raina, 2009). Gangotri glacier has been less affected by climate change by comparison with other glacierised catchments, maybe because it is at a higher elevation at which precipitation continues to fall as snow. Air temperature is still sufficiently cold to prevent large scale ice losses. Glaciers located at around 5,000 m a.s.l. lost ~24% by comparison with 14% in those glaciers located above 5,400 m (Kulkarni and Alex, 2003). Raina (2009) and Kulkarni and Alex (2003) evidently show that glacier reduction has been slower at high elevation, meaning that glaciers starting higher will probably continue to exist while smaller glaciers at lower elevation will recede more rapidly and ultimately cease to exist.

4.7 Water towers

Himalayan mountain ranges provide large quantities of runoff to people downstream with more than 1.4 billion people depending on water from Indus, Ganges, Bhahmaputra, Yangtze and Yellow Rivers (Immerzeel *et al*., 2010). Viviroli (2003; 2004; and 2007) saw high mountain regions as water towers, with Himalayas contributing a disproportionally high runoff compared to lowlands. As the Himalayas are vital for water availability downstream it is important to understand how much comes from glacier runoff and how it will be altered with future climatic change. In the Indus 90% of lowland flow originates from mountain areas, mainly from Hindukush, Karakorum, and western Himalaya (Liniger *et al*., 1998). In western Himalaya, Winiger (2005) suggested that 70% of annual runoff in the Indus is produced by seasonal monsoon. Downstream gauging stations show greater monsoon influence in western Himalayas. Water availability downstream in the three major rivers will be a major issue in the future. Immerzeel *et al*. (2010) estimated that by 2050 the number of people with access to fresh water downstream will decrease by -34.5 ± 6.5 million in Brahmaputra, −2.4 ± 0.2 million in Ganges and −26.3 ± 3.0 million in Indus basin. Indus and Brahmaputra are likely to be severely impacted in regions and in countries with already a high dependence on irrigated agriculture and melt water as runoff from glaciers is particularly sensitive to changes in precipitation and temperature (Barnett and Lettenmaier, 2005). High mountains are often labelled as 'water towers' however; future changes in runoff characteristics are needed to be understood with specific focus on seasonal variation, water balance and reaction time of runoff in relation to climate change (Winiger, 2005).

4.8 Discharge trends

As air temperature and precipitation have changed over the 20th century, inevitably runoff levels across the Himalayan arc have greatly transformed. In north-west Himalaya Bhutiyani *et al*., (2008) identified that annual and monsoon discharge decreased in Beas River from 1961-65 and in Ravi River but insignificantly. Both river basins have relatively few glaciers in their catchments, with runoff mostly monsoon fed. Beas and Ravi rivers surprisingly have heavy influence from summer monsoon even though located in north-west Himalaya. Snow and ice are the main contributors towards river flow in the Sutlej and Chenab Rivers also in the north-west Himalaya Recently, discharge in the Sutlej River declined from 1991 to 2004 (Bhutiyani *et al*., 2008). Sutlej follows a similar decreasing trend to the monsoon-fed Beas and Ravi Rivers. However, discharge in the Chenab River increased annually and during winter, differing from the three other rivers studied by Bhutiyani *et al*., (2008). In central Himalaya, Neupane *et al*. (2010) identified a rise in discharge in pre monsoon in Narayani River from 1963-2006. Increase in runoff in this region is accredited to melting of ice in high mountains but this study does not look at the more important summer period. In Nepal, the Bagmati River showed post and pre monsoon periods to be more or less constant from 1965-2005 (Sharma and Shakya, 2006). Differing from the increasing trends observed in pre monsoon along the Narayani River. Sharma and Shakya (2006) identified a significant decrease in mean annual and monsoon river flow supporting Bhutiyani *et al*., (2008). Zongxing *et al*.,

(2009) observed that runoff from glacier areas has increased seasonally and annually in Yanggong basin (1979-2003) and Hailuogou basin (1988-2006). Across the Himalayas runoff seems to be decreasing especially annually and during summer, but more research is needed with limited data records.

4.9 Future trends of river flow and climate change

In basins across Himalaya changing meteorological conditions have led to declining flows. Water security will become a major issue as glaciers decline but as Rees and Collins (2006) pointed out, future warming is unlikely to affect river flow uniformly throughout. Rees and Collins (2006) modelled two catchments at opposite ends of Himalayan arc, west and east. Under a warming climate Rees and Collins highlighted that glacier fed rivers throughout will respond in a similar manner to climatic warming, except summer snowfall in the east that will reduce rate the of initial flow increase with timing of peak discharge delayed postponing eventual disappearance of ice. As glaciers decline, impact will be much greater in smaller, highly glacierised catchments in both east and west Himalaya. Immerzeel (2008) predicts accelerated seasonal increases in temperature and precipitation from 2000-2100 in Brahmaputra basin, with the greatest changes in more highly elevated Tibetan Plateau compared to floodplains. As meteorological variables inevitably change, river flow will increase and most significantly during summer (Immerzeel, 2008). However, runoff during autumn and spring will not see a large rise until about 2050. Future climatic change will increase river flow in Brahmaputra basin as snow and ice melt rise due to rising air temperature but will also depend on whether summer snowfall that may increase runoff contributions.

Akhtar *et al*. (2008) modelled discharge in a changing climate under different percentages of glacier coverage in three Hindu Kush river basins. Under a 100% glacier scenario discharge will generally increase 60% and 88% for HBV-PRECIS and HBV- Met models respectively. Runoff will increases under 50% glacier coverage in both HBV-PRECIS and HBV-Met models with 24% and 10% respectively. However, under 0% coverage there is a noticeable decrease in water resources with up to 94% in HBV-Met and 15% in the HBV-PRECIS model. In the future, basins with high glacier coverage will observe a huge increase in runoff compared to less glacierised basins that will experience a smaller increase or decrease in regions with very few glaciers.

4.10 Deglaciation discharge dividend

Initially runoff in glacierised catchments increases due to enhanced melting of snow and ice as climatic warming takes place (Immerzeel 2008; Jansson et al., 2003). As glaciers eventually reduce to an insignificant mass and ultimately cease existence altogether, ice free areas increase which alters the discharge patterns in what where highly glacierised basins. As glacier recession occurs, a percentage of river flow in excess of that related to existing precipitation, a deglaciation discharge dividend, is added to basin runoff from reduction of water stored as ice (Collins, 2008). Collins (2008) highlighted that river flow will solely reflect future levels of precipitation as the dividend declines and glaciers cease to exist. In catchments with high percentage glacierization, discharge variation will reflect temporal variations in heat energy available for melt (Collins, 1989). Therefore in lower glacier coverage basins, runoff will be greatly influenced by precipitation with higher year-to-year variations (Collins, 2008). Kaser *et al*., (2010) outlined that a period of strong melt in response to on-going warming will lead to a temporary rise in average precipitation and possibly maximum monthly precipitation. While glacier extent decreased at same time, this effect of "deglaciation discharge dividend" would be compensated (Kaser et al., 2010). Kaser *et al*., (2010) found that generally low maximum monthly precipitation with an assumed doubling of glacier melt water production during strong glacier volume loss on large river basin would have little impact. Future Himalayan basins will be precipitation dominated as glacier mass is insufficient to provide large quantities of flow downstream as the deglaciation discharge dividend ultimately declines.

4.11 Evaluation of journal limitations

Data collection in journals researching into climate across Himalaya has to be assessed on its validity and accuracy. Studies often fail to use real data when investigating air temperature and precipitation trends. Bhutiyani *et al*. (2007, 2008 and 2010) used standardised indices for temperature discharge and precipitation across north-west Himalaya. Rather than using real data the use of standardised index is used to reflect a desired outcome, such as rising temperatures (Bhutiyani *et al.* 2007). Shekhar *et al.* (2010) used minimum, maximum and mean temperature anomalies to identify trends. Anomalies only give an idea of whether rare occurrences have accelerated instead of complete data that accounts for every day. Most accurate representation of precipitation trends would be annual rainfall data. However, Shekhar *et al*. (2010) used cloud cover and number of snowfall days that do not give total precipitation amounts. Journal papers concentrate on particular regions of the Himalaya rather than the range as a whole. Bhutiyani *et al.,* (2007, 2008 and 2010) researched into north-west Himalaya, while Immerzeel (2008) looked at the Brahmaputra basin and Shrestha *et al*. (1999) investigated Nepal. Each study reflects three different catchments, but a combination of these studies would identify climatic change across Himalayan arc. Some research projects looking at Himalayan climate change use very short data records to identify trends over 20th century into the 21st century. Shekhar et al., (2010) and Shrestha et al., (1999) only used data sets 30 to 40 years long, which is insufficiently long accurately to reflect climate change. Research into discharge is limited with very few studies and poor coverage of river gauges across different river catchments. Bhutiyani et al., (2008) and Neupane *et al*., (2010) are good studies but lack long data series for comparison with relevant precipitation and air temperature trends. Ideally a more extensive river gauging network across Himalayas would give a better indication of river flow.

5 Methods

Across the Himalayas, stations were selected with particular criteria for each meteorological variable. Collins (1989) chose stations in close proximity to glaciers, at high elevation and a degree of correlation between hydro meteorological variables derived from records with mass balance or runoff. Ideally, this method of approach would be adopted when selecting stations across the Himalaya. The Collins (1989) approach is preferred over using standard annual meteorological seasons as used for example by (Dimri, 2009). Dimri (2009) defined short seasons of summer from June to August and December through to February for winter.

Long uninterrupted data series should be favoured over shorter sets even if the latter are situated at high elevation and in close proximity to glaciers. Longer data series are favoured, as they give a better indication of long term change since 1850, by comparison with short data sets that are limited in reflecting the changes from the late 19th century through to the early 21st century. Climate normal is defined over a 30-year period that is long enough to filter out any inter-annual variation or anomalies, but still enough to show long term trends (World Meteorological Organisation, 2014). A long data set will be more than 30 years, but preferably around 100 years in length. Sites must have long homogenous data series for air temperature, precipitation and discharge and form a set distributed across different river basins of Himalaya. A homogeneous series is one that is collected in a uniform manner throughout so that changes observed are changes that actually occur and are a result of a change in the measurement technique. In each region there should be at least one station at high elevation with a long continuous data set. Chosen sites should cover the three main river catchments in the Himalaya - Indus, Ganges and Brahmaputra. Chosen stations are situated therefore within the study area covering the foothills of Himalaya through to high elevations and the Tibetan Plateau between 38^0 N-28^0 N and 70^0 E - 95^0 E, the area defined by Collins and Hasnain (1995).

5.1 Air temperature stations

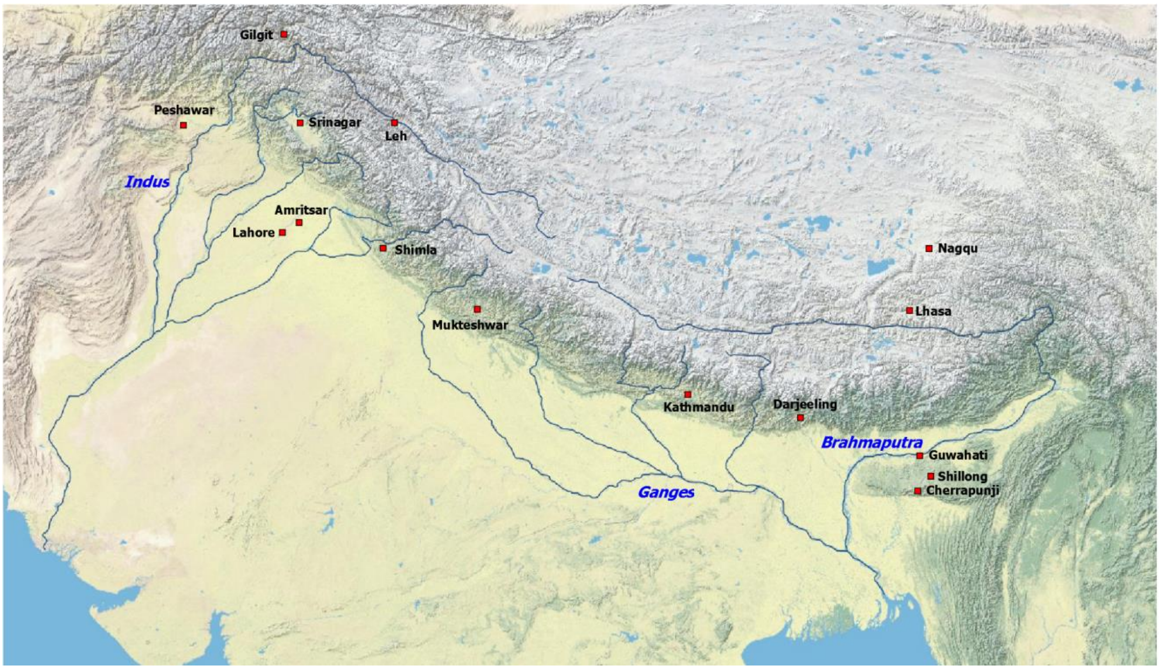

Figure 5.1 Air temperature stations across the Himalayan arc (Map adapted from ArcGIS Explorer).

Air temperature stations cover a large expanse of Himalaya with 15 sites in total as shown in Figure 4.1. Ideally sites should be situated at high altitude because a range in different elevations would help identify whether warming is accelerated in high mountains. Many stations for which data exist did not fit the selected criteria. More than 70 stations were not used because data series do not continue through the 20th century into the 21st century or contain too many missing years of record. The majority of stations were rejected because they were located at low elevations, with only five of more than 250 meteorological stations in Nepal being located above 4000 m a.s.l. (Chalise, 1996). In north-west Himalaya, Peshawar, Gilgit, Srinagar and Lahore were selected. In the western Himalaya Amritsar, Leh and Shimla were chosen, whilst Mukteshwar and Kathmandu were the only available stations meeting the criteria in the central Himalaya. Lhasa and Nagqu in Tibetan Plateau are the highest stations at 3656 and 4513 metres respectively (Table 5.1). In east Himalaya Darjeeling, Cherrapunji and Shillong are at higher altitudes than lowly Guwahati at 51 m a.s.l. (Table 5.1). Collins (1989) measured summer air temperature from 1 May to

30 September ($T_{5\text{-}9}$) at Sion. Average summer air temperature data collected for this study was also taken from May to September and annual air temperature from 1 January to 31 December. Both of these measurements will be a weighted average, that takes into account the number of days in a month so that each month will be weighted individually rather than collectively which makes the average more accurate. Air temperature data were taken from British Meteorological Office, Royal Dutch Meteorological Institute (KNMI), Pakistan Meteorological Department and NASA (Table 5.1). Data for some stations data can be found at more than one source; in this case the longest and most complete record was selected. Data had to be sorted by removing -999.9 and sequentially removing those years that cannot be used (because individual monthly elements are missing) which can short or long gaps in records.

5.2 Precipitation stations

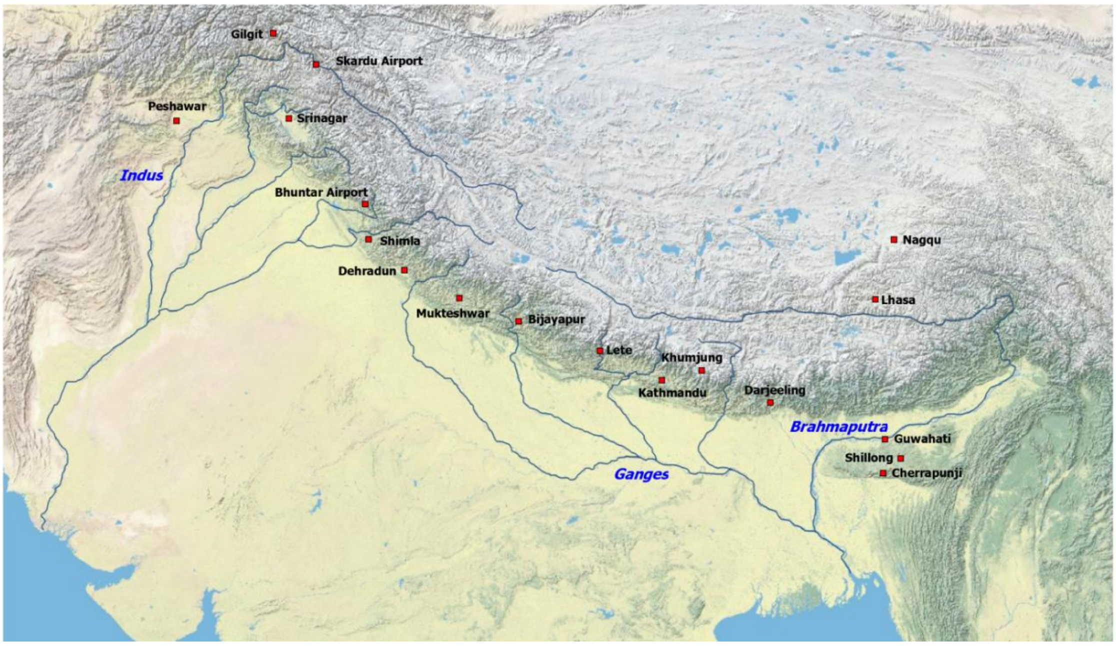

Figure 5.2 Precipitation stations across the Himalayan arc (Map adapted from ArcGIS Explorer).

In total 18 stations were selected to characterise precipitation across the Himalaya, covering from Karakoram in the far west to monsoon-dominated eastern Himalaya (Table 5.2 and Figure 5.2). A large number of stations was rejected with around 150 stations not being used because the data set was either not long enough or had too many values missing. The majority of the data provided by EarthInfo ended abruptly around 1970 meaning that many stations were rejected because they did not run towards or into the 21st century. Several stations have long records for both air temperature and precipitation, but for some locations data are not available at all or exist only for a short period. In this case, another suitable location would be chosen that would be long and uninterrupted, providing a better indication of trends.

In north-west Himalaya Peshawar provides the best available long record but it is at only 359 m a.s.l. so Skardu Airport, Gilgit and Srinagar at higher elevations were chosen (Table 5.2). In western Himalaya, Shimla and Bhuntar Airport only run up to 1980 yet have to be correlated with Skardu Airport and Srinagar. Dehradun in central Himalaya only runs until 1970 and compares with Mukteshwar that continues until 2008. Other central Himalaya stations in Nepal have much shorter records with Bijayapur, Lete, and Khumjung only covering 1970-1990 while Kathmandu is from 1951 to 2008 (Table 5.2). Lhasa and Nagqu on the Tibetan plateau are the highest stations in the Himalaya, and allow examination of the question whether precipitation changes have been accelerated at altitude. Stations in the monsoon driven eastern Himalaya are Darjeeling, Cherrapunji, Guwahati and Shillong with 3 of the stations at high elevations.

Variables selected were total winter (1 November to 31 May) and total summer precipitation (1 June to 30 September). There are a few different sources of data collection for precipitation records that are EarthInfo, Royal Dutch Meteorological Institute (KMNI) and Pakistan Meteorological Department (Table 5.2). Earth Info data only runs until 1970 at Indian stations so other sources were needed to provide longer data sets that run until more recently, such as KNMI.

5.3 Gauging stations

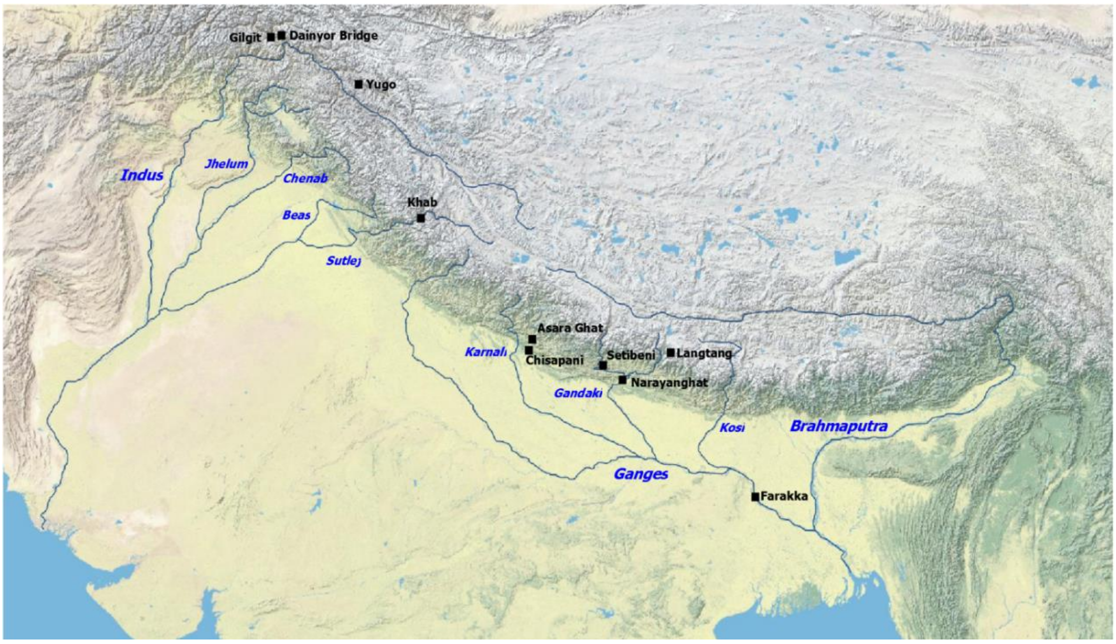

Figure 5.3 Map of three main Himalayan Rivers and tributaries with gauging stations (Map adapted from ArcGIS Explorer).

With a limited number of gauging stations across the Himalayas it is important to select those with long records in as many catchments as possible. There are four gauging stations in the Indus and Sutlej basins and six on rivers that drain into the Ganges (Table 5.3 and Figure 5.3). Farakka on the Ganges River was used only for annual hydrographs because it is far downstream and water is abstracted upstream for irrigation, so it is unlikely therefore to reflect runoff from mountains or glaciers. Data collected from gauging stations are taken from partially-glacierised headwater basins that feed into the Indus and Ganges. However, over 15 gauging stations were not selected as they did not meet the criteria. They were rejected due to insufficient data series or too many years that would lead to huge gaps in the plotted graphs. Sites in north-west and western Himalaya are within close proximity to glaciers with elevations ranging from 1432 to 2550 metres. Stations at higher altitude will more accurately reflect snow and ice melt fluctuations. Gauging stations across Nepal are mainly located in lower foothills not high mountains. At stations at great distances away from high mountains, runoff levels will not truly reflect glacial melt as precipitation influence increases downstream. As river flow is dominated by summer monsoon in central Himalaya, gauging stations situated at high elevation are not as important with less percentage contributed from snow and ice melt. Langtang station is located close to glaciers but data records are shorter than other central Himalaya sites. Ideally discharge data sets would be as long as those of air temperature and precipitation. However, length of discharge series only extends from 1960 to 2006 (Table 5.3). Generally records are approximately 30 years in length, with the longest set 43 years at Narayanghat and shortest only 10 years at Langtang. Most data sets ends in the 1990s but correlation was used to compare overlapping sections of records in similar partially-glacierised basins. Discharge is measured on annual basis between 1 January and 31 December and during summer (1 June to 30 September), to compare with glaciologically-relevant climatic parameters. With a wide coverage of gauging networks there are number sources; Global Runoff Data Centre (GRDC), Indian Hydroelectric (IHE) Project, Nepal Department of Hydrology & Meteorology (DHM) and Water and Power Development Authority (WAPDA) (Table 5.3).

5.4 Coefficient correlation

Correlation analysis was undertaken to provide an indication of whether temperature, precipitation and discharge trends are related, and whether behaviour of variables at the various stations occurs in parallel in the regions of the Himalaya. Correlation was also used to identify connexions between variables across the range of the diverse and complex regions of the Himalaya by comparing behaviour of variables within and between the river catchments. Correlation was also used to access climate-runoff relationships, using overlapping parts of the data series.

Table 5.1 List of air temperature stations across the Himalaya.

Station	Country	Elevation (metres)	Length of series	Longitude	Latitude	Source
Peshawar	Pakistan	359	1931-2010	71°30'51.09"E	33°59'53.04"N	UK Met Office
Gilgit	Pakistan	1465	1968-1985	74°19'51.10"E	35°55'03.35"N	Pakistan Met Department
Srinagar	India	1590	1893-2010	74°48'30.57"E	34° 4'58.01"N	UK Met Office
Lahore	Pakistan	216	1876-2009	74°20'33.18"E	31°32'21.48"N	UK Met Office
Amritsar	India	232	1949-2008	74°52'20.29"E	31°38'02.21"N	UK Met Office
Leh	India	3411	1882-1968	77°34’38.77”E	34°09’07.82”N	KNMI
Shimla	India	2143	1881-1986	77°10'47.20"E	31°05'19.21"N	NASA
Mukteshwar	India	2289	1898-2008	79°38'50.65"E	29°28'26.76"N	UK Met Office
Kathmandu	Nepal	1302	1951-2007	85°19'5.68"E	27°42'10.34"N	UK Met Office
Lhasa	China	3656	1935-2007	91°8'27.08"E	29°38'43.99"N	UK Met Office
Nagqu	China	4513	1955-1995	92°3'32.05"E	31°28'35.39"N	UK Met Office
Darjeeling	India	2128	1884-1981	88°15'45.26"E	27°02'08.89"N	NASA
Cherrapunji	India	1489	1903-2010	91°42’00.04”E	25°17’59.82”N	UK Met Office & NASA
Guwahati	India	51	1903-2009	91°44'10.10"E	26°08'49.25"N	UK Met Office
Shillong	India	1439	1903-2000	91°53’35.90”E	25°53’35.90”N	NASA

Table 5.2 List of precipitation stations across the Himalaya.

Station	Country	Elevation (metres)	Length Of Series	Longitude	Latitude	Source
Peshawar	Pakistan	359	1862-2000	71°30'51.09"E	33°59'53.04"N	Pakistan Met Department
Skardu Airport	Pakistan	2222	1900-1999	75°32’30.54"E	35°20’20.16"N	KNMI
Gilgit	Pakistan	1465	1901-1999	74°19'51.10"E	35°55'03.35"N	Pakistan Met Department
Srinagar	India	1590	1893-2010	74°48'30.57"E	34° 4'58.01"N	KNMI
Bhuntar Airport	India	1092	1901-1965	77°09'13.42"E	31°52'38.80"N	Earth Info
Shimla	India	2143	1863-1988	77°10'47.20"E	31°05'19.21"N	KNMI
Lhasa	China	3656	1936-2004	91°8'27.08"E	29°38'43.99"N	KNMI
Nagqu	China	4513	1957-1997	92°3'32.05"E	31°28'35.39"N	Earth Info
Dehradun	India	653	1901-1970	78° 1'54.94"E	30°19'17.16"N	Earth Info
Mukteshwar	India	2289	1897-2009	79°38'50.65"E	29°28'26.76"N	KNMI
Bijayapur	Nepal	1857	1971-1990	80°36’57.96"E	29°33’58.42"N	KNMI
Lete	Nepal	2208	1971-1990	83°38’24.85"E	28°36’25.40"N	KNMI
Kathmandu	Nepal	1302	1951-2008	85°21'31.16"E	27°41'31.16"N	Earth Info and KNMI
Khumjung	Nepal	3806	1971-1990	86°43’00.07"E	27°49’00.47"N	KNMI
Darjeeling	India	2128	1901-1970	88°15'45.26"E	27°02'08.89"N	Earth Info
Cherrapunji	India	1489	1873-2004	91°42’00.04"E	25°17’59.82"N	KNMI
Guwahati	India	51	1865-2010	91°44'10.10"E	26°08'49.25"N	KNMI
Shillong	India	1439	1867-2000	91°53’35.90”E	25°53’35.90”N	KNMI

Table 5.3 List of gauging stations with basin characteristics and available records in the Indus and Ganges River catchments.

Gauging Station	River	Country	Elevation of gauge (metres)	Basin area (km^2)	Percentage glacierisation (%)	Length of series	Longitude	Latitude	Source
Gilgit	Gilgit	Pakistan	1457	12800	27	1960-1999	74°18’39.21"E	35°55’30.33"N	WAPDA
Dainyor Bridge	Hunza	Pakistan	1432	13733	30	1967-2004	74°22'18.77"E	35°55'29.71"N	WAPDA
Yugo	Shyok	Pakistan	2434	19540	-	1973-1997	76°10’18.06"E	35°11’07.08"N	WAPDA
Khab	Sutlej	India	2550	44135	-	1972-2002	78°38’25.06"E	31°48’04.71"N	IHE Project
Chisapani	Karnali	Nepal	191	42890	4.97	1962-1993	81°16'56.35"E	28°38'30.24"N	GRDC
Asara Ghat	Karnali	Nepal	629	19240	-	1962-1993	81°26’41.03"E	28°57’03.82"N	GRDC
Setibeni	Kali Gandaki	Nepal	410	6630	7.83	1964-1993	83°34'48.32"E	27°58'42.91"N	GRDC
Narayanghat	Narayani	Nepal	189	31100	6.94	1963-2006	84°25'33.33"E	27°42'04.29"N	Nepal DHM
Langtang	Langtang	Nepal	3688	354.75	29.51	1988-1998	85°33’50.57"E	$28^0$12’26.64"N	GRDC
Farakka	Ganges	India	21	835000	-	1949-1973	87°55'31.34"E	24°47'59.73"N	GRDC

6. Results

6.1 Annual hydrological regimes

6.1.1 North-west Himalaya- Hunza River hydrological regime

The Hunza River is an example of a typical north-west Himalaya hydrological regime. On average discharge in the Hunza River at Dainyor Bridge peaked in July above 3,000 (x 10^6 m^3) with August only slightly lower (Graph A, Figure 6.1.1). Average runoff gradually rises from April onwards as snow and ice melt increase, as temperatures gradually warm from April to August. The Hunza hydrograph is similar that of Alpine basins as they both generally peak around July or August. In the year of maximum discharge in 1970, the river flow in Hunza River at Dainyor Bridge increased more sharply than on average peaking during August (Graph B; Figure 6.1.1). However, during the minimum flow year of 1989 discharge peaked earlier, in July. During years of average and minimum flow, runoff peaks earlier than in the maximum flow year suggesting that runoff will continue to rise for longer in years of higher runoff.

At Skardu Airport, most of the precipitation occurs during winter from December to May, with an average of 25 mm per month, with a maximum in March (Graph B, Figure 6.1.1). Skardu Airport receives a large amount of winter precipitation due to the winter westerlies, but there is a small influence from summer monsoon as summer precipitation increases from July to September (Graph B, Figure 6.1.1). It would be thought that monsoon does not influence far north-west Himalaya, as summer precipitation strength decreases from east to west. Air temperature at Srinagar peaks in July and August, which is the same time as Hunza average discharge reaches its maximum (Graph C, Figure 6.1.1). On average, temperatures at Srinagar are very warm and reach a maximum of 24°C, but from December to February temperatures are much lower around 1-3°C. As temperatures start to rise during spring through summer, river flow in the Hunza at Dainyor Bridge follows the increasing trend (Graph C, Figure 6.1.1). However, the discharge curve lags behind air temperature, rising in May and March respectively.

6.1.2 Western Himalaya- Sutlej River hydrological regime

Sutlej River at Khab is situated to east of the Hunza River at Dainyor Bridge. On average discharge in the Sutlej at Khab peaks in July at around 1700 (x 10^6 m^3). (Graph A, Figure 6.1.2). River flow starts to rise from April onwards as snow and ice melt increases as temperature warms, but also summer precipitation adds to the runoff contributions with the highest discharges between June to August. As available heat energy to melt snow and ice declines and monsoon weakens, meaning that discharge is greatly reduced and runoff is much lower in September. In 1973, the year of the highest discharge in Sutlej River at Khab, river flows peak earlier in June.

Through June to September Shimla receives its majority of precipitation (Graph B, Figure 6.1.2), indicating that summer monsoon provides a greater percentage of runoff contribution in the Sutlej River basin. At Shimla a majority of precipitation falls during July and August with over 400 mm per month, compared to less than 100 mm per month from October to May (Graph B, Figure 6.1.2). As precipitation increases from June to August, runoff rises from April to July with a majority of runoff contributed during summer (Graph B, Figure 6.1.2). As precipitation reduces from August, discharge also declines from August until the next spring when it starts to rise again in April (Graph B, Figure 6.1.2). Air temperature at Amritsar gradually increases from January onwards until reaching its maximum in June (Graph C, Figure 6.1.2). Warming temperatures in Amritsar are followed by an increase in runoff in the Sutlej River at Khab, with a similar rising curve until June. However, in July temperatures dipped while discharge continued to rise (Graph C, Figure 6.1.2). As river flow reached its maximum in July, it would be expected that runoff would follow a decline in temperatures, but discharge remained high as summer precipitation continued to contribute.

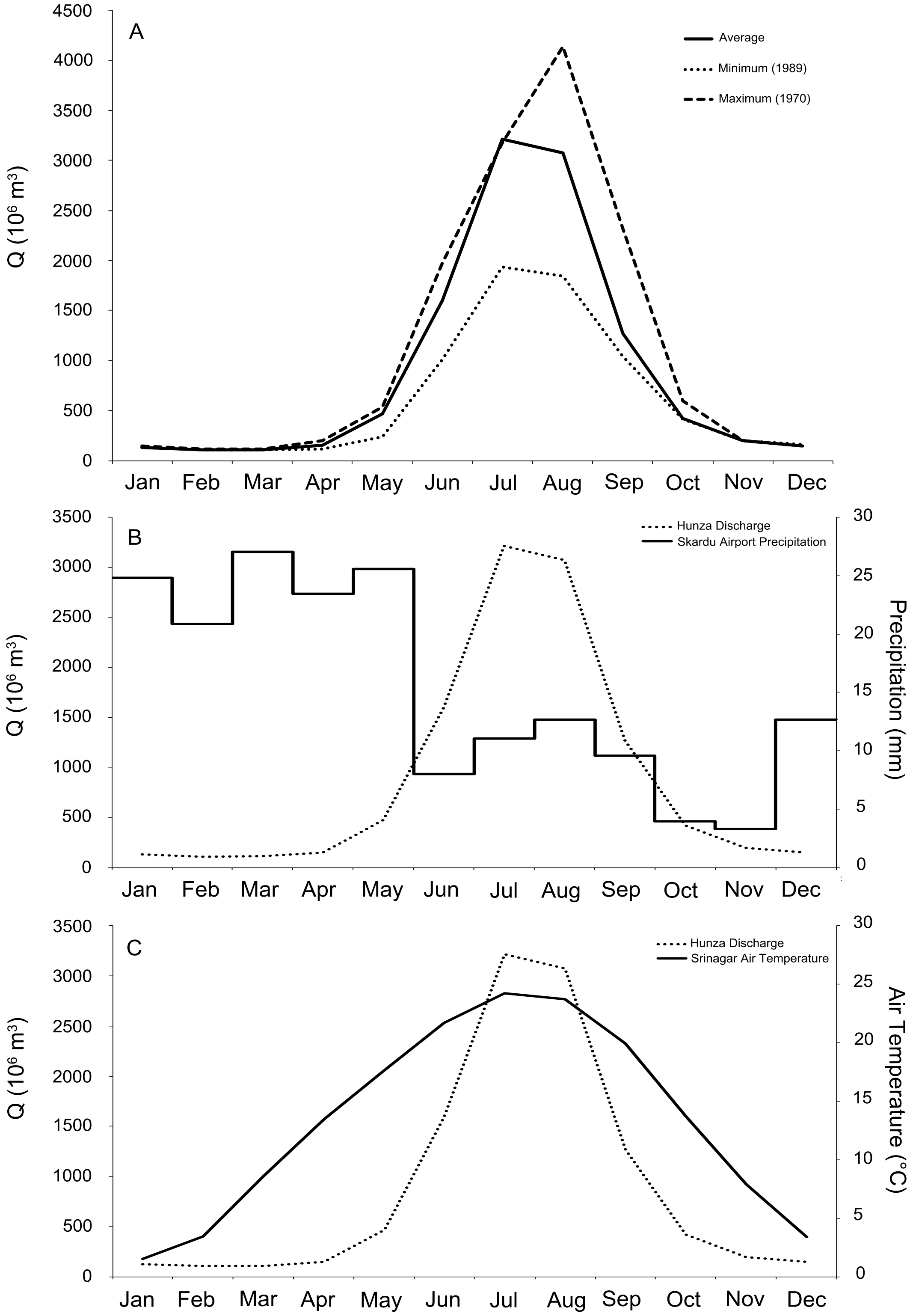

Figure 6.1.1 A) Annual hydrological regime for Hunza River at Dainyor Bridge (1967-1998), B) Skardu Airport precipitation (1900-1999) compared against annual discharge and C) Srinagar air temperature (1893-2008) against annual river flow. Elevations: Dainyor Bridge- 1457m a.s.l., Skardu Airport- 2222m a.s.l. and Srinagar- 1590m. a.s.l.

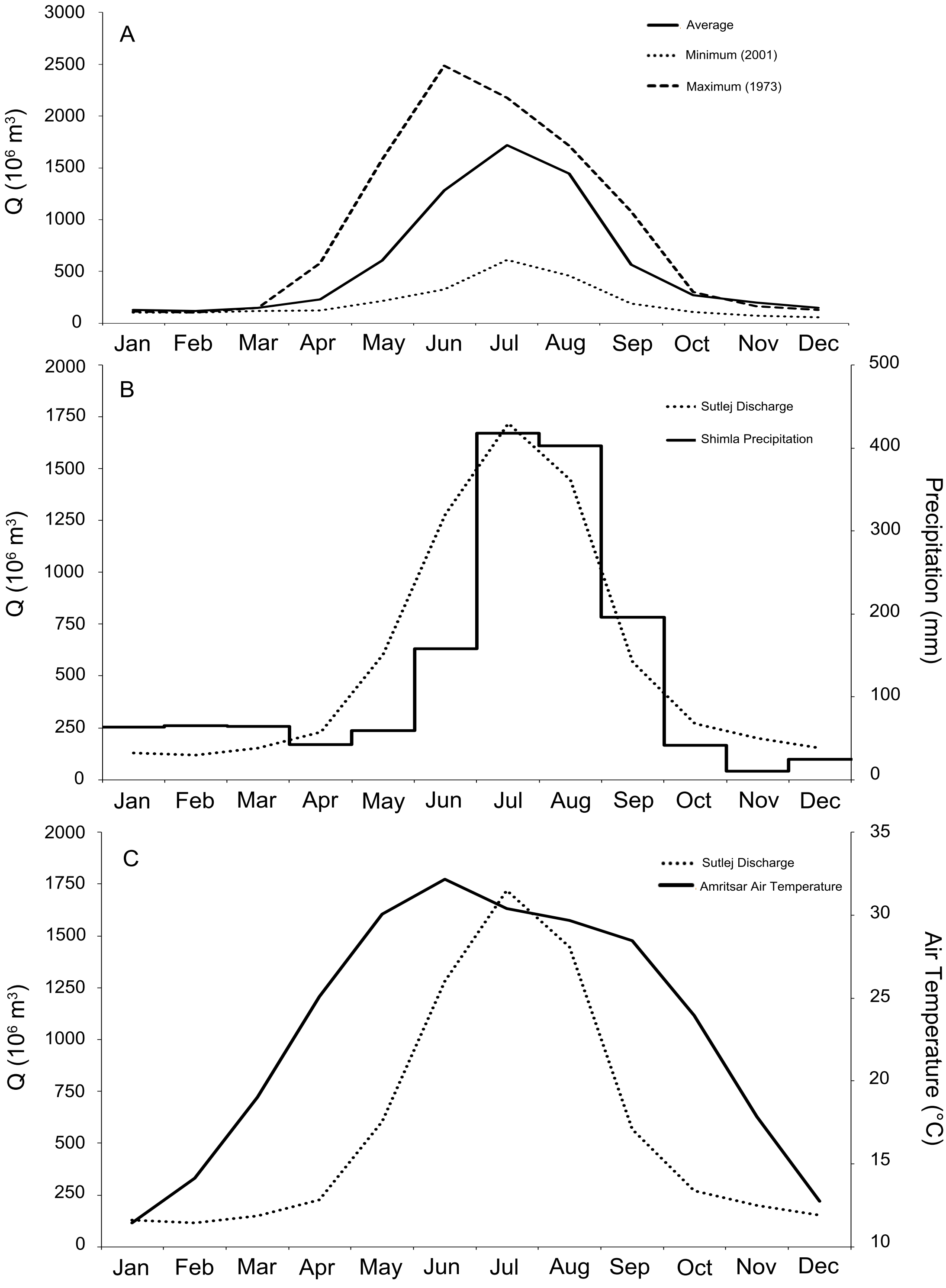

Figure 6.1.2 A) Annual hydrological regime for Sutlej River at Khab (1972-2002), B) Shimla precipitation (1901-1975) compared against annual discharge and C) Amritsar air temperature (1948-2008) against river flow. Elevations (metres): Khab- 2550m, Shimla- 2143m and Amritsar- 232m.

6.1.3 Central Himalaya- Narayani River hydrological regime

On average river runoff in the Narayani River at Narayanghat increased from May onwards, and then reaching its maximum in August of 13,341 (x 10^6 m^3) (Graph A, Figure 6.1.3). During July and August, Narayani river flow is at its seasonal highest. During year of minimum flow (2006), discharge peaked earlier in July compared to average and maximum years, when it reached its highest levels in August (Graph A, Figure 6.1.3).

Precipitation in Kathmandu from November to March is very low with less than 20mm a month (Graph B, Figure 6.1.3). Most precipitation falls between May and September, with the highest levels in July and August with over 300 mm each month (Graph B, Figure 6.1.3). As precipitation increases, discharge similarly increases in the Narayani River at Narayanghat from May to July. However, runoff continues to rise into August while precipitation experiences a small dip. Air temperatures at Kathmandu increased from January onwards until reaching maxima in August (Graph C, Figure 6.1.3). Temperatures from November to March ranged between 10 and 16°C (Graph C, Figure 6.1.3). Temperatures during July and August are at their highest around 24°C. Snow and ice melt rise from May to August as temperatures gradually increase, with both temperature and average discharge peaking in August. However, a greater percentage of runoff is contributed from summer precipitation. Even though precipitation drops during August; discharge continues to rise with still much precipitation to maintain high flows, along with increasing snow and ice melt which add to runoff.

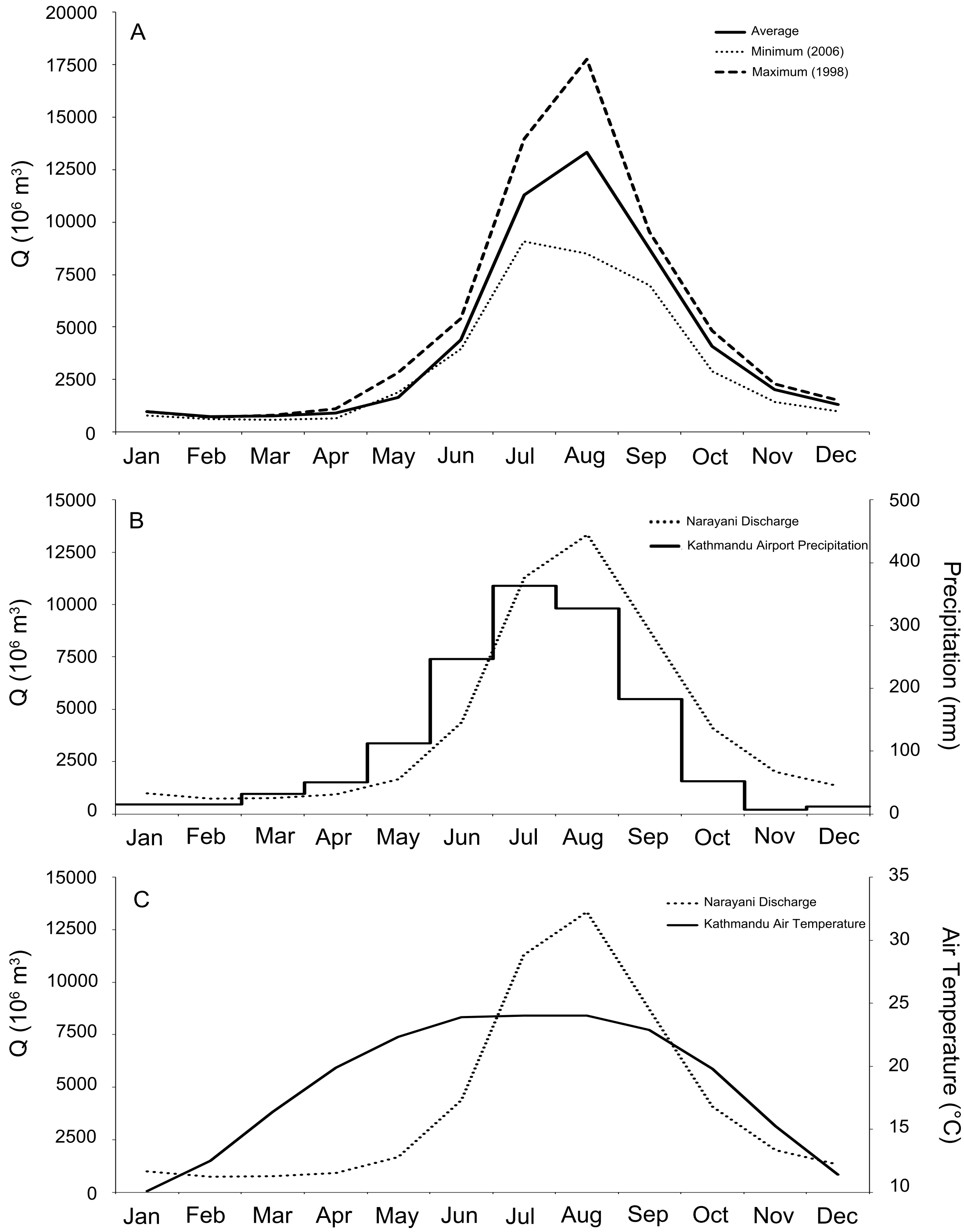

Figure 6.1.3 A) Annual hydrological regime for Narayani River at Narayanghat (1963-2006), B) Kathmandu precipitation (1950-2000) compared with annual discharge and C) Kathmandu air temperature (1951-2009) with river flow. Elevations (metres): Narayanghat 189m and Kathmandu 1302m a.s.l.

6.1.4 Eastern Himalaya- Ganges River hydrological regime

River flow in the Ganges at Farraka is very low from November to June with river flow only 4, 538 up to 25,282 (x 10^6 m^3) (Graph A, Figure 6.1.4). At Farakka, on average discharge from July experiences a sharp incline and peaks in August at 115,251 (x 10^6 m^3) (Graph A, Figure 6.1.4), and during its maximum year of 1955, its reaches 164,354 (x 10^6 m^3). However, during the minimum discharge year (1972), runoff peaks a month later and showed less of sharp incline compared to average and maximum flows. Ganges River at Farakka experiences a much shorter period of high runoff with discharge only considerably higher for four months, compared to other river basins that experience high runoff season that last five to six months.

At Shillong the majority of precipitation falls from June to September (Graph B, Figure 6.1.4). Highest precipitation totals are in June and July with the highest June amount of 455 mm. Precipitation peaks in June, but runoff only starts to rise significantly from June, as discharge lags behind precipitation with the highest discharge level observed much later in September (Graph B, Figure 6.1.4). Shillong experiences fairly mild winter temperatures between 10°C and 12°C from December to February. Temperatures slowly rise from March until peaking in July (Graph C, Figure 6.1.4). Air temperatures at Shillong are constantly high during summer between June and September with averages of over 20°C. As air temperatures warm, inevitably more snow and ice melt increase but in monsoon driven catchments summer precipitation contributes far more to runoff in the Ganges.

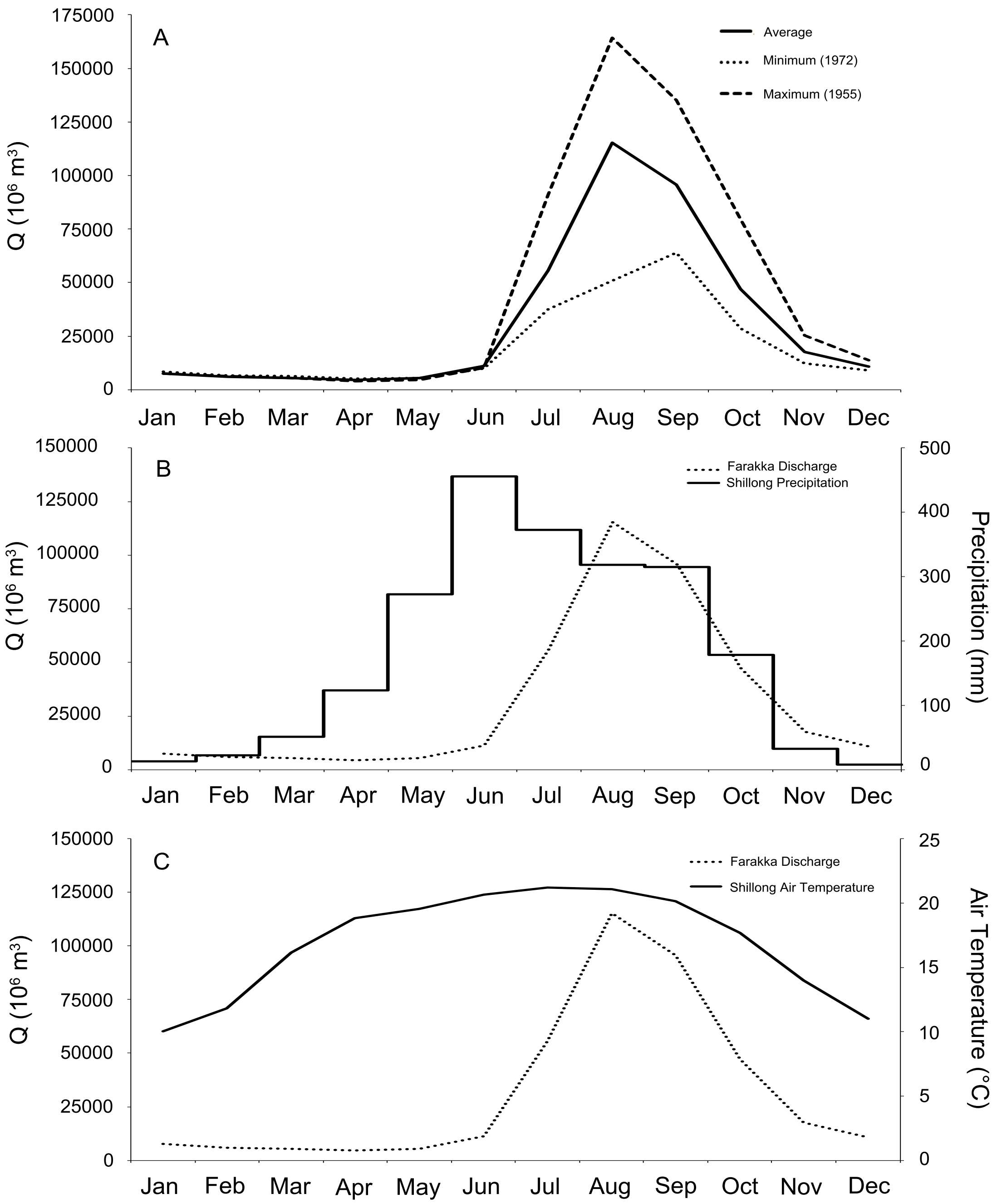

Figure 6.1.4 A) Annual hydrological regime for Ganges River at Farakka (1949-1973), B) Shillong precipitation (1950-2000) compared against annual discharge and C) Shillong air temperature(1951-2000) against river flow. Elevations (metres): Farakka 21m and Shillong 1439m a.s.l.

6.2 Long term summer air temperature trends

Summer temperature trends across the north-west Himalaya give an indication of how much heat energy is available to melt snow and ice. At low elevation, summer temperatures at Peshawar declined from the late 1930s to 1980, with a fall of 1.5°C between 1941 and 1981 (Figure 6.2.1). Since the 1980s, summer temperatures in Peshawar increased to levels that were experienced before the 1940s with a 5-year moving average of 32.04°C centred on 1999. At a higher elevation, summer temperatures at Gilgit fluctuated up and down without a major change from 1968-1985 (Figure 6.2.1). During same period, summer temperatures at Srinagar and Lahore were steady (Figure 6.2.1). Summer temperatures at Srinagar warmed from 1893-2010 (Figure 6.2.1). Initially temperatures warmed between 1900s and 1950s from 19.89°C in 1909 to 23.17°C in 1948. Most recently, summer temperatures at Srinagar increased from 1989 after a cooling period which split the two warming trends. Summer temperatures at Lahore increased from 1886 to 1949 (Figure 6.2.1). The warming period at Lahore between late 1890 and 1950 was similarly followed at Srinagar. However, from 1950s to 1985 summer temperatures at Lahore declined by about 1°C lower than the previous 55 years. Temperatures during summer warmed between 1985 and 2009 at Lahore, but not above the levels observed before 1950. Warming summer temperatures were observed from late the 19th century until the 1950s at Lahore and Srinagar, but cooled from 1950 to 1985 at all four stations. Most recently, summer temperatures increased at Peshawar, Srinagar and Lahore but not above the earlier maxima of 1950.

Sutlej river catchment in western Himalaya has three stations recording summer temperature to highlight climatic trends. At the highly elevated Leh, summer temperatures cooled from the early 1880s to 1960 (Figure 6.2.2), with a loss of just above 1°C. Leh differs from Srinagar and Lahore in Figure 6.2.1 that warmed since the late 19th century to 1950. Summer temperatures in Shimla declined from 1880s to 1920s, and during this Leh also cooled (Figure 6.2.2). Cool period at Shimla is much shorter than Leh because it continued after 1920 until 1960. Summer temperatures at Shimla experienced two warming periods, firstly from 1920 to 1950 and most recently between 1970 and 1986 (Figure 6.2.2). At lower elevation, temperatures during summer at Amritsar declined between 1949 and 1990 (Figure 6.2.2), as temperatures dropped by about 0.9°C. However, from 1990 summer temperatures warmed slightly, but not reaching the high temperatures of the 1950s. Summer temperature across the western Himalaya cooled in the century until 1980, when temperatures started to warm at Shimla and Amritsar. With limited data records in the western Himalaya, it is difficult to identify long term trends as Amritsar covers the later part of the 20th century, but Leh and Shimla only have records up 1960 and 1986 respectively.

During summer, temperatures at Mukteshwar cooled between 1900 and 1920 and again from 1960 to 1990 (Figure 6.2.3). Summer temperatures at Mukteshwar warmed from 1920s to 1960s and secondly from 1990s to 2008 (Figure 6.2.3). At Mukteshwar It is hard to identify a clear trend as summer temperature experiences cooling and warming fluctuations. Since 1920 summer temperatures have risen and most recently temperatures are at their highest and above the earlier maxima of 1960. Summer temperatures at Kathmandu decreased by about 1.2°C from 1950s to the mid-1970s (Figure 6.2.3). Temperatures at both Kathmandu and Mukteshwar declined during the same period between 1960 to mid-1970s. Summer temperatures at Kathmandu start to warm from 1975 to 2007, with an approximate increase of 1.6°C (Figure 6.2.3). Temperatures in Kathmandu most recently were warmer than the levels observed in the early 1950s. Warming in Kathmandu starts earlier than Mukteshwar, as it does not rise until 1990s. In the central Himalaya, temperatures experienced cooling and warming trends, with summer temperatures most recently rising above earlier maxima.

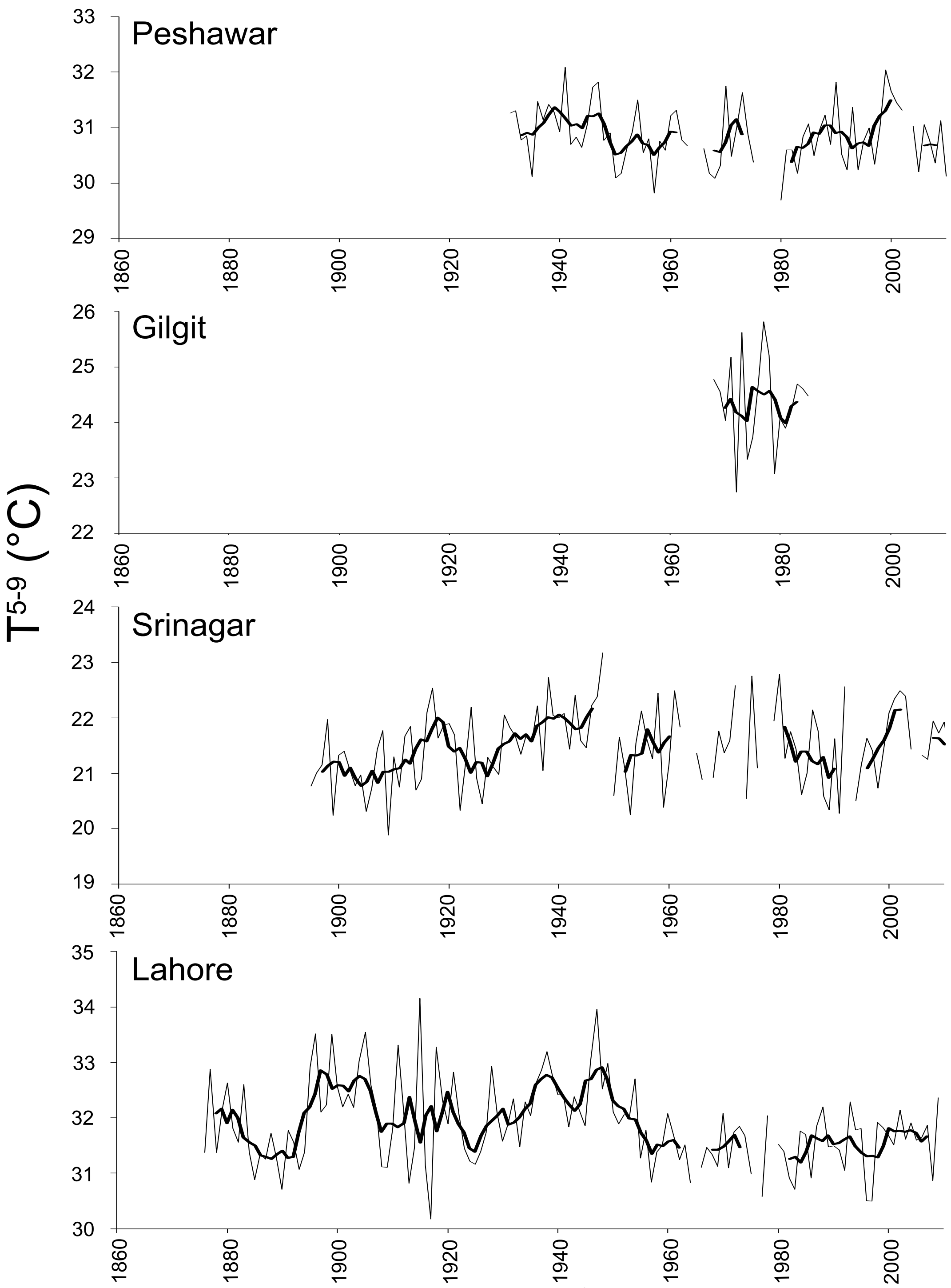

Figure 6.2.1 Summer average air temperature (1st May to 30th September) for Peshawar (359m), Gilgit (1465m), Srinagar (1590m) and Lahore (216m a.s.l.) (Thin line- real data, Thick line- 5 year moving average).

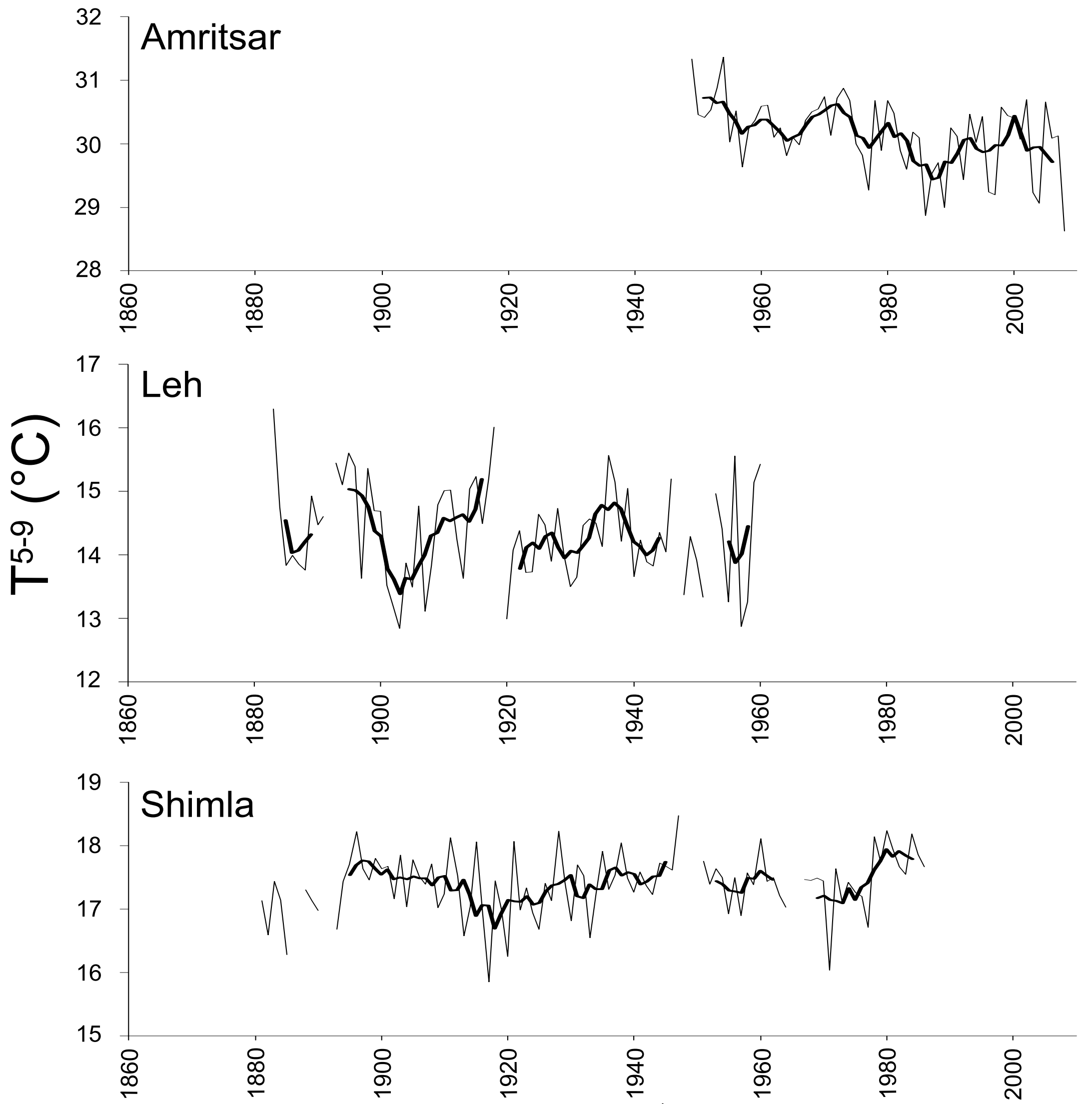

Figure 6.2.2 Summer average air temperature (1st May to 30th September) for Amritsar (232m), Leh (3411m) and Shimla (2143ma.s.l.) (Thin line- real data, Thick line- 5 year moving average).

Tibetan Plateau consists of high elevation stations for comparison with others at lower altitude. At Lhasa, summer temperature clearly warmed from the 1960s to 2007 (Figure 6.2.3), with an increase of 1.8°C. At Nagqu summer temperatures have increased between 1955 and 1995 as temperatures warmed by about 0.5°C (Figure 6.2.3). Summer temperatures in the Tibetan Plateau warmed since the 1960s.

At Darjeeling, summer temperatures decreased by about 0.5°C from 1880 to 1981 (Figure 6.2.4). However, temperature records at Darjeeling only run up 1981 making it difficult to be sure whether temperatures at that elevation would have warmed most recently like Mukteshwar, Kathmandu, Lhasa and Nagqu, (Figure 6.2.3), and Srinagar and Lahore (Figure 6.2.1). Summer temperatures at Cherrapunji were stable from 1903 to 1980 as temperatures were around 20°C (Figure 6.2.4). However, summer temperatures at Cherrapunji cooled from 1980 to the early 1990s, which was followed by warming period until 2009 (Figure 6.2.4). Summer temperatures increased at Guwahati until 1960 and then declined between 1960 and 1990 (Figure 6.2.4). Most recently, temperatures at Guwahati rose sharply from 1980 to 2009 with an increase of about 1°C (Figure 6.2.4). Summer temperatures in 2009 were much higher than in the 1960s. Guwahati and Cherrapunji both observed warming in summer, but Cherrapunji lagged behind as temperature increases took place about 10 years later. Summer temperatures at Shillong increased between 1900 and 1945, followed by a dip that only lasted until about 1990 (Table 6.2.4). Guwahati and Shillong observed similar patterns through the 20th century as the cool periods between 1940 and 1980 that prevented summer temperatures from continuing to rise. Temperature during summer across eastern Himalaya showed an increase, with temperatures reaching their highest most recently after the cooling period between 1945 and 1990.

6.3 Long term annual air temperature trends

Annual temperatures warmed at Srinagar from 1893 to 2008 (Figure 6.3.1). The first warming period at Srinagar was between the 1890s and late 1940s, but temperatures cooled from the late 1940s to 1970 (Figure 6.3.1). A second warming period occurred most recently from the mid-1970s to late 2000s when annual temperatures increased by about 1°C. However, temperatures in the 2000s did not exceed the earlier maxima in the late 1940s. At Lahore, annual temperatures were fairly steady between 1876 and 1950 (Figure 6.3.1). Between 1950 and 1970 annual temperatures dipped below the levels experienced before 1960. Most recently, annual temperatures warmed from 1980s to 2008 (Figure 6.3.1). Over a ten year period from 1998 to 2008 annual temperatures increased at Lahore and Srinagar. Annual temperatures at Lahore and Srinagar both cooled from 1950 and 1980 but most recently temperature rose to the highest in the early 21st century.

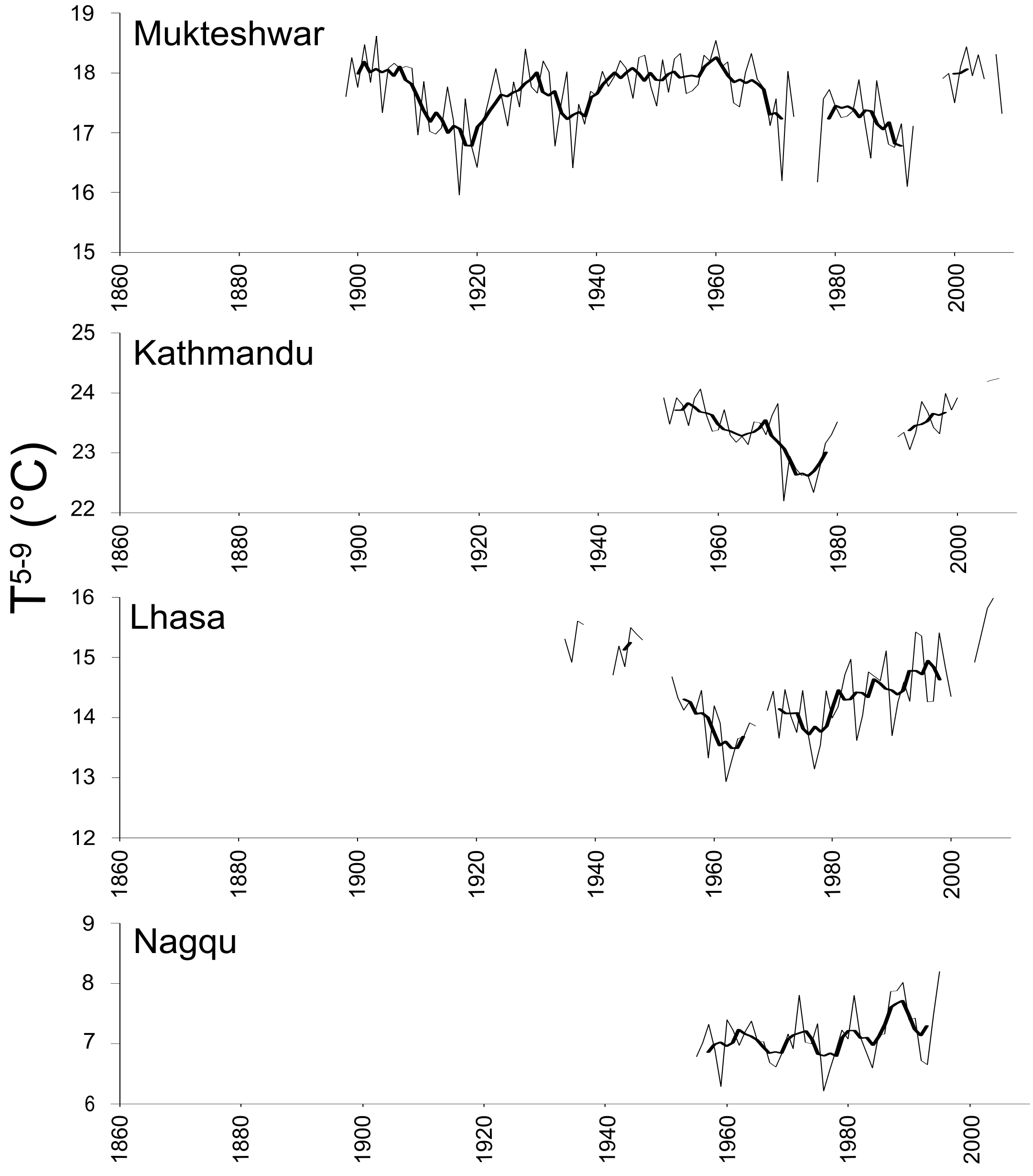

Figure 6.2.3 Summer average air temperature (1st May to 30th September) for Mukteshwar (2289m), Kathmandu (1302m), Lhasa (3656m) and Nagqu (4513m a.s.l.) (Thin line- real data, Thick line- 5 year moving average).

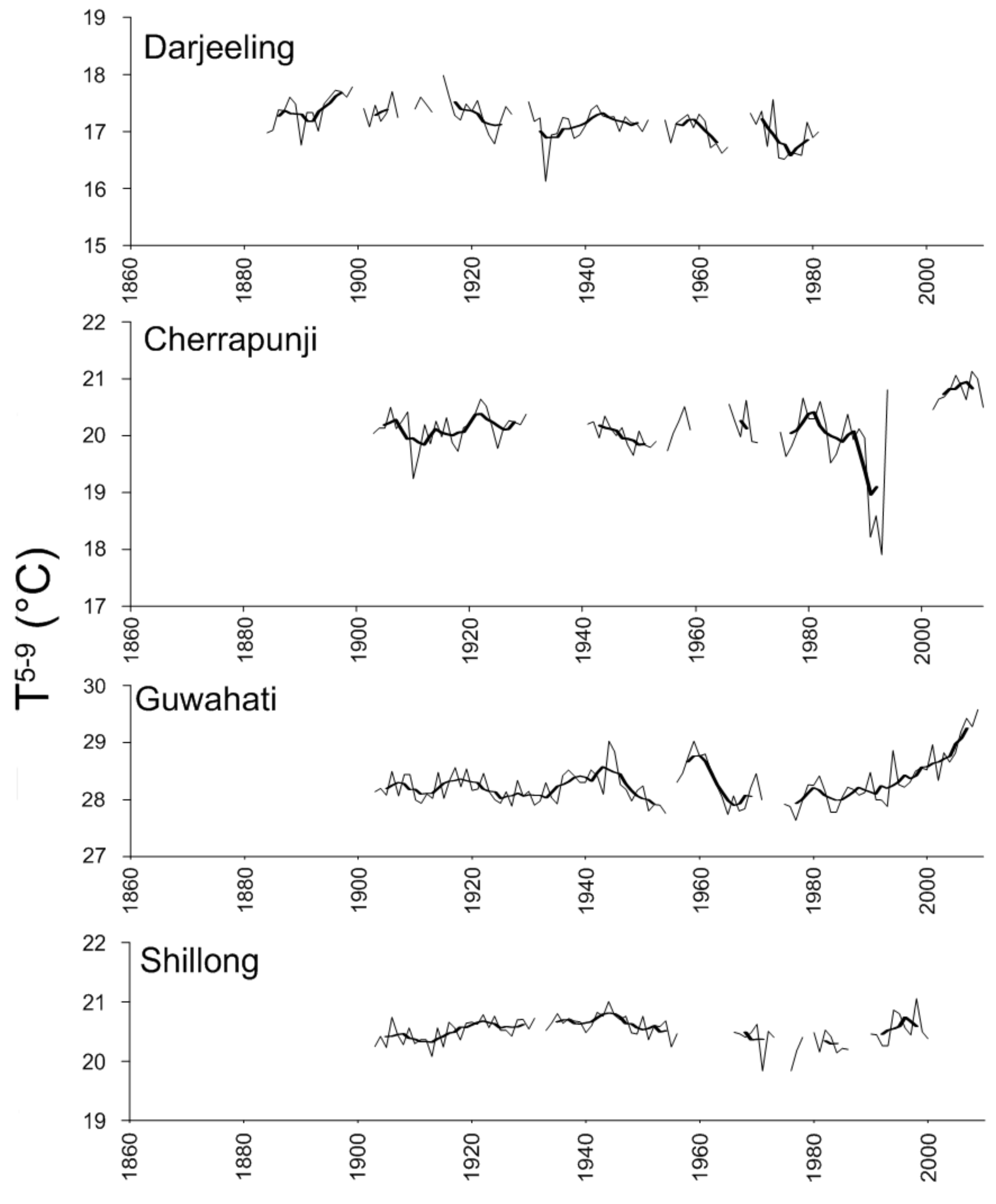

Figure 6.2.4 Summer average air temperature (1st May to 30th September) for Darjeeling (2128m), Cherrapunji (1489m), Guwahati (51m) and Shillong (110m a.s.l.) (Thin line- real data, Thick line- 5 year moving average).

Annual temperatures at Amritsar declined by about 0.9°C from the 1950s to the late 1990s (Figure 6.3.1). After 1989, annual temperatures begin to increase until 2006 (Figure 6.3.1). In general, annual temperatures decreased at Amritsar, but recent warming did not exceed the earlier maxima in the early 1950s. Annual temperatures increased by about 0.7°C at Shimla from 1900 to the mid-1980s (Figure 6.3.1). Annual temperatures at Shimla and Srinagar warmed over 20th century, but Srinagar cooled from 1950s to 1970s, while Shimla stayed steady t during this period. Cooling annual temperatures occurred at three stations; Srinagar, Lahore and Amritsar during 1960s and 1970s, but at Amritsar the decline lasted until 1989. Srinagar and Amritsar both had much lower annual temperatures in the 1950s. Annual temperatures warmed across north-west and western Himalaya, but the cooling periods from the 1950s was followed by recent rising temperatures since the 1980s.

Mukteshwar in the early 20th century experienced low annual temperatures with no great change between 1898 and 1940 (Figure 6.3.2). After 1940, annual temperatures rise through the rest of 20th century into the early 21st century (Figure 6.3.2). First warming period was between 1940 and the mid-1960s with annual temperatures increasing by about 1°C (Figure 6.3.2). Annual temperatures then cooled from the mid-1960s to the early 1990s. Second warming was most recently between 1999 and 2004 (Figure 6.3.2). Mukteshwar experienced two noticeable cool periods, but generally annual temperatures trended upwards across the last century and reaching warmest most recently. Annual temperatures declined at Kathmandu from the 1950s to the late 1970s (Figure 6.3.2) as temperatures decreased by about 1.2°C. Kathmandu and Mukteshwar both observed cooler periods in the 1970s, but annual temperatures at Kathmandu cooled much earlier. Annual temperatures at Kathmandu warmed most recently by about 1.3°C from the late 1970s to the late 1990s (Figure 6.3.2). Most recently, annual temperatures were just above the earlier maxima observed in the 1950s. In central Himalaya annual temperatures warmed through the 20th century but started to cool from around 1950 at Kathmandu but Mukteshwar lagged behind, but the warmest temperatures since the 1990s were experienced.

Annual temperatures at Lhasa declined by about 1°C between 1935 and the mid-1960s as the climate cooled dramatically (Figure 6.3.2). After this, annual temperatures started to rise from the mid-1960s to 2007, with a 1.3°C increase (Figure 6.3.2). Lhasa and Kathmandu experienced both cooling periods in the 1950s and 1960s, but both stations followed a warming trend, which started much earlier at Lhasa.

Annual temperatures at Guwahati experienced warming and cooling periods over the 20th century. First warming period was between 1900 and 1960 with a 1.1°C increase in annual temperatures (Figure 6.3.2). Much milder temperatures occurred from the mid-1960s to the mid-1980s, as annual temperatures dropped to similar levels experienced in the early 20th century. A second warming cycle showed a rise in annual temperatures from the mid-1980s to 2008 of 1.2°C (Figure 6.3.2). Annual temperatures at Guwahati generally increased through the twentieth century, but a cooling period suppressed warming until most recently, as the climate warmed. Similarly, annual temperatures in Shillong increased until the early 1960s (Figure 6.3.2). At Shillong annual temperatures then cooled until starting to warm again in 1990, eventually reaching the highest temperatures of the century in 1999 (Figure 6.3.2). The cooling period at Shillong is much shorter than at Guwahati, which dipped much earlier in 1960, but Shillong reached much warmer levels most recently in comparison to its earlier maximum in the early 1960s. Annual temperatures in the eastern Himalaya increased over the 20th century with recent warming exceeding the earlier maxima.

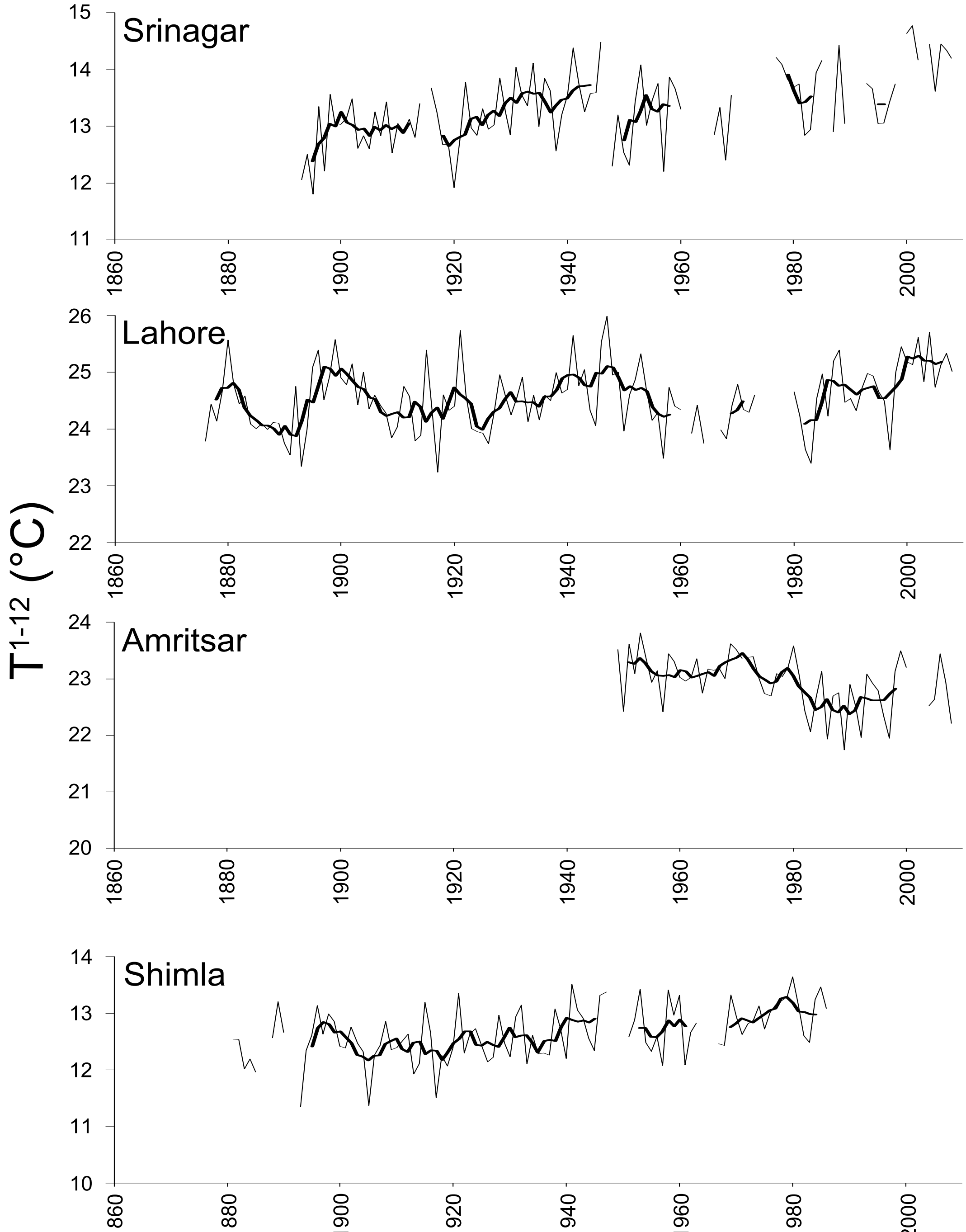

Figure 6.3.1 Average annual air temperature (1st January to 31st December) for Srinagar (1590m), Lahore (216m), Amritsar (232m) and Shimla (2143m) (Thin line- real data, Thick line- 5 year moving average).

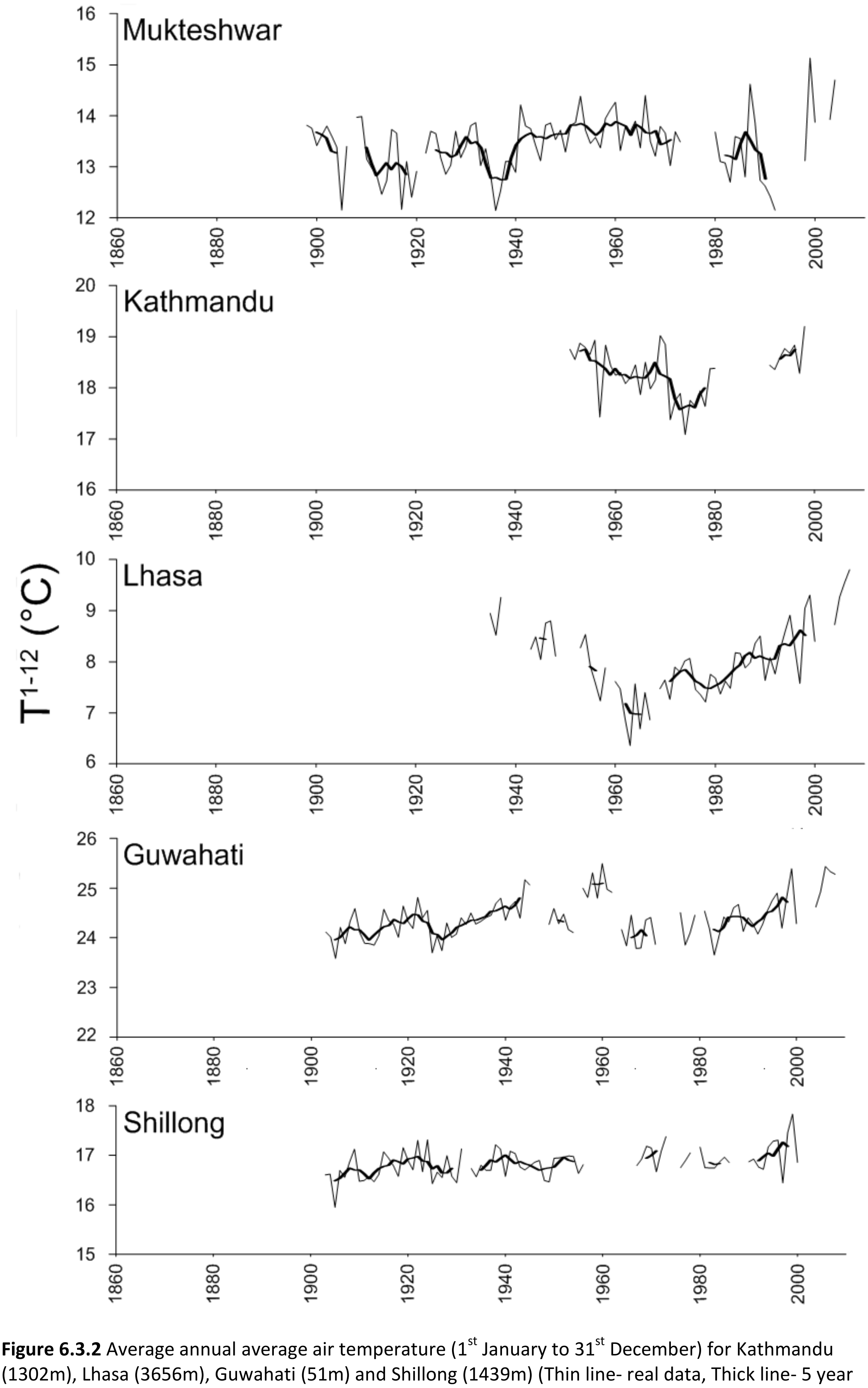

Figure 6.3.2 Average annual average air temperature (1st January to 31st December) for Kathmandu (1302m), Lhasa (3656m), Guwahati (51m) and Shillong (1439m) (Thin line- real data, Thick line- 5 year moving average).

6.4 Long term winter precipitation trends

Winter precipitation at Peshawar fluctuated steadily with no clear trends from 1863 to 2000 (Figure 6.4.1). Total winter precipitation can vary greatly with individual years receiving both large and small precipitation amounts, with huge contrasts at Peshawar, as 1978 received 625.6 mm compared with 41.1 mm in 1880. Winter precipitation cyclically fluctuated over time with high levels often followed by much lower levels. Moving averages at Peshawar were around 200 to 300 mm compared with actual data which has a much greater range of about 50 to 450 mm (Figure 6.4.1). Winter precipitation at Skardu Airport showed neither major increases nor decreases between 1906 and 1970 (Figure 6.4.1). Skardu Airport experienced a much wetter period but only for short time from 1969 to 1974. After 1974, winter precipitation at Skardu Airport declined until 1982, but after which precipitation started to increase (Figure 6.4.1). At Skardu Airport, precipitation increased from 359.7 mm in 1996, compared to a much lower amount of 53.6 mm in 1982. Winter precipitation slightly increased by just less than 20% at Srinagar from the 1890s to the 1980s (Figure 6.4.1). However, winter precipitation declined by about 18% between 1980 and 2009, showing that winter precipitation at Srinagar has not greatly changed over the 20th century with the gains and losses cancelling each other out. Winter precipitation at Gilgit fluctuated up and down but with no major changes between 1894 and 1999 (Figure 6.4.1). Winter precipitation on average is around 50 to 150 mm. Gilgit is similar to Peshawar and Srinagar as winter precipitation was stable over the 20th century. Winter precipitation in north-west Himalaya is generally stable, but with some small increases since 1960 at Peshawar and Skardu Airport, but Srinagar declined from 1970.

Winter precipitation at Bhuntar Airport was stable as precipitation fluctuated between 1900 and 1960 (Figure 6.4.2). Winter precipitation from 1900 to 1930 at Bhuntar Airport declined, but then increased between the 1930s and 1960. However, the trends are both relatively minor changes (Figure 6.4.2). Moving average at Bhuntar Airport is generally around 350-450 mm, but real data ranges much more from 200 to 600 mm. Winter precipitation at Shimla had a clear drying trend from

1865 to 1988 (Figure 6.4.2). Moving average declined from 531.06 in 1876 to 249.08 in 1986. Shimla experienced a significant decrease in winter precipitation of more than 50 % between 1876 and 1986. Western Himalaya observed a drying trend in winter precipitation over the 20th century at Shimla and Bhuntar Airport stations, but Shimla has showed a great reduction in winter snowfall.

At Lhasa there is very little winter precipitation with many fluctuations but no major trends from 1940 to 2000 (Figure 6.4.2). However, there are many gaps in data making it difficult to identify clear patterns. Moving average values were generally around 20 to 70 mm compared to real data which varied from 0 to 100 mm. Winter precipitation at Nagqu experienced a wet period between 1985 and 1995, but only over a small time and an insignificant change. Moving average at Nagqu ranges from 75 to 125 mm, whereas the real data has a greater range between 50 and 175 mm. Winter precipitation in the Tibetan Plateau generally showed little trend, but longer data sets are needed to identify long term changes in the Plateau.

Dehradun experienced a fall in winter precipitation from 1901 to 1970, but relatively small with a loss of about 8% (Figure 6.4.3). In general precipitation during winter varied between 200 and 400 mm and around 200 to 300 mm for moving average. Winter precipitation declined at Mukteshwar from 1900s to 2009 (Figure 6.4.3). At Mukteshwar the most significant decline in winter precipitation was between 1900 and 1970 with 16% loss. Winter precipitation at Dehradun and Mukteshwar observed a drying period in the 20th century from 1900 to 1970, but Mukteshwar showed a more significant change. Moving average for winter precipitation at Mukteshwar is generally between 250 to 350 mm and 150 to 500 mm for real data. Real precipitation data can vary greatly at Mukteshwar so often not entirely accurate representation of trends, whereas the moving average can give a better indication. Winter precipitation increased from 1971 to 1990 (Figure 6.4.3). Moving average increased from 344 mm in 1975 to 504 in 1988. However, this increase was over a short period making it difficult to identify long term trends in the central Himalaya.

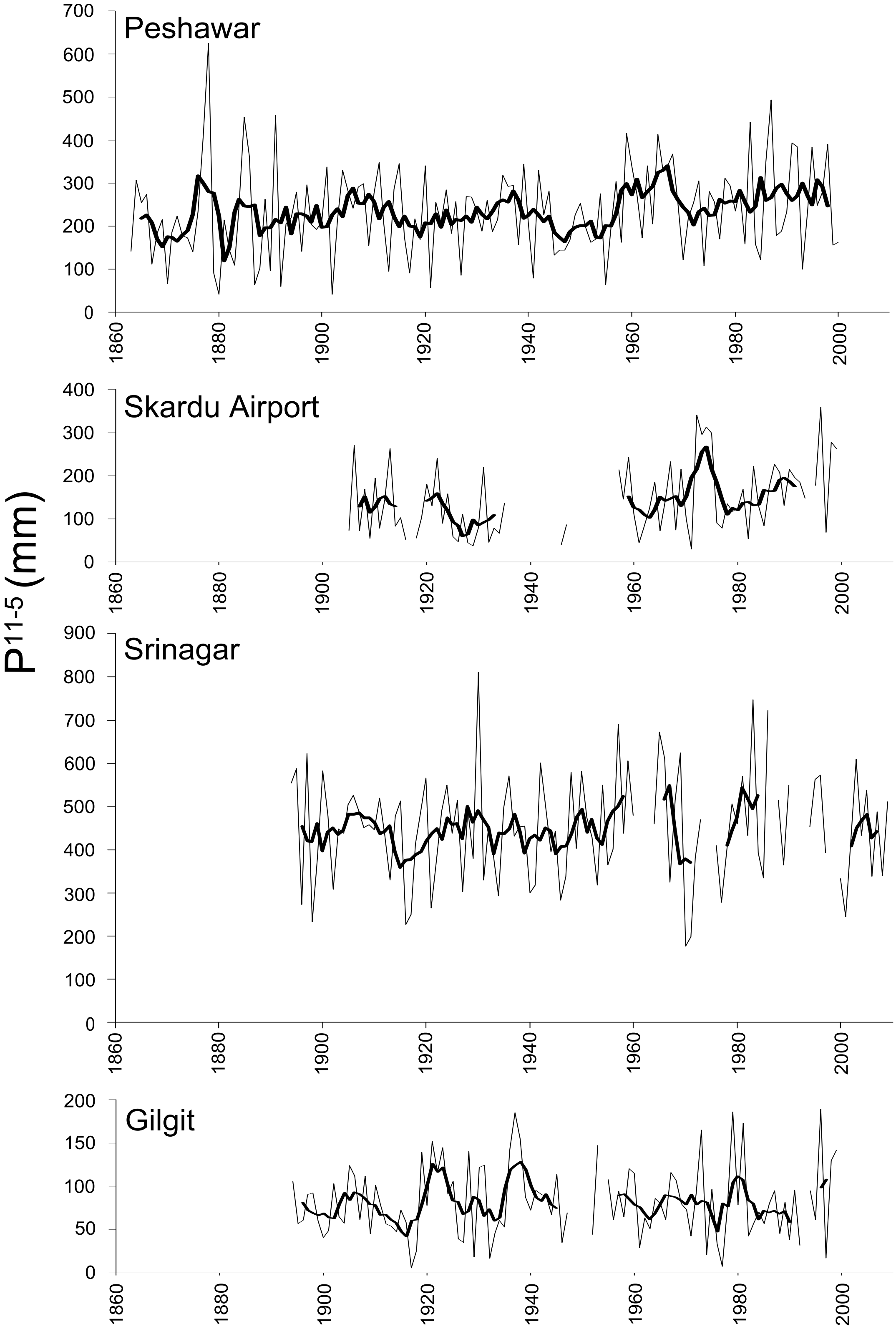

Figure 6.4.1 Total winter precipitation (1 November to 31 May) for Peshawar(359m), Skardu Airport (2222m), Srinagar (1590m) and Gilgit (1465m) (Thin line- real data, Thick line- 5 year moving average).

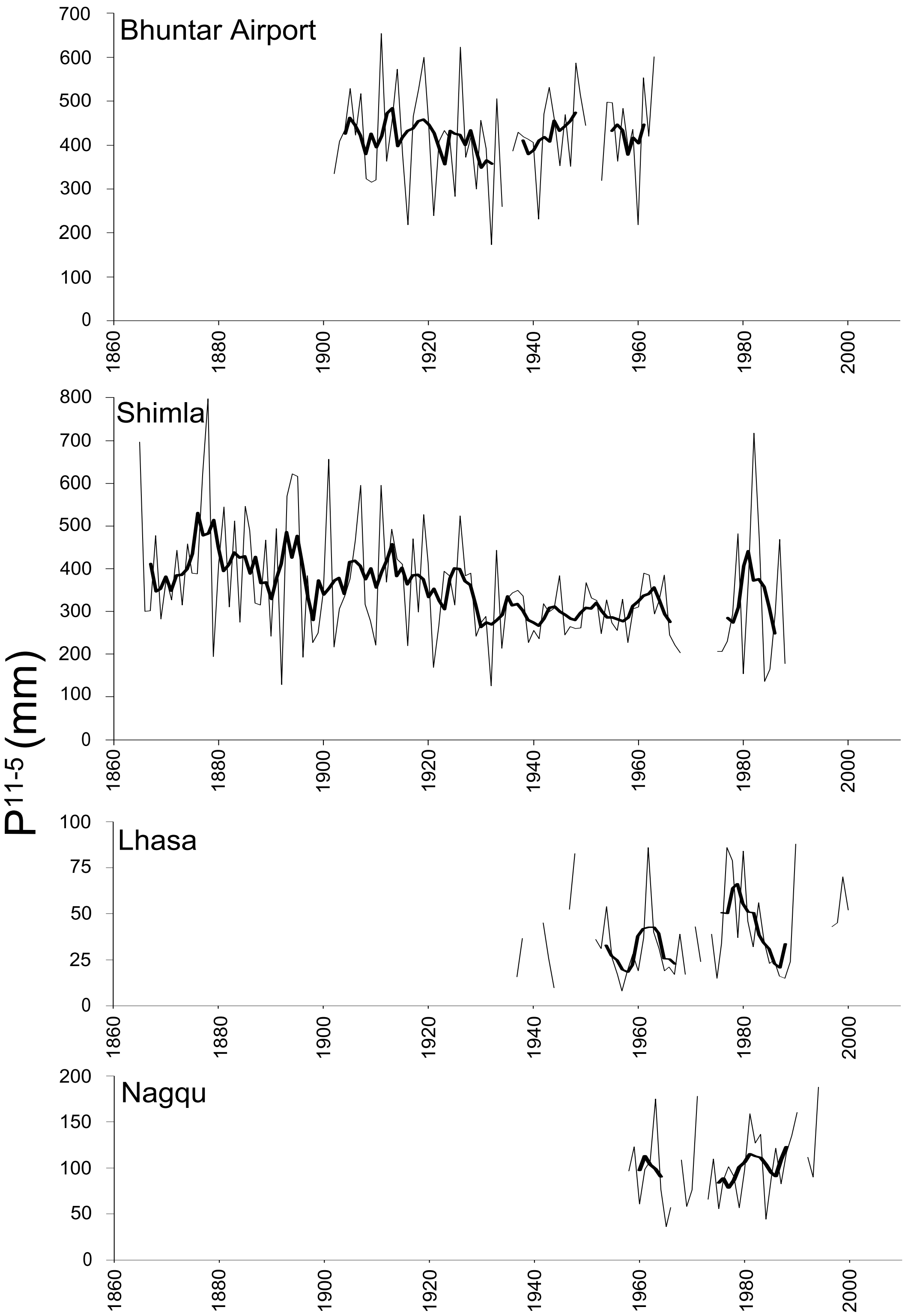

Figure 6.4.2 Total winter precipitation (1 November to 31 May) for Bhuntar Airport (1092m), Shimla (2143m), Lhasa (3656m), Nagqu (4513m) (Thin line- real data, Thick line- 5 year moving average).

During winter, precipitation at Kathmandu observed a rise from 1952 to 2000 (Figure 6.4.3). Kathmandu increased by about 39% in moving averages from 1974 to 1998. Winter precipitation at Kathmandu and Lete experienced wet periods since 1970 and 1970 respectively. Winter precipitation in central Himalaya experienced a drier period as precipitation declined until 1970, but in Nepal both stations saw wet periods after 1970.

Winter precipitation at Darjeeling generally declined from 1880 to 1972, with two drying periods, firstly from 1895 to 1930 and secondly between 1952 and 1972 (Figure 6.4.4). Winter precipitation declined more significantly from 531.8 in 1935 to 345.08 in 1972 which a loss of about 35%. Cherrapunji observed little change in total winter precipitation from 1874 to the early 1940s (Figure 6.4.4). Winter precipitation at Cherrapunji increased between 1940to 1960, but after 1960 data is scarce making it difficult identify trends. At Guwahati, winter precipitation from 1868 to 2010 showed no significant trends as levels stayed fairly stable. At Shillong there is a clear increase in winter precipitation from the 1870s to 1950 with a 55% increase (Figure 6.4.4). Winter precipitation at Shillong from 1950 onwards is much drier until 2000, as winter precipitation especially declines from 1950 to 1960 (Figure 6.4.4). Real data decreased from 1065.8 mm in 1946 to 600 in 2000. However, precipitation data collected at Shillong from 1950 to 2000 has many gaps making it difficult to identify changes. Between 1940 and 1950 Shillong and Cherrapunji observed an increase in winter precipitation. Winter precipitation started to decline around 1950 at Shillong, with Darjeeling soon following in 1956. Winter precipitation across eastern Himalaya has stayed fairly stable with some periods of drying and wetting but not over substantially long periods.

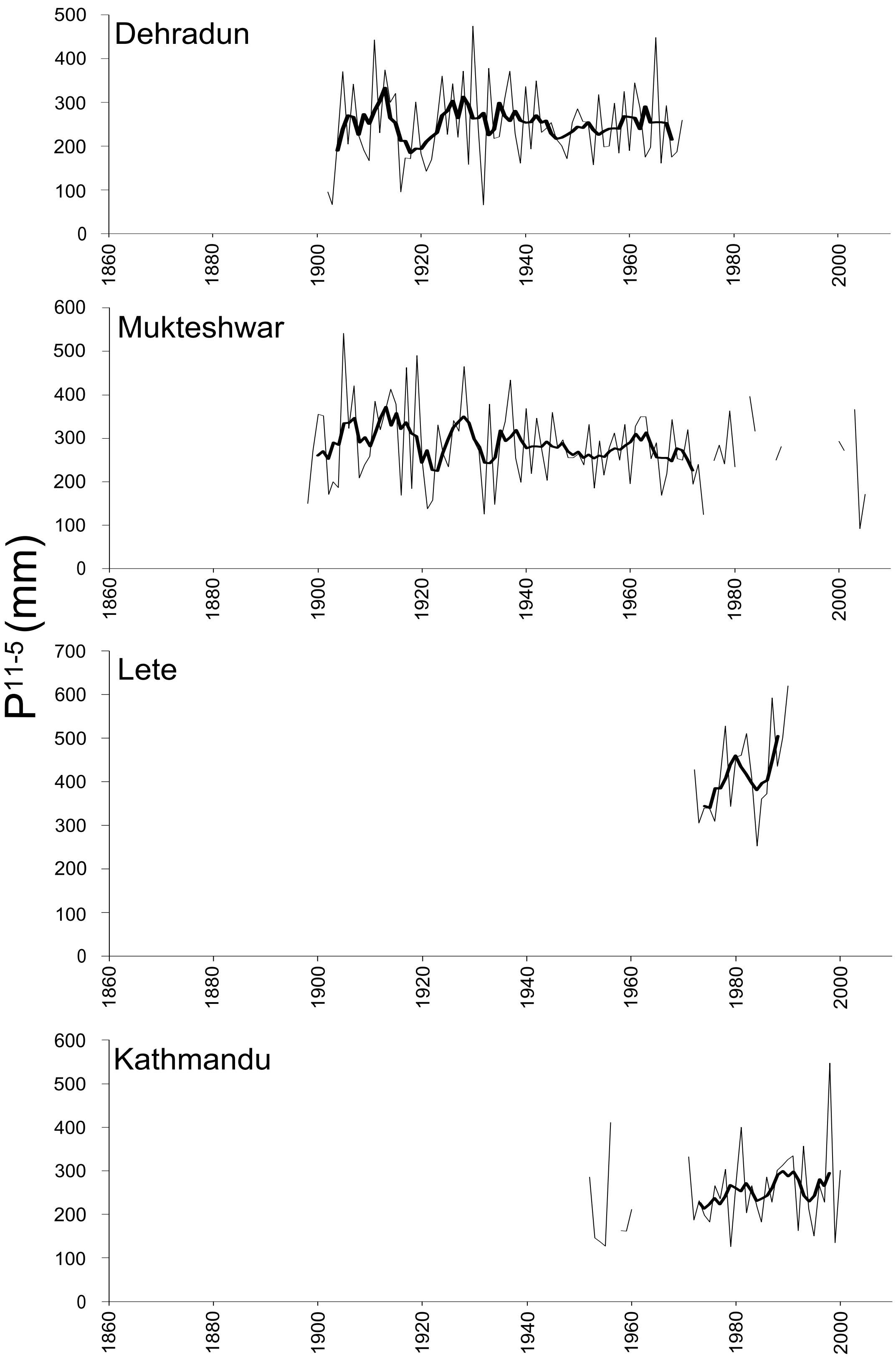

Figure 6.4.3 Total winter precipitation (1st November to 31st May) for Dehradun (653m), Mukteshwar (2289m), Lete (2208m) and Kathmandu (1302m) ((Thin line- real data, Thick line- 5 year moving average).

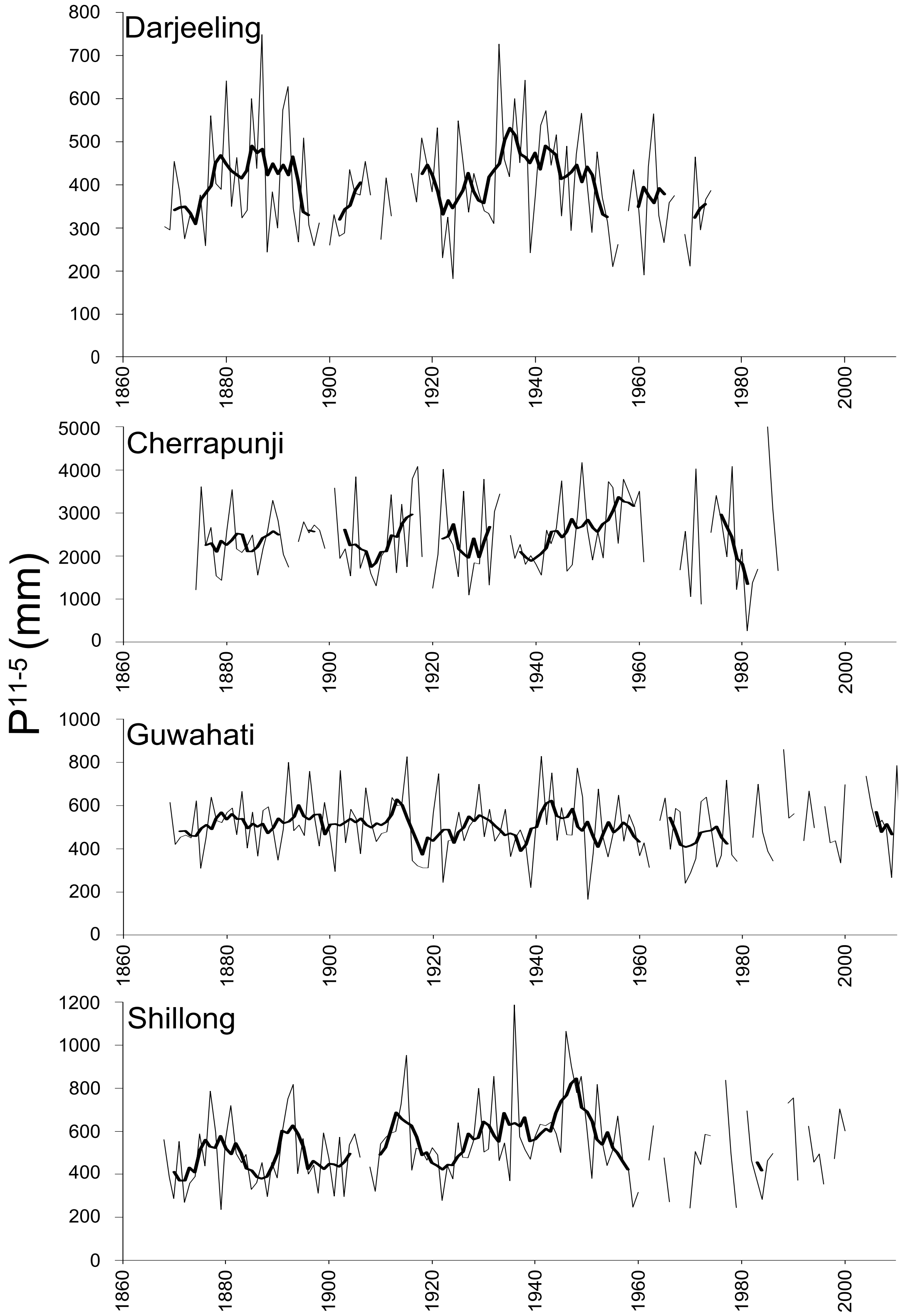

Figure 6.4.4 Total winter precipitation (1st November to 31st May) for Darjeeling (2128m), Cherrapunji (1489m), Guwahati (51m) and Shillong (1439m)(Thin line- real data, Thick line- 5 year moving average).

6.5 Long term summer precipitation trends

At Peshawar, summer precipitation was steady from 1862 to 2000, with moving average very low around 60 to 180 mm (Figure 6.5.1). Peshawar experienced a wet period was between 1975 and 1996, but summer precipitation reached its maximum level of 556.9 mm in 1892 which is an anomaly year (Figure 6.5.1). Downward trend in summer precipitation is clear at Skardu Airport with a 47% loss from 1907 to 1969 (Figure 6.5.1). After 1970, summer precipitation steadied and gradually increased to similar levels that were experienced in early 20th century. At Shimla, summer precipitation was fairly stable until 1960s, until declining by 50% between 1960 and the mid-1980s (Figure 6.5.1). During summer, precipitation at Mukteshwar decreased over the 20th century (Figure 6.5.1). However, the monsoon rainfall took a while to experience a significant decline as summer precipitation decreased by 50% from 1980 to 2009. Summer precipitation at Kathmandu was stable from 1951 to 1990, until experiencing a short but insignificant wet period between 1990 and 2003. Summer precipitation in north-west Himalaya declined, but the most significant losses of 50% at Shimla and Mukteshwar in the lower Sutlej and north-west Ganges regions while monsoon rainfall has increase since 1990 in Nepal.

Summer precipitation at Lhasa experienced two wet periods from 1941 to 1968 and secondly from 1983 to 2003 (Figure 6.5.2). These increasing trends are divided by a drying trend from 1968 to 1983 (Figure 6.5.2). Summer precipitation at Guwahati was mostly stable between 1867 and 2010 (Figure 6.5.2). Cherrapunji also showed no clear wetting or drying periods as summer precipitation is steady through the 20th century into the 21st century (Figure 6.5.2). Cherrapunji receives a considerably large amount of summer precipitation, with precipitation ranging from 6,000 to 10,000 mm, which is much greater than other stations in the eastern Himalaya. Summer monsoon rainfall at Guwahati and Cherrapunji changed little over 20th century (Figure 6.5.1). However, Guwahati generally receives around 750 to 1250 mm compared to 6,000 to 10,000 mm at Cherrapunji. Summer precipitation at Shillong declined by 10% from 1867 and 1999 and 15% from 1950 to 1999 (Figure

6.5.2). Darjeeling experienced a reduction in summer precipitation from 1869 to 1978 by about 40% (Figure 6.5.2). The wettest period at Darjeeling was in early 20th century from 1868 to 1905. Summer precipitation declined at both Darjeeling and Shillong over the 20th century. Summer precipitation in the eastern Himalaya has been steady at two lowland stations, but there are significant declines of 40% at Darjeeling and 10% at Shillong.

6.6 Long term annual discharge trends

Annual discharge in the Gilgit River at Gilgit initially reduced from 1960 to the early 1980s decreasing from 11586.18 (x 10^6 m^3) in 1960 to 6654.7 (x 10^6 m^3) in 1982 (Figure 6.6.1). River flow declined by 42%, which is a huge reduction in snow and ice melt. During the 1990s, annual discharge at Gilgit increased and peaked in 1999 with 10988.24 (x 10^6 m^3) (Figure 6.6.1). Annual River flow in the Gilgit River at Gilgit over 40 years is fairly constant with only a 0.5% loss (1960-1999).Annual runoff in the Hunza River at Dainyor Bridge showed a clear downward trend from 1966 to 1999 with a 24% loss (Figure 6.6.1). However, discharge data needs to be compared with meteorological parameters to identify the cause of declining runoff.

Annual discharge in the Shyok River at Yugo showed little trend from 1973 to 1997 with a minimal increase of 1% (Figure 6.6.1). Shyok and Gilgit Rivers both had little trend from 1970 to the late 1990s as annual discharge fluctuated without major changes. Annual runoff in the Sutlej River at Khab reduced from early 1970s to early 2000s, with a 31% loss between 1972 and 2002 (Figure 6.6.1). Along the Sutlej River at Khab, annual runoff decreased most significantly from 1991 to 2002 after a more stable period (Figure 6.6.1). River flows annually declined along the Hunza and Sutlej Rivers since 1970. River flow in the upper Indus was generally stable, but declined by 24% in the Hunza River at Dainyor Bridge and in the upper Sutlej by 31%.

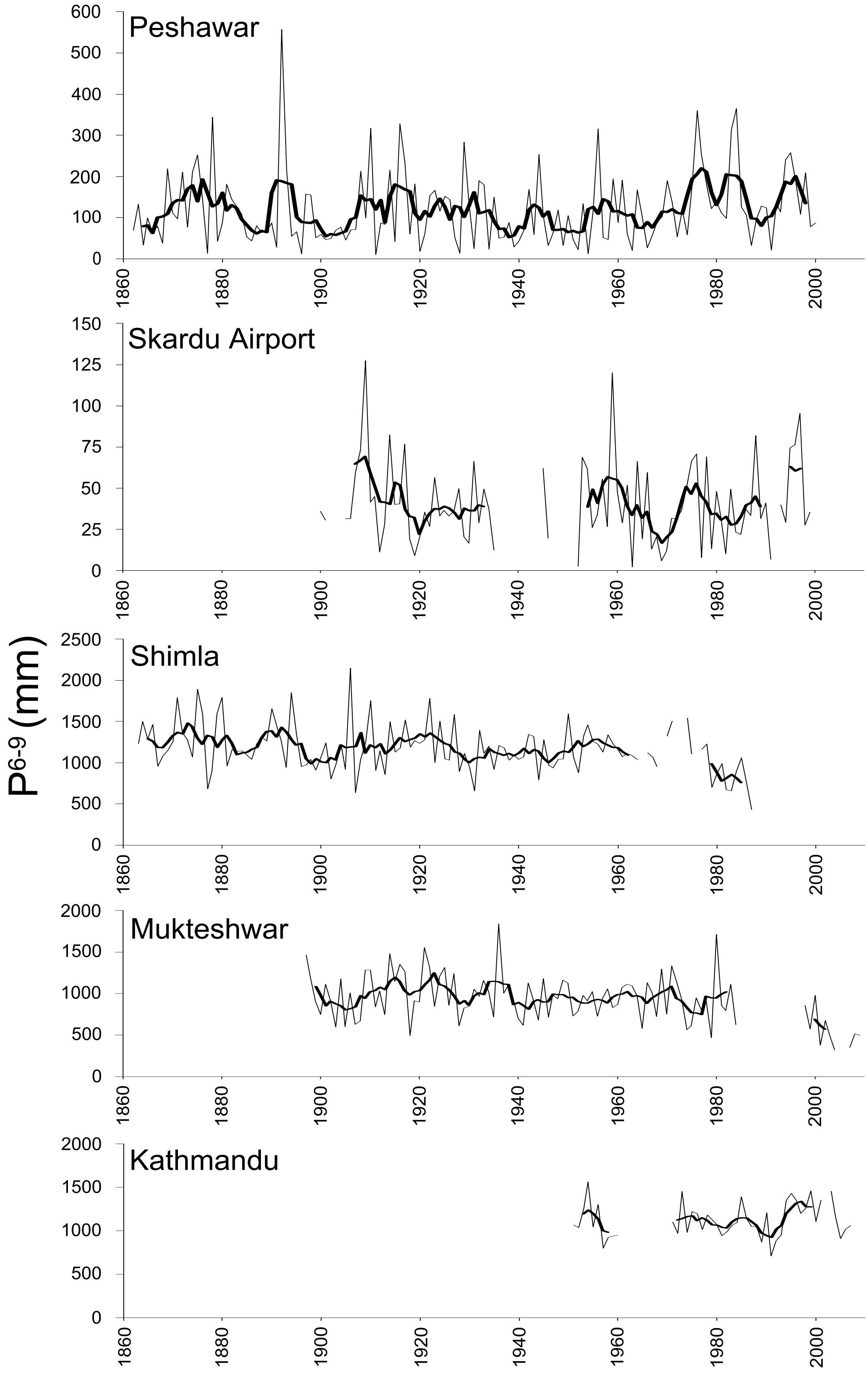

Figure 6.5.1 Total summer precipitation (1st June to 30th September) for Peshawar (359m), Skardu Airport (2222m), Shimla (2143m), Mukteshwar (2289m) and Kathmandu (1302m) (Thin line- real data, Thick line- 5 year moving average).

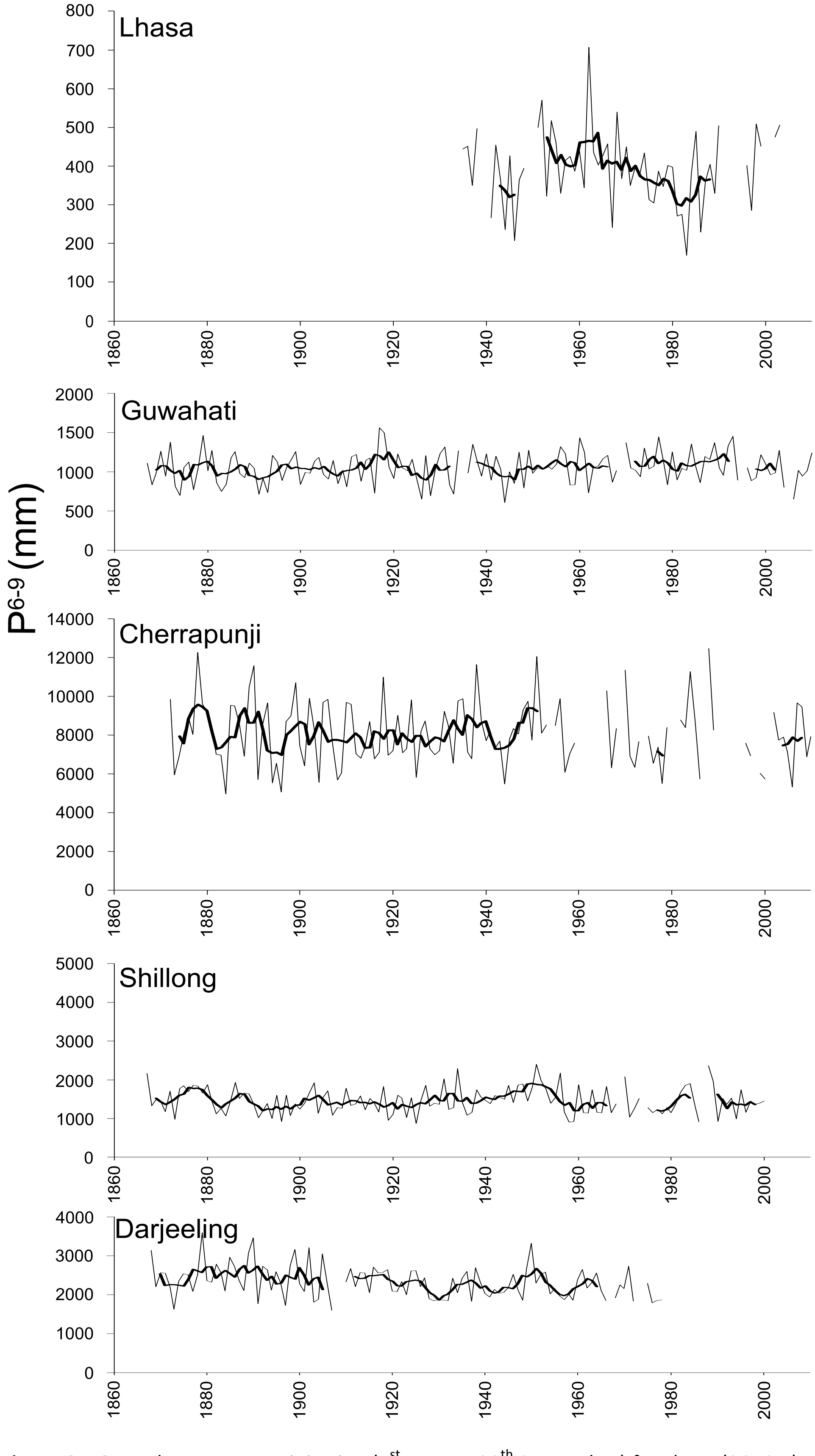

Figure 6.5.2 Total summer precipitation (1st June to 30th September) for Lhasa (3656m), Guwahati (51m), Cherrapunji (1489m), Shillong (1439m) and Darjeeling (2128m) (Thin line- real data, Thick line- 5 year moving average).

Across Nepal snow and ice melt drains into the Ganga River, but greater contributions downstream arise from summer monsoon. Annual river flow in the Karnali River at Chisapani was stable from 1962 to early 1990s, with only an insignificant loss of 0.2% (Figure 6.6.2). Karnali River at Chisapani observed a dip in annual runoff from the mid-1960s, but discharge recovered and stabilised after 1970. Annual discharge in the Karnali River at Asara Ghat showed no major changes between 1975 to early 1990s with a minor loss of 3% (Figure 6.6.2). Annual river flow at Asara Ghat dipped in the mid-1960s, which followed runoff at Chisapani with a decline since the 1960s. Kali Kandaki River at Setibeni, experienced a decline by 12.3% in annual discharge from 1973 to 1993 (Figure 6.6.2). Annual runoff across Nepalese river basins dipped in the mid-1960s with Setibeni following Chisapani and Asara Ghat. Annual discharge in the Narayani River at Narayanghat showed three small periods of peaks and troughs which generally last around 10 years (Figure 6.6.2). Since 2000, annual discharge declined at Narayanghat, but will this trend continue to fall or follow the up and down trend observed since 1960? Annual runoff in the Langtang River at Langtang decreased from 1988 to 1998 (Figure 6.6.2). The data set for Langtang is very short with only 10 years, which makes it difficult to identify any long term trends at high elevation. Declining annual discharge at Langtang from 1988 to 1998 is similar to Setibeni, Chisapani, Asara Ghat and Narayanghat as they fall during the same period. Annual river flow in the tributaries of the Ganges across central Himalaya dipped in the mid-1960s, followed by a recovery but runoff stayed relatively stable from the 1960s to the 2000s. The Kali Kandaki at Setibeni was the only noticeable decline.

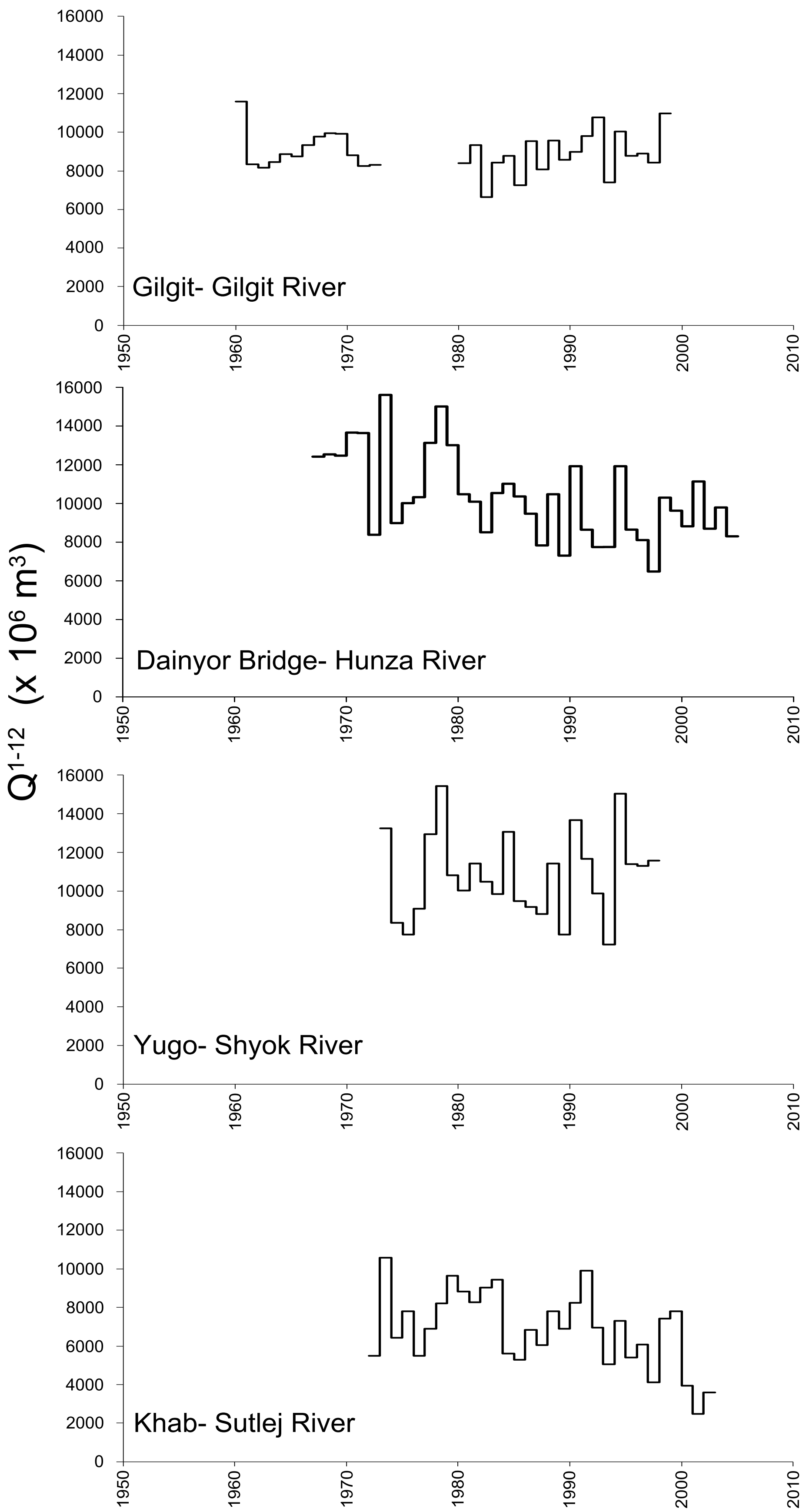

Figure 6.6.1 Annual discharge (1st January to 31st December) in the Indus and Sutlej River catchments. Elevations: Gilgit (1457m), Dainyor Bridge (1432m), Yugo (2434m) and Khab (2550m).

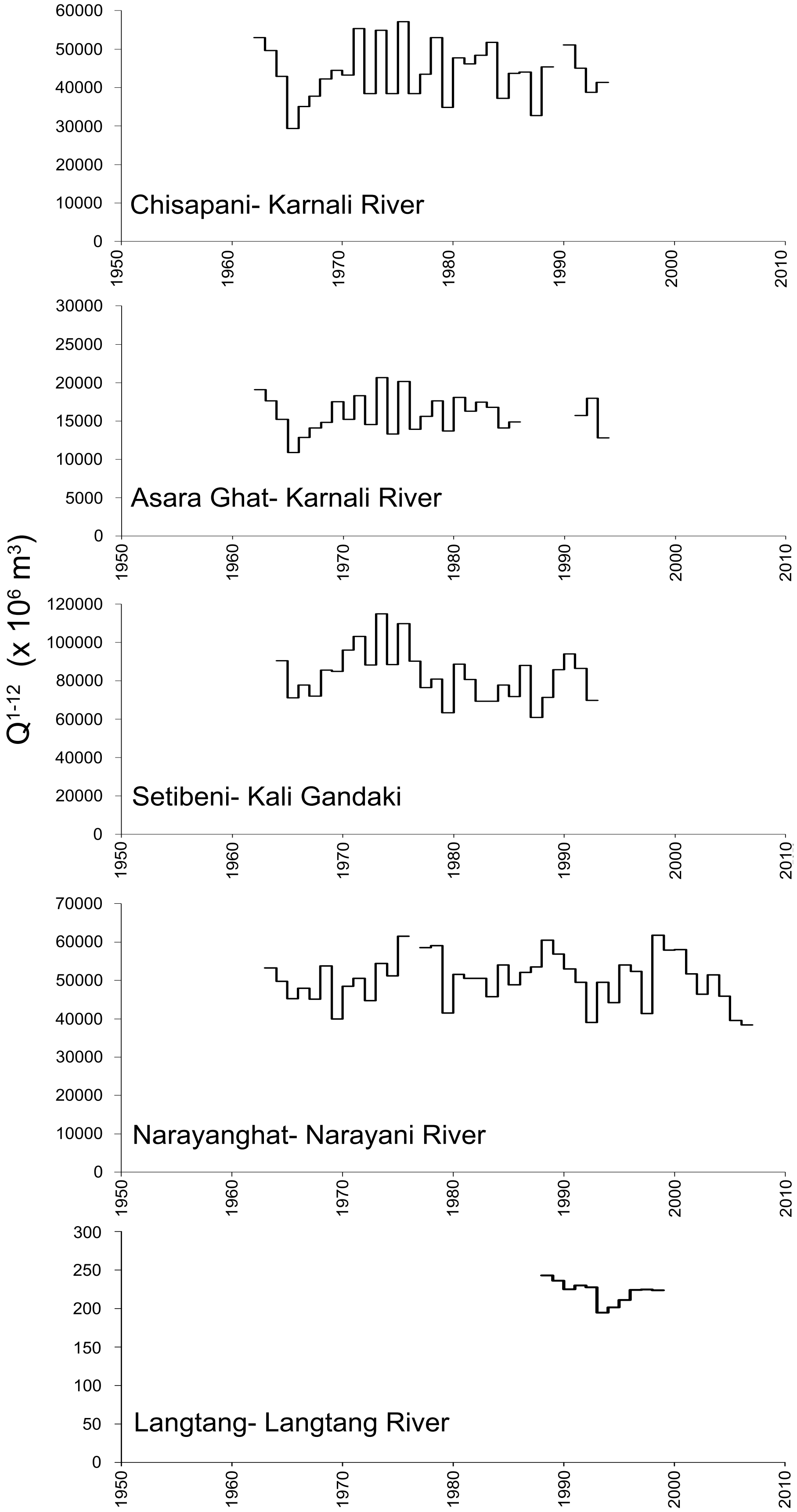

Figure 6.6.2 Annual discharge (1st January to 31st December) across Nepal, Ganges River catchment. Elevations: Chisapani (191m), Asara Ghat (629m), Setibeni (410m), Narayanghat (189m) and Langtang (3688m).

7. Discussion

7.1 Evaluation of data

Although the intention was to use long homogenous series, the available data are limited in many aspects. Ideally, records would be uninterrupted and located at high elevation sites to reflect climatic change in highly glacierised catchments. There are large variations in the length of air temperature records with Lahore, Srinagar, Shimla, Mukteshwar, Cherrapunji, Guwahati and Shillong each having a hundred years or more of data, compared with the shorter records of Gilgit, Kathmandu, Nagqu and Amritsar. Precipitation records also range in length of series with Peshawar, Srinagar, Gilgit, Shimla, Mukteshwar, Guwahati and Shillong being longer, whereas Lhasa and Nagqu, Lete, Kathmandu, Bijayapur and Khumjung are shorter. Particularly in Nepal, it was difficult to find complete enough data sets for both air temperature and precipitation. Precipitation records mainly range from 1970 to 1990 and only one station was available in Nepal for air temperature outside this period. To find long term changes in meteorological conditions in Nepal, correlations was used to compare short records with overlapping portions of longer data records at stations located to the east and west.

As discharge data records range from 1960 to 2006, only recent changes in river flows can be analysed. Runoff records collected cover the upper Indus, upper Sutlej and tributaries of the Ganges in central Himalaya, but there are no data available in the Brahmaputra basin in the far eastern Himalaya. The Central Water Commission highlighted that river flow at gauging stations across the major rivers of India are 'classified' thus not accessible (Ministry of Water Resources, 2013), meaning there are limited available data for research into river flows draining from partially-glacierised basins. Farakka along the Ganges River was used for annual hydrographs in the eastern Himalaya (Figure 6.1.4). However, Farakka was not used for annual long term trends because it is too far downstream and water has been abstracted at dams so that the record does not reflect runoff from high mountain regions.

Collected data for air temperature and precipitation have many missing months of data, which are represented as -999.9 in the data set. However, this leads to missing years of data, as one missing month prevents calculation of winter and summer averages or totals. This leaves short or sometimes very long periods of missing data. Precipitation records in India often disappear after 1970, leaving large gaps in recent records particularly unhelpful for comparison with temperature and river flows. The intention to identify whether climate change is enhanced with elevation was hindered by there being limited data and few stations above 2 000 m a.s.l. As Stott and Thorne (2010) highlighted, that there are virtually no stations in Himalaya above 2,000 metres (Figure 7.1). With poor coverage of high altitude stations, it is difficult to determine whether precipitation or air temperature is accelerated across the Himalaya.

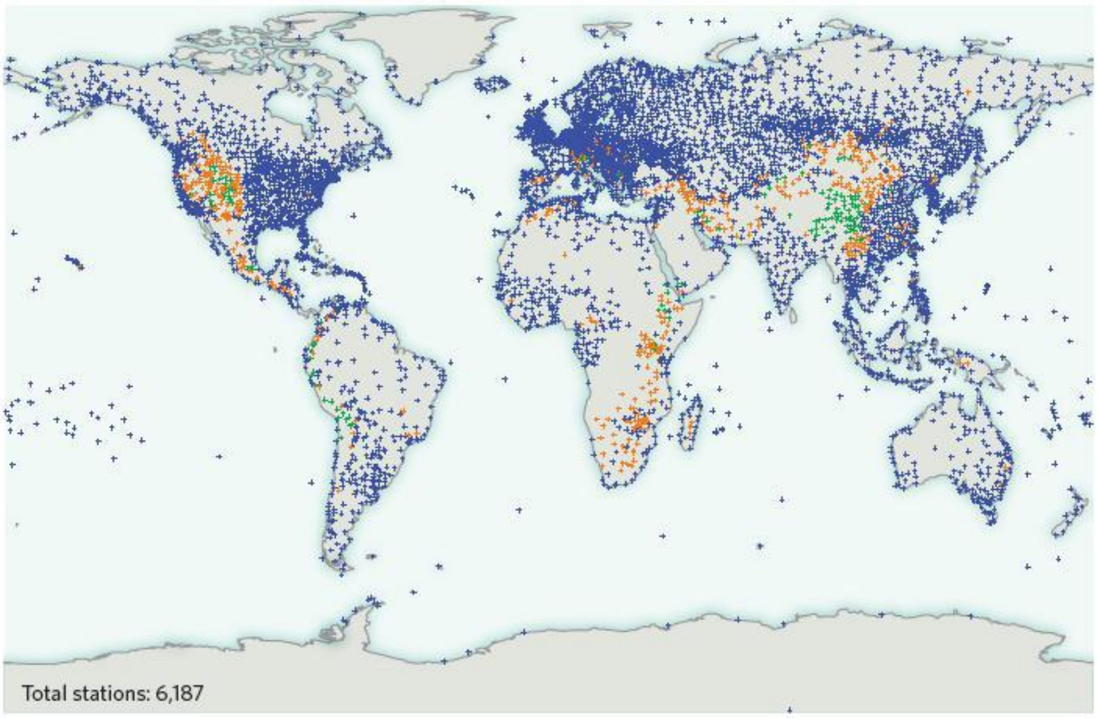

Figure 7.1 Map of stations that report at three hourly intervals for at least 15 years with stations in blue below 1,000 metres; orange are at 1,000 to 2,000 metres; green are higher than 2,000 m a.s.l. (Stott and Thorne, 2010).

7.2 Glaciologically-relevant graphs

Both air temperature and precipitation strongly influence Himalayan catchments, as they ultimately determine whether glaciers advance or decline, and precipitation contributes directly to runoff. In years of low winter precipitation followed by warm summers, ice free area will increase, but glaciers will increase in mass in years of heavy winter snowfall followed by cool summer temperatures. However, this does not occur uniformly along the Himalayan arc due to the summer monsoon strengthening from west to east, while winter westerlies are very influential in the north-west Himalaya. Therefore it is important to determine which season contributes the majority of precipitation for each catchment. Precipitation in the north-west falls mainly during winter whereas in the central and eastern Himalaya precipitation is at its highest in summer. However, in western Himalaya both summer and winter precipitation influences runoff. Ultimately changes in meteorological conditions indicate current and future river flows, which determine the amount of discharge provided to downstream lowland areas. As the majority of discharge occurs during summer, therefore discharge from May to September is related in this section to summer temperature, and winter and summer precipitation.

7.2.1 North-west Himalaya catchment

Winter westerlies provide a large amount of snowfall as a contribution to runoff in the heavily glacierised north-west Himalaya. Summer temperatures at Lahore in the lowlands and at Srinagar at higher elevation follow similar trends over the last century. Summer temperatures warmed from 1900 to 1950, followed by a cool period until around 1985 (Figure 7.2.1). Most recently, summer temperatures during summer have risen especially at Srinagar (Figure 7.2.1), which has increased available heat energy to melt snow and ice. However, warming temperatures do not necessarily lead to enhanced runoff, as the amount of snow and ice melted depends also on the amount of precipitation. Srinagar precipitation during winter increased by 20% from the 1890s to the 1980s (Figure 7.2.1), but decreased by 18% between 1980 and 2009 (Figure 7.2.1). Winter precipitation at Gilgit over the 20th century showed no clear wetting or drying trend (Figure 7.2.1). However, more recently winter precipitation increased from 1992 to 1999 insignificantly (Figure 7.2.1). This would have little impact on halting glacier area loss. A long sustained wet period would be needed for glaciers to grow.

Summer river flow in the Gilgit River at Gilgit dipped from 1960 to 1980, but then increased more recently (Figure 7.2.1). Summer runoff at Gilgit decreased by 46.7% (1960-1982), compared to a rise of 74.8% from 1982 to 1998. Between 1960 and the 1990s runoff during summer in the Gilgit River showed no major losses or gains. Summer discharge in the Hunza River at Dainyor Bridge declined by nearly a 25% from 1966 to 1998 (Figure 7.2.1). As available heat energy increases and enhances the melting of snow and ice with recent warming summer temperatures it might be thought that river flow would be much greater. However, declining winter precipitation at Srinagar since 1980, and stable runoff at Gilgit, suggest that there was insufficient snow to melt, hence summer river flows in the Hunza River declined - a 25% loss between the mid-1960s and the late-1990s. Summer runoff was steady in the Gilgit River at Gilgit in the upper Indus even though summer temperatures warmed since the mid-1980s again because there was insufficient snow accumulation in winter. Across north-west Himalaya, rising summer temperatures have not been enough to increase runoff as there is insufficient total winter precipitation available for melting during summer, with summer runoff either stable or declining.

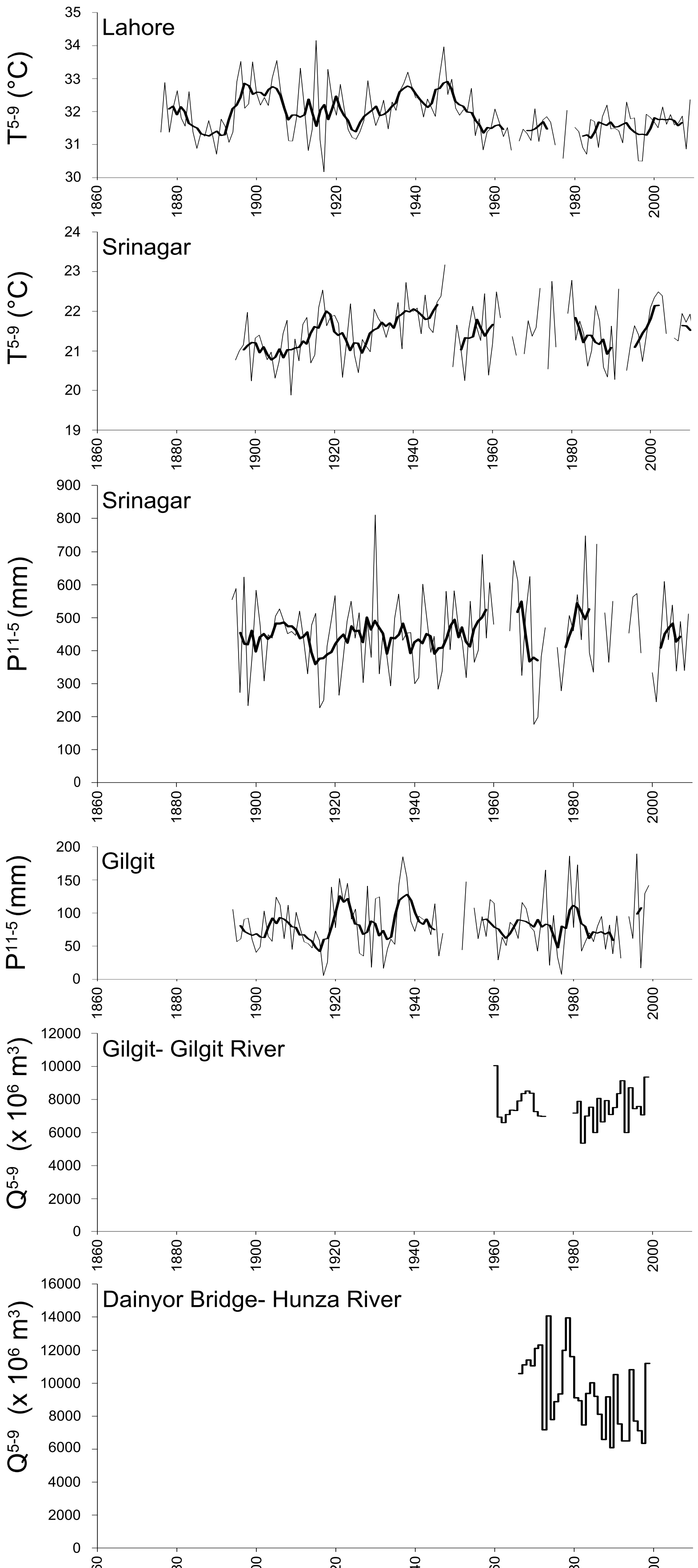

Figure 7.2.1 Glaciologically relevant meteorological variables in north-west Himalaya compared against summer discharge from the upper Indus River.

7.2.2 Western Himalaya catchment

In the western Himalaya region, monsoon influence gradually strengthens, with both winter and summer precipitation influencing runoff in the Sutlej River. In the early 20th century, summer temperatures cooled from 1880 to 1920, but then warmed until 1950 at Shimla (Figure 7.2.2). After 1950, summer temperatures at Shimla and Amritsar cooled until 1970, but Shimla warmed from 1970 to 1986 and Amritsar took until 1985 to warm slightly (Figure 7.2.2). Winter precipitation at Bhuntar showed no clear wetting or drying trend between 1902 and 1963 (Figure 7.2.2). At Shimla, winter precipitation declined by about 50% from 1876 to 1986 (Figure 7.2.2). However, in Shimla there is a greater percentage of annual precipitation falling between June to September. Average summer precipitation was 1170 mm with 353 mm during winter from 1863 and 1988. Summer precipitation at Shimla declined by about 50% from 1960 to the-mid 1980s (Figure 7.2.2). Both summer and winter precipitation experienced a decline, but more importantly there was a greater loss during summer when total precipitation is at its highest.

Summer runoff in the Sutlej River at Khab decreased by 34% from 1972 to 2002 (Figure 7.2.2). As summer temperatures have warmed since 1985, as at Amritsar, snow and ice melt in the high mountains will have been enhanced. However, summer discharge has declined in the Sutlej River at Khab, because of the decrease in winter and summer precipitation. There has been insufficient snow to sustain glacier areas, and high monsoon rainfall is needed as the monsoon contributes a large fraction of runoff. However, Shimla precipitation data exists only up to 1987, making it difficult to compare assess declining runoff in a warming climate.

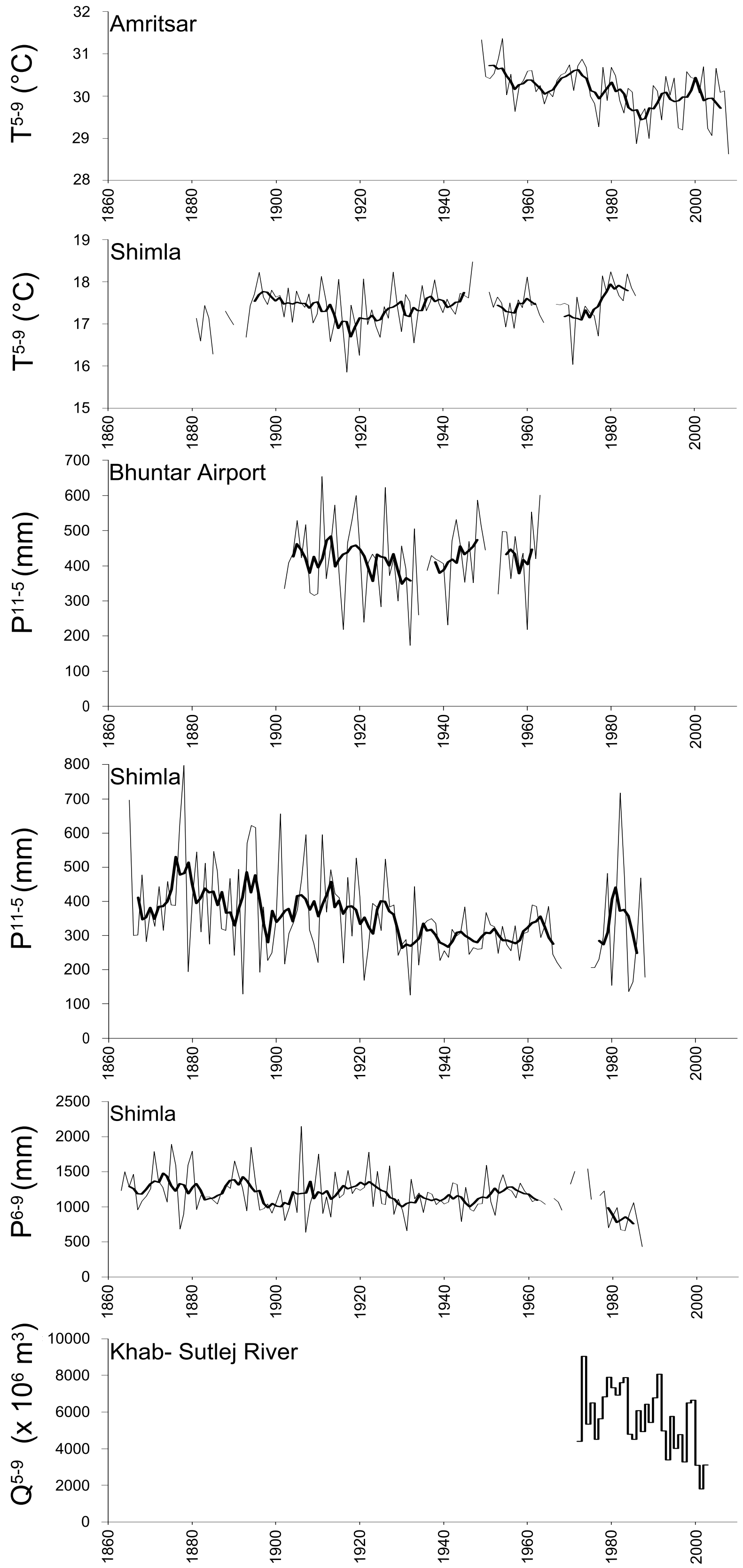

Figure 7.2.2 Glaciologically relevant meteorological variables in western Himalaya compared with summer discharge in the Sutlej River.

7.2.3 Central Himalaya catchment

River flows across the central Himalaya are heavily influenced by summer monsoon with large quantities of rainfall contributed to discharge. At Mukteshwar, summer temperatures had two cooling periods, from 1900-1920 and from 1960-1990, and two warming periods, one between 1920 and 1960 and the other after 1990 (Figure 7.2.3). Summer temperatures at Kathmandu were very similar with a declining trend until the mid-1970s, which was followed by warming temperatures since 1980. In a warming climate across central Himalaya there is increased heat energy available to melt snow and ice, but this is not the main influence on river flow changes in a monsoon-fed catchment.

Initially summer precipitation at Mukteshwar was stable with no major change up to the 1960s, but declined by around 50% between 1960 and 2009 (Figure 7.2.3). Similarly, summer precipitation at Kathmandu was reduced by 33.2% from 1951 to 1991 (Figure 7.2.3). However, monsoon rainfall at Kathmandu showed an insignificant decrease before rising from 1990 onwards (Figure 7.2.3). Data records at Mukteshwar are very limited between 1984 and 1999 because of missing data. During the same period at Kathmandu, data is also incomplete, so it is hard to determine whether summer precipitation increased or continued to fall. More stations are needed to examine summer trends in Nepal, especially at high elevation, to help identify trends across central Himalaya.

Summer precipitation drives summer runoff in the central Himalaya. Summer river flows in the Kali Kandaki River at Setibeni initially increased, from 1960 until the late 1970s, but from 1974 to 1992 discharge declined by 13% (Figure 7.2.3). In the Narayani River at Narayanghat, summer discharge had three cycles of increase and decrease. Between 1963 and 2006, summer runoff showed a small increase of 5% (Figure 7.2.3). Although summer temperatures have warmed since 1980, summer runoff has been fairly stable in the tributaries of the Ganges across Nepal because of the monsoon being steady, as at Kathmandu.

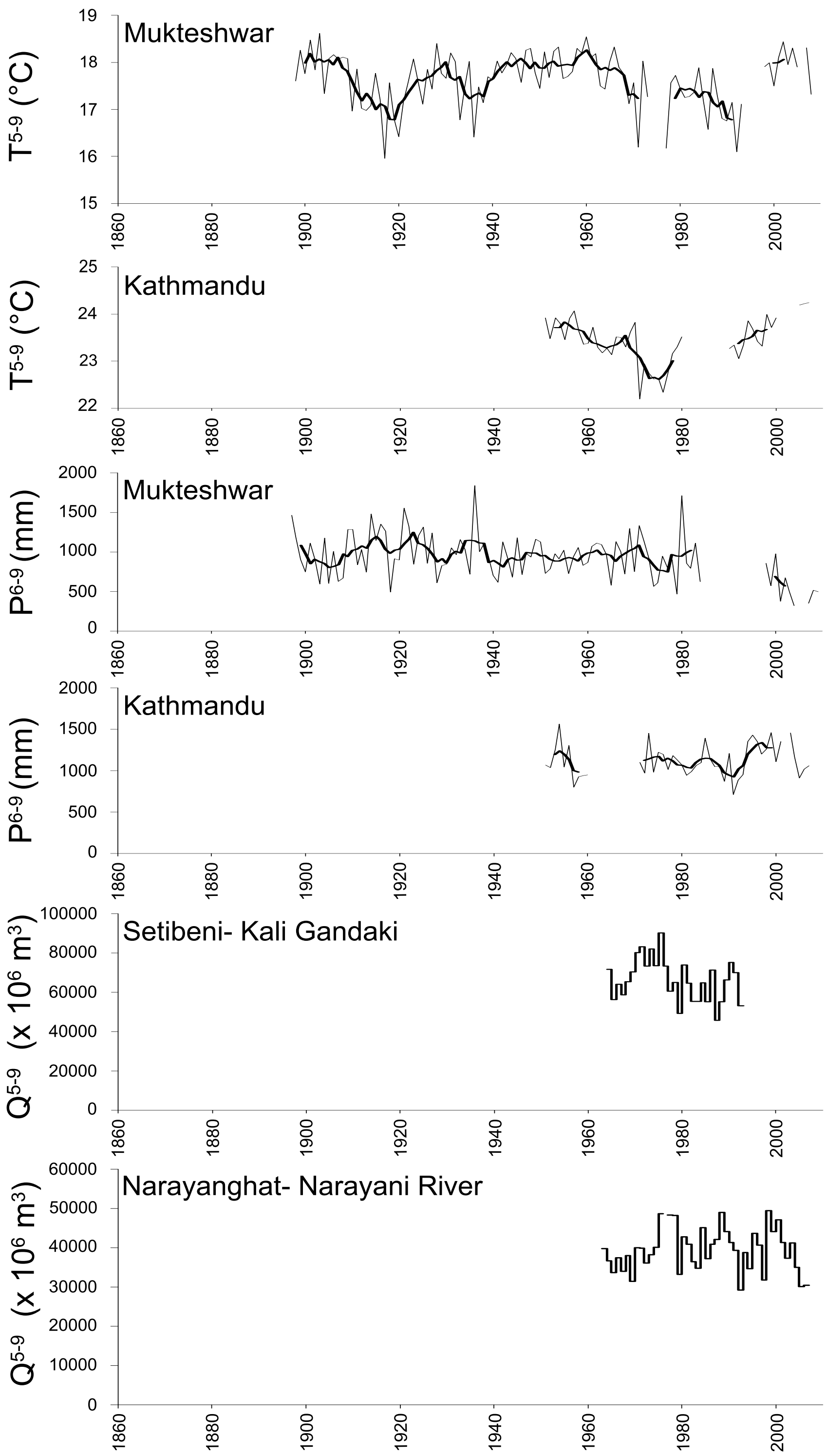

Figure 7.2.3 Glaciologically relevant meteorological variables in central Himalaya compared against summer discharge from the Ganges River catchment.

7.2.4 Eastern Himalaya catchment

Monsoon strength is at its greatest in the eastern Himalaya with total summer precipitation playing an important role in tributaries east of Nepal draining into the Ganges. Initially summer temperature decreased throughout 20th century at Darjeeling and Guwahati until about 1980 (Figure 7.2.4). Since 1980, summer temperatures at Guwahati warmed to their highest level with a 1°C increase from 1980 to 2009 (Figure 7.2.4). Darjeeling is the only high elevation station in eastern Himalaya. Summer temperatures declined by 0.5°C from the early 1880s to 1980s. However, after 1980 there are no data for Darjeeling, but as it closely followed Guwahati until 1980 it is likely that summer temperatures would have increased in parallel after 1980.

Total summer precipitation is an important indicator of runoff levels, being the main contributor to river flow during summer. Darjeeling is a high altitude station and summer precipitation declined by 40% from 1869 to 1978 (Figure 7.2.4). Summer precipitation declined 10% at Shillong from 1867 to 1999 (Figure 7.2.4). However, at low elevation Guwahati, summer precipitation has been steady over the 20th century. Summer monsoon change seems to be accelerated at higher elevation with losses enhanced with altitude. Darjeeling is located in close proximity to glacierised basins, so it would be easy to suggest that glacier runoff has been heavily affected. Downstream, most of the summer runoff is contributed by precipitation, which has changed little at Guwahati. With insufficient rates of snowfall during summer in high mountain areas in eastern Himalaya, along with low rates of winter precipitation, runoff reduced as summer precipitation declined.

Summer discharge data availability is limited, with Farakka and Hardinge Bridge the only gauging station records in the far eastern Himalaya. However, these records do not reflect discharge draining from glaciers, as water has been abstracted upstream of the gauges. Instead Setibeni and Narayanghat in Nepal have been chosen as they are upstream on tributaries of the Ganges. Flow at Setibeni declined 13% (1964-1992) but Narayanghat saw a minor increase of 5% in the same period (Figure 7.2.4).

Rivers flowing into the Ganges have not had great losses, even where summer precipitation declined by up to 40%, but precipitation monitoring stations are located a long distance away from gauging stations. It would be expected that summer river flows would have declined in the eastern Himalaya as summer precipitation decreased, just as the Sutlej River at Khab declined as monsoon rainfalls significantly declined.

7.3 Correlation Matrices

7.3.1 Correlation of summer air temperatures

North-West Himalaya has Peshawar and Lahore with long records, but only Srinagar provided an extensive data series at high altitude. Summer temperatures at Lahore and Peshawar showed a strong positive correlation of 0.64 (Table 7.1.1), as expected with both following similar trends since 1940, and both located at less than 400 m a.s.l. Summer temperatures in Gilgit and Srinagar at high elevation are fairly closely related with a positive correlation of 0.51 (Table 7.1.1), but this is only taken over a short period because the Gilgit record only ranges from 1968 to 1985. Summer temperature at Srinagar showed a strong positive correlation with that at Peshawar, but only moderately with Lahore, 0.65 and 0.34 respectively (Table 7.1.1). Leh is located north-east of Srinagar and showed a high correlation of 0.65, compared with Shimla which is much weaker at 0.21 (Table 7.1.1). This is due to Leh being much closer to Srinagar and being at a similar elevation. Summer temperature stations at similar altitudes are closely connected, but Srinagar shows a direct link also with stations situated at lower elevations. Correlations are generally higher between variables at stations in close proximity.

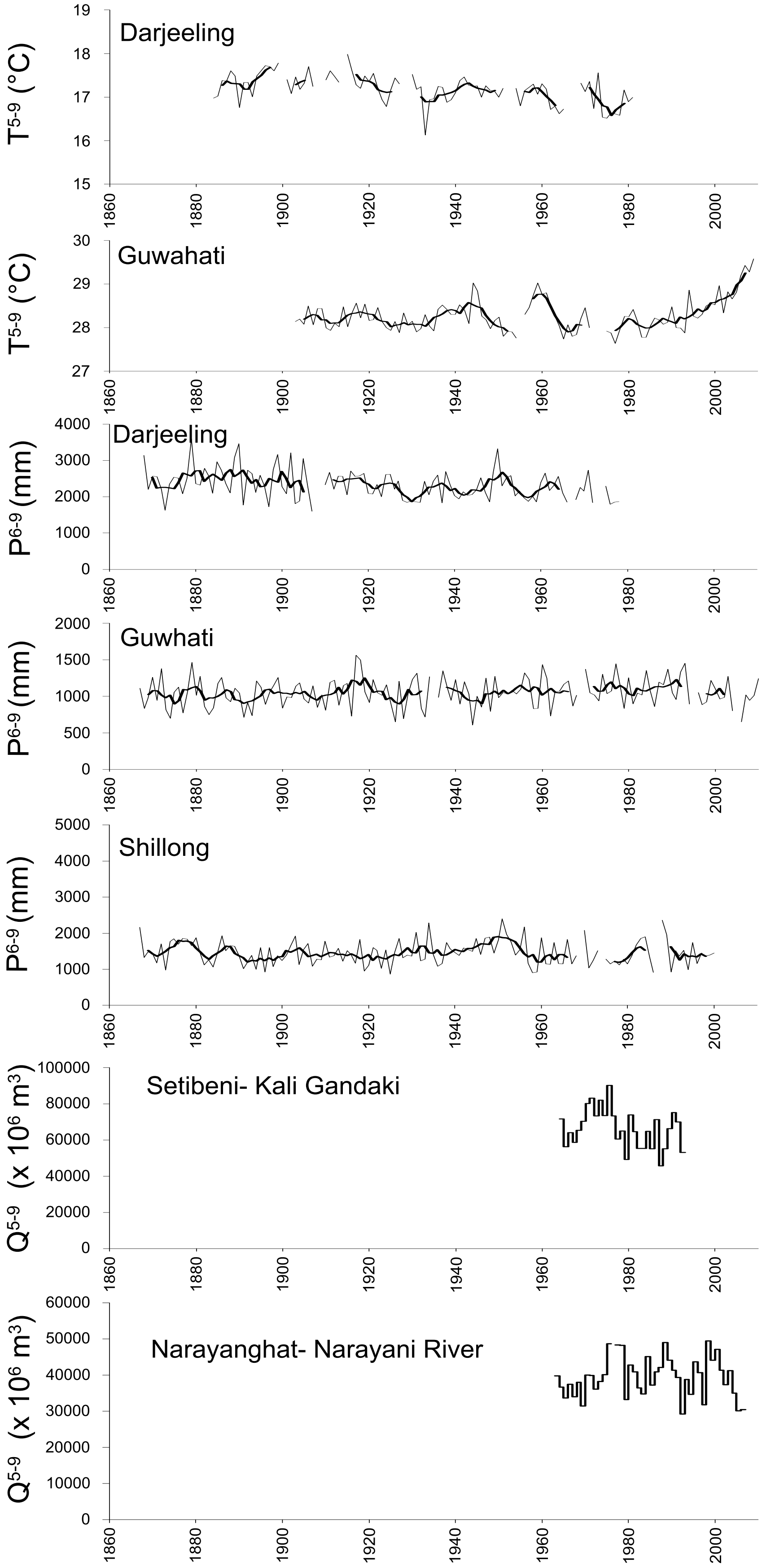

Figure 7.2.4 Glaciologically relevant meteorological variables in eastern Himalaya compared against summer discharge from the Ganges River catchment.

Table 7.1.1 Correlation matrix of summer air temperature for north-west and western Himalaya.

	Gilgit	Srinagar	Lahore	Leh	Shimla
Peshawar	-0.02	0.65	0.64	0.36	0.31
Gilgit	-	0.51	-0.17	-	-0.38
Srinagar	-	-	0.34	0.65	0.21
Lahore	-	-	-	0.09	0.73
Leh	-	-	-	-	-0.04

Amritsar in western Himalaya showed a strong positive correlation of summer temperature with Leh of 0.6, but at 0.2 much weaker with Shimla (Figure 7.1.2). Shimla and Leh have no relationship which is surprising, considering they are closely located and at high elevation. Summer temperatures in the western Himalaya are not that closely related, with only Amritsar and Leh showing a strong correlation. Summer temperature at Shimla showed a positive correlation with Mukteshwar of 0.67 (Table 7.1.2), as might be expected with Shimla being the station closest to the central Himalaya. However, Amritsar and Leh show little relation with the central Himalayan stations of Mukteshwar and Kathmandu, as they are hundreds of kilometres further west. Summer temperatures at Mukteshwar and Kathmandu in central Himalaya have a positive correlation of 0.57 (Table 7.1.2). Correlation of summer temperatures between stations in the central Himalaya showed a close correlation, as the stations follow similar trends, but Mukteshwar has a strong association with Shimla so could be classed within the same region.

Table 7.1.2 Correlation matrix of summer air temperature for western and central Himalaya.

	Shimla	Amritsar	Mukteshwar	Kathmandu
Leh	-0.04	0.60	-0.12	-0.21
Shimla	-	0.20	0.67	0.08
Amritsar	-	-	0.14	0.11
Mukteshwar	-	-	-	0.57

Tibetan Plateau is at a consistent high elevation with Lhasa and Nagqu both above 3000 m a.s.l. Summer temperature at Lhasa and Nagqu showed a strong positive correlation with 0.61 (Table 7.1.2). Nagqu does not correlate well with other stations from the Himalaya. However, summer temperature at Lhasa showed a strong positive correlation with Kathmandu and Darjeeling with 0.71 and 0.69 respectively (Table 7.1.2). Correlation of summer temperatures in Lhasa showed some strong relationships with stations from the Himalaya, but Nagqu shows no correlation making hard to determine whether there is close link between the Tibetan Plateau and stations from the Himalayan arc.

Table 7.1.3 Correlation matrix of summer air temperature for Tibetan Plateau against selected stations.

	Nagqu	Mukteshwar	Kathmandu	Darjeeling
Lhasa	0.61	0.04	0.71	0.69
Nagqu	-	0.08	0.20	0.09
Mukteshwar	-	-	0.57	0.48
Kathmandu	-	-	-	0.23

It is important to correlate Darjeeling as it is within close proximity to glaciers. Summer temperatures at Darjeeling showed negligible correlation with Cherrapunji, Guwahati and Shillong with correlation coefficients of -0.10, 0.09 and 0.26 respectively (Table 7.1.4). Stations in the lowland eastern Himalaya showed little correlation of air temperatures at Darjeeling suggesting that trends occurring near glaciers are not occurring at lower elevations. Summer temperatures at Cherrapunji showed positive correlations of 0.43 and 0.55 with Guwahati and Shillong respectively and 0.48 between Guwahati and Shillong (Table 7.1.4). Mukteshwar is the closest station to the eastern Himalaya with a long uninterrupted data set in the high mountains. Summer temperatures at Mukteshwar showed moderate to strong positive correlations with Darjeeling, Cherrapunji and Shillong of 0.48, 0.29, 0.48 respectively (Table 7.1.4). Summer air temperatures across the eastern Himalaya are poorly correlated between high and low altitude, but the relationship at lowland stations is close.

Table 7.1.4 Correlation matrix of summer air temperature for eastern Himalaya.

	Cherrapunji	Darjeeling	Guwahati	Shillong
Mukteshwar	0.29	0.48	-0.17	0.58
Cherrapunji	-	-0.10	0.43	0.55
Darjeeling	-	-	0.09	0.26
Guwahati	-	-	-	0.48

7.3.2 Correlation of winter precipitation

Winter precipitation at Peshawar showed a strong positive correlation with Srinagar of 0.6 (Table 7.2). At other high altitude stations, winter precipitation at Peshawar showed negligible correlation with Gilgit and Skardu Airport. Peshawar and Srinagar experience much more total winter precipitation than Skardu Airport and Gilgit that receive low winter snowfall. Highly elevation sites in north-west Himalaya show little connection with only Skardu Airport and Gilgit showing a weak positive correlation of winter precipitation of 0.27 (Table 7.2). Correlation between winter precipitation at Bhuntar and Shimla in western Himalaya is strong with 0.82 (Table 7.2). There is some relation between western and north-west Himalaya as Srinagar showed positive correlations of 0.37 and 0.42 with Shimla and Bhuntar Airport respectively (Table 7.2). At much drier stations of Gilgit and Skardu Airport, the correlation of winter precipitation is negligible with western Himalaya. The problem with correlating precipitation is that there are large differences in precipitation between stations which are close together. Additionally within regions in mountains precipitation is considerably variable.

Table 7.2 Correlation matrix of winter precipitation for north-west and western Himalaya.

	Skardu Airport	**Srinagar**	**Gilgit**	**Shimla**	**Bhuntar Airport**
Peshawar	0.20	0.60	0.02	0.37	0.28
Skardu Airport		-0.11	0.27	-0.01	0.05
Srinagar			0.22	0.37	0.42
Gilgit				-0.04	0.01
Shimla					0.82

7.3.3 Correlation of summer precipitation

Summer precipitation in Peshawar and Skardu are only weakly correlated (r = 0.19). (Table 7.3.1). Summer precipitation at these stations show no relation with Shimla and Mukteshwar located in western and central Himalaya respectively. Summer precipitation at Shimla and Mukteshwar showed moderate correlation with 0.44

(Table 7.3.1). Correlation of summer precipitation between north-west and western Himalaya is weak, but at stations affected by summer monsoon there is a stronger connection at Shimla and Mukteshwar.

Whilst only two stations were selected for summer precipitation in Nepal for examining long term trends, extra stations at higher elevations were needed to investigate spatial associations. Additional sites (Table 7.3.1) were located at elevations between 1835 and 3970 m a.s.l., but records exist only for the period between 1970 and 1990. For short data series, correlation with overlapping portions of long uninterrupted series was undertaken. Summer precipitation in Mukteshwar was relatively high from 1950 to 1984 before declining from 1984 onwards (Figure 7.3). Summer precipitation at Mukteshwar was poorly correlated with Kathmandu and Bijayapur, but showed some correlation with Lete and Khumjung with correlation coefficients of 0.29 and 0.42 respectively (Table 7.3.1). Mukteshwar is the most complete precipitation data set available for central Himalaya, but only correlates with precipitation at two other stations and not strongly enough to suggest that the summer precipitation trends at Mukteshwar are matched high elevation stations across Nepal. Kathmandu differs from Mukteshwar as summer precipitation declined initially up to 1990, followed by a wetter period between 1990 and 2010. Summer precipitation at Kathmandu showed little correlation with the other stations which may be due to regional differences in meteorological conditions. Both Lete and Khumjung showed similar temporal patterns of summer precipitation increasing after 1980 but Bijayapur was fairly stable from 1970 to 1990 (Figure 7.3). Khumjung is the highest station at 3970 m a.s.l., but showed limited correlation with the other high altitude sites of Lete and Bijayapur. Summer precipitation in the central Himalaya showed some moderate inter-station correlation, but many stations were poorly related because of local differences making it difficult to assess regional trends.

Table 7.3.1 Correlation matrix of summer precipitation for north-west, western and central Himalayan catchments.

	Skardu Airport	Shimla	Mukteshwar	Kathmandu	Bijayapur	Lete	Khumjung
Peshawar	0.19	0.11	0.05	0.30	0.09	0.28	-0.38
Skardu Airport		-0.04	0.28	0.29	0.51	0.15	0.01
Shimla			0.44	0.29	0.31	-0.17	0.27
Mukteshwar				-0.06	0.17	0.29	0.42
Kathmandu					0.02	0.19	0.25
Bijayapur						0.62	0.20
Lete							0.24

Summer precipitation at Shillong in the lowlands showed a strong positive correlation with that at upland Darjeeling and Cherrapunji with correlation coefficients of 0.62 and 0.59 respectively (Table 7.4.1). However, summer precipitation at Darjeeling and Cherrapunji was poorly correlated which may be due to Cherrapunji being located further east where average summer precipitation was more than double that at Darjeeling. Cherrapunji has much missing data after 1950 meaning correlation of summer precipitation is possible over only a short period. Summer precipitation at Lhasa in the Tibetan Plateau showed little relation with all three stations in east Himalaya (Table 7.4.1). This is expected because the Tibetan Plateau is a dry region to which rain-bearing monsoon winds that affect the eastern Himalaya penetrate only infrequently. Summer precipitation in the Central Himalaya is poorly correlated with east Himalayan stations with a very strong negative correlation of -0.95 between Darjeeling and Khumjung (Table 7.4.1). Correlation of summer precipitation at stations in the eastern Himalaya was poor, though stronger at higher elevations. But there is negligible correlation with the Tibetan Plateau and central Himalaya.

Table 7.3.2 Correlation matrix of summer precipitation for eastern Himalaya compared against Tibetan Plateau and central catchments.

	Darjeeling	Shillong	Lhasa	Khumjung	Mukteshwar
Cherrapunji	0.12	0.59	0.24	0.14	-0.13
Darjeeling	-	0.62	-0.07	-0.95	0.29
Shillong	-	-	0.08	-0.27	-0.11
Lhasa	-	-	-	0.58	-0.06
Khumjung	-	-	-	-	0.42

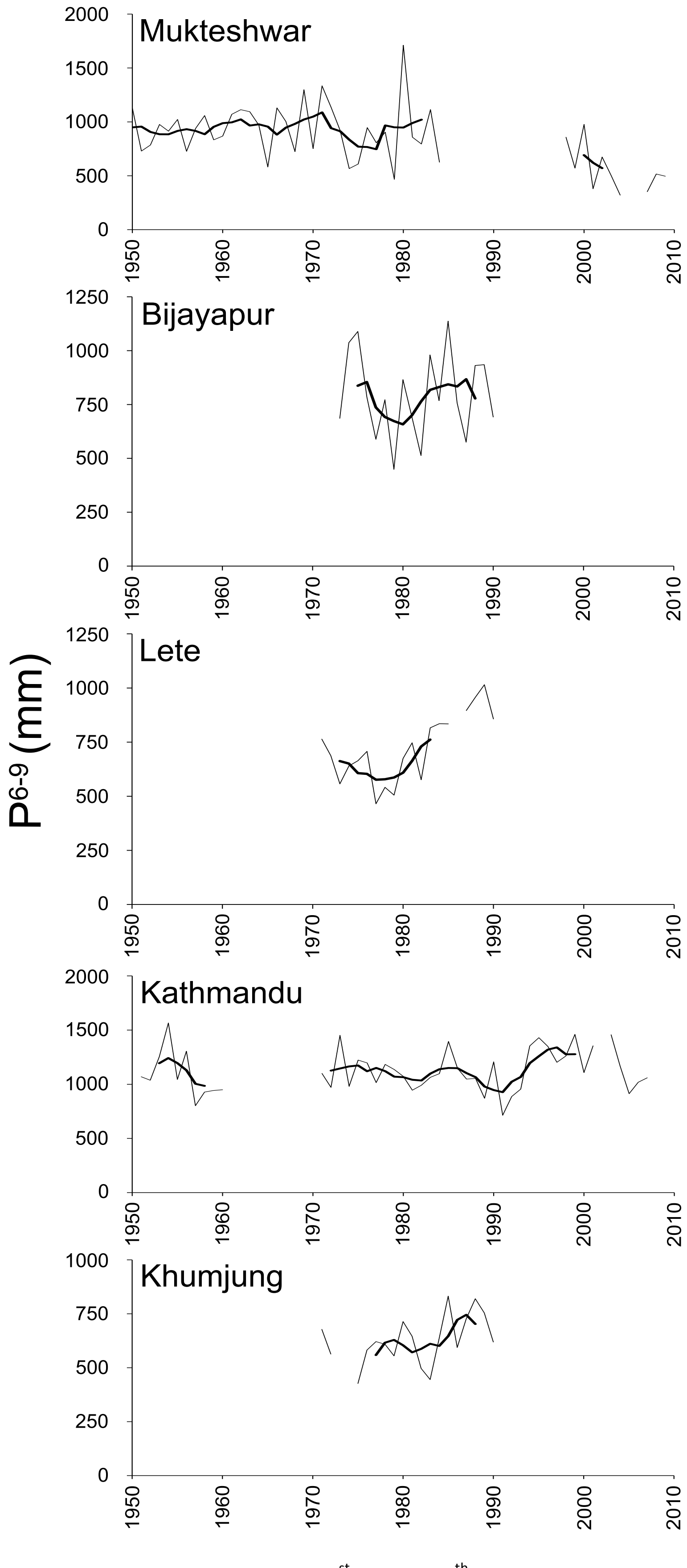

Figure 7.3 Summer precipitation (1st June to 30th September) across central Himalaya between 1950 and 2010 (Thin line- real data, Thick line- moving average).

7.3.4 Correlation of Annual Discharge

Annual discharge at Gilgit and Dainyor Bridge are moderately correlated with r= 0.34 (Table 7.4.1). This is very low considering their proximity to each other with the gauging stations located within 10 kilometres of each other. Annual runoff in The Shyok River at Yugo station relates much more strongly with Gilgit/Gilgit and Hunza/Dainyor Bridge with r=0.42 and 0.66 respectively (Table 7.4.1). To the west, annual discharge of the Sutlej River at Khab was weakly positively related with Shyok/Yugo and Gilgit/Gilgit, but strongly with Hunza/Dainyor Bridge (Table 7.4.1). Correlation of annual discharge amongst stations in the upper Indus shows some strong positive relationships suggesting that similar climatic conditions influence all. By contrast, correlation between the Sutlej and the Indus rivers is mostly poor.

Table 7.4.1 Correlation matrix for Indus and Sutlej River catchments.

	Dainyor Bridge	Yugo	Khab
Gilgit	0.34	0.42	0.2
Dainyor Bridge		0.66	0.51
Yugo			0.27

Annual discharge in the Sutlej River at Khab however showed a strong positive correlation with Chisapani, Asara Ghat and Langtang with r = 0.59, 0.63 and 0.42 respectively (Table 7.4.2). Sutlej River in western Himalaya correlates well with a few river stations in Nepal because Khab is in a region influenced by summer monsoon. Discharge at Asara Ghat strongly correlates with all other stations in Nepal especially with Chisapani and Langtang with 0.95 and 0.87 respectively (Table 7.4.2). As Asara Ghat is closely connected to all Nepalese stations, suggesting that river flow trends occurring at this station would be the general pattern across central Himalaya. Chisapani also showed a strong positive correlation of annual discharge with Setibeni, Narayanghat and Asara Ghat (Table 7.4.2). Most stations in Nepal are linked closely together especially Asara Ghat and Chisapani. The station at Langtang is situated in close proximity to glaciers and flow is moderately correlated with that at Chisapani and Narayanghat, but is much more strongly related to discharge at Asara Ghat (Table 7.4.2). However, Langtang has only a short data set making correlation with annual runoff at lower altitudes less convincing based on a small subset of the data. With the exception of one negative relationship, correlation of annual discharge between tributaries of the Ganges across Nepal suggests that runoff is subject to similar monsoonal controls across the central Himalaya.

Table 7.4.2 Correlation matrix for discharge at stations in the Ganges River catchment and at Khab on the Sutlej.

	Chisapani	Setibeni	Asara Ghat	Narayanghat	Langtang
Khab	0.59	0.18	0.6	-0.04	0.42
Chisapani	-	0.58	0.95	0.64	0.3
Setibeni	-	-	0.6	0.37	-0.43
Asara Ghat	-	-	-	0.53	0.87
Narayanghat	-	-	-	-	0.34

7.4 Is climate change accelerated with elevation?

7.4.1 Is climatic warming enhanced at altitude?

Beniston *et al.* (1997) highlighted that climate records have shown that temperature changes over last century have been amplified at higher elevation. However has this occurred at air temperature and precipitation stations selected in this study? In north-west Himalaya temperatures during summer at Srinagar showed an increase of 1.17°C at the highest station in this region compared to 0.98°C at Lahore at 216 m a.s.l. (Table 7.5.1). However, summer temperatures at Peshawar experienced a decrease of 1.14°C which is situated in the lowland areas at 359 metres. Across north-west Himalaya summer temperatures both increased and decreased irrespective of elevation. At Shimla (at 2143 m a.s.l.) summer temperatures increased by 0.53°C, whereas at the highest station, Leh, temperatures in summer decreased by 0.87°C. Cooling at lowland Amritsar was 1.21°C (Table 7.5.1). Observations in western Himalaya suggest that summer temperatures did not experience enhanced warming at high elevation. A small decrease was recorded at Shimla. Summer temperatures in the central Himalaya at Mukteshwar (2289 m a.s.l.)

increased by 0.51°C and those at Kathmandu (1302 m) by 0.3°C (Table 7.5.1). Summer temperature trends in central Himalaya seem to be enhanced with increased elevation, but more observations at high elevation are needed to support such a view.

Observed warming on the Tibetan Plateau at Lhasa (3656 m a.s.l.) and at Nagqu (4513 m) includes summer temperatures increasing by 0.67°C at Lhasa and by 1.41°C at Nagqu (Table 7.5.1). In the eastern Himalaya, summer temperatures at Guwahati increased 1.44°C at 115 m a.s.l. (Table 7.5.1). At Cherrapunji and Shillong summer temperatures only increased by 0.45 and 0.25°C at higher elevations - 1489 and 1438 m a.s.l respectively. The highest station in the eastern Himalaya is Darjeeling, which experienced a 0.29°C decline in summer temperatures (Table 7.5.1). However, this data set only runs until 1981 and since 1980 summer temperatures have warmed, so it might be expected that Darjeeling would have followed this trend. The Eastern Himalaya shows little evidence of enhanced warming with elevation as the greatest increases in summer temperature are in the lowlands.

7.4.2 Is winter precipitation change enhanced at high elevation?

Winter precipitation in Peshawar (359 m a.s.l.) changed little over last century with a minor increase of only 20.8 mm (Table 7.5.2). Winter precipitation at Srinagar and Gilgit at around 1500 m a.s.l. showed little change with small decrease and increase respectively. Winter precipitation at Skardu Airport increased by 226.3 mm, which is a substantial change in a dry area (Table 7.5.2). Winter precipitation increase with elevation is not proven as only Skardu changed whilst winter precipitation was stable at Peshawar, Srinagar and Gilgit.

Table 7.5.1 Summer air temperature change with relation to elevation.

Station	Elevation	Region	Series Length	Temperature change (°C)	Temperature change per year (°C a^{-1})
Nagqu	4513	Tibetan Plateau	1955-1995	1.41	0.035
Lhasa	3656	Tibetan Plateau	1935-2007	0.67	0.009
Leh	3411	Western	1881-1987	-0.87	-0.008
Mukteshwar	2289	Central	1898-2010	0.51	0.005
Shimla	2143	Western	1883-1960	0.53	0.007
Darjeeling	2128	Eastern	1882-1981	-0.29	0.003
Srinagar	1590	North-west	1895-2010	1.17	0.01
Cherrapunji	1489	Eastern	1903-2011	0.45	0.004
Shillong	1439	Eastern	1903-2000	0.25	0.002
Kathmandu	1302	Central	1951-2007	0.32	0.006
Peshawar	359	North-west	1931-2010	-1.14	0.014
Amritsar	232	Western	1949-2007	-1.21	0.02
Lahore	216	North-west	1876-2009	0.98	0.007
Guwahati	51	Eastern	1903-2009	1.44	0.013

Shimla experienced a substantial decrease in winter precipitation of 517.8 mm (Table 7.5.2). However, Shimla in the western Himalaya is best compared with stations from in the north-west region. Winter precipitation was enhanced at stations above 2000 m throughout north-western and western Himalaya.

Table 7.5.2 Winter Precipitation change in relation to elevation.

Station	Elevation	Region	Series Length	Precipitaiton change (mm)	Precipitation change per year (mm a^{-1})
Skardu Airport	2222	North-west	1901-1999	226.3	2.31
Shimla	2143	Western	1865-1987	-517.8	4.24
Srinagar	1590	North-west	1894-2009	-43	0.37
Gilgit	1465	North-west	1894-1999	36.4	0.35
Peshawar	359	North-west	1863-2000	20.8	0.15

7.4.3 Is summer monsoon change enhanced at higher altitude?

Summer precipitation at Shimla and Mukteshwar was reduced by 802.6 and 971 mm respectively. Both are located above 2,000 m a.s.l. (Table 7.5.3). At a lower elevation, summer precipitation at Kathmandu in central Himalaya only went down by 9 mm (Table 7.5.3). Precipitation during summer reduced with elevation in western and west-central Himalaya, but did not change at Kathmandu in central

Himalaya. At the highest available station on the Tibetan plateau, Lhasa, summer precipitation increased by 61.1 mm (Table 7.5.3), compared with substantial losses in the western and west-central Himalaya.

Above 1400 m a.s.l., summer precipitation at Shillong in the eastern Himalaya declined by 710.4 mm (Table 7.5.3). Summer precipitation in Darjeeling at 2128 m a.s.l. and located closer to glaciers went down by 670 mm (Table 7.5.3). Summer precipitation at Cherrapunji showed a small decline of 92.22 mm (Table 7.5.3), but Guwahati in the lowlands gained 132.4 mm (Table 7.5.3), completely different behaviour by comparison with the other stations in the eastern Himalaya. Summer precipitation thus appears to have increased considerably at high by comparison with low elevation sites.

Table 7.5.3 Summer precipitation changes with relation to elevation.

Station	Elevation	Region	Series Length	Precipitaiton change (mm)	Precipitation change per year (mm a^{-1})
Lhasa	3656	Tibetan Plateau	1935-2003	61.1	0.78
Mukteshwar	2289	Central	1897-2009	-971	8.67
Shimla	2143	Western	1863-1987	-802.6	6.47
Darjeeling	2128	Eastern	1901-1970	-670	9.57
Cherrapunji	1489	Eastern	1872-2010	-92.22	0.65
Shillong	1439	Eastern	1867-2000	-710.4	5.34
Kathmandu	1302	Central	1951-2007	-9	0.16
Guwahati	51	Eastern	1867-2010	132.5	0.92

7.5 Summary of results

Annual and summer temperatures across the north-west Himalaya warmed between 1900 and 1950 followed by a dip until around 1985 as global solar radiation dimmed from 1950 to 1980 (Huss *et al*., 2009). Temperatures increased since 1985, just exceeding levels experienced around 1950. Bhutiyani *et al*., (2007) also showed that the highest rate of warming through the 20th century in the Himalaya was since 1980 in the north-west. In the western Himalaya summer temperatures cooled after 1950, with warming picking up from the 1980s, large increases in temperatures from

1984 and 2007 also being detected by Shekhar *et al*., (2010). In the central Himalaya, annual and summer temperatures increased since 1980. Shrestha *et al*. (1999) had observed rising temperatures from mid-1960s to mid-1990s across Nepal. As temperatures warm, glacier melt is increases but in a monsoon fed catchment total summer precipitation masks any influence of temperature on river flow. As snow and ice melt are enhanced with increased available heat energy in a warming climate across central Himalaya, monsoon influence will increase further as runoff from the increasingly large ice-free portions of basins is directly related to precipitation.

Annual and summer temperatures on the Tibetan plateau warmed dramatically after the cooling period between 1935 and 1960, as reported by Liu and Chen (2000) who identified an annual increase of 0.16°C between 1950 and 1990. Significant increases in temperatures on the plateau suggest warming seems to be enhanced with elevation. Summer temperatures cooled between 1945 and 1990 in the eastern Himalaya, but then temperatures reached levels not previously reached in the early 21st century. Jharharia and Singh (2010) showed rising temperatures during the summer monsoon period between 1965 and 2000. Annual temperatures increased throughout the 20th century with two warming periods between 1900-1960 and 1990-2010. Immerzeel (2008) suggested year on year increases in temperatures in the Brahmaputra basin from 1900 to 2000. Temperatures in the eastern Himalaya have increased from 1900 to 2010 with the most significant warming occurring since 1980. However the cool period varies in length and timing within the region. Increasing temperatures will reduce glacier area in the eastern Himalaya, leading to reduced contributions from snow- and ice-melt in a region that already produces limited glacier runoff.

Across the Himalaya temperatures generally warmed through the 20th century until 1950, followed by cooling from 1950 to 1980, but picked up again after 1980-85 as temperatures reached just above earlier maxima. Across the Himalaya runoff initially changes because of enhanced melting of snow and ice as temperatures rise (Immerzeel, 2008; Jansson *et al*., 2003). However, glaciers will decline and ultimately cease existence altogether, and runoff will solely reflect future levels of precipitation as the deglaciation discharge dividend declines (Collins, 2008). In the eastern

Himalaya with a lower percentage of glacierisation, total monsoon precipitation already dominates river flow.

Winter precipitation across the north-west Himalaya was generally stable over the 20th century, but declined by 18% at Srinagar from 1969 to 2009. Bhutiyani *et al*., (2010) interestingly reported an increase in winter precipitation from 1866 to 2006. The Western Himalaya has been becoming drier throughout, especially at Shimla where winter precipitation reduced by 41% between 1878 to 1987. Shekhar *et al*. (2010) also observed a decline in winter precipitation. As winter westerlies influence weakens, with winter precipitation either stable or drying across north and western Himalaya, there has been insufficient total winter precipitation to offset reduced summer ice-melt.

Summer monsoon rainfall at Shimla reduced by 50% between 1960 and the mid-1980s, supporting the view that total summer precipitation is declining (Bhutiyani et al., 2010). Shimla is in marginal region in which summer monsoon just dominates runoff contributions from snow- and ice-melt. Summer precipitation at Mukteshwar in west central Himalaya similarly decreased by 48% (1910-2002) as also indicated by Basistha et al., 2009). Significant fall in monsoon precipitation over the lower Sutlej and north-western tributaries of the Ganges, has led to reduced river flows. In contrast, in Nepal, in the central Himalaya, summer precipitation was steady since the 1950s with small increases since 1980s.

In the eastern Himalaya, summer precipitation at Cherrapunji and Guwahati showed no clear wetting or drying trends over the 20th century. Immerzeel (2008) considered highlighted that precipitation showed no distinct trends. However, summer precipitation lost around 40% and 10% at Darjeeling (1869-1978) and Shillong (1867-1999) respectively. Summer precipitation has generally reduced considerably in the western and eastern regions, leading to lower levels of runoff in catchments dominated by monsoon. However, runoff in Nepal has remained unchanged with fairly stable monsoon. Future flows will reflect year-to-year variations of precipitation especially in less glacierised catchments.

Annual runoff in Hunza River at Dainyor Bridge in the upper Indus basin reduced by nearly 24% between 1967 and 2004 in parallel with the decline in flow of the Ravi and Beas tributaries of the Indus (Bhutiyani *et al*., 2008). However, annual discharge has generally fluctuated without trend in the Gilgit and Shyok rivers so that river flow is fairly stable in the main-stem upper Indus. Annual discharge in the Sutlej River at Khab declined by 31% from 1972 to 2002. The monsoon extends sufficiently west to the Sutlej basin to start to influence runoff (Bhutiyani *et al*., 2008). Flow in the main-stem Indus is derived from basins with differing climatic and hydrological conditions. Flows at both Khab and Dainyor Bridge declined but for differing reasons.

Annual river flow dipped in the mid-1960s in the tributaries of the Ganges in Nepal, but recovered until in the early 1970s. Between 1960s and 1990s annual and summer runoff observe no significant losses or gains. Only the Kali Gandaki River at Setibeni lost 12% and 13% of annual and summer discharge respectively from 1964 to 1993 indicative of declining annual and monsoon precipitation in Nepal (Sharma and Shakya, 2006). Annual and summer runoff in the tributaries of the Ganges across Nepal were without major trends, and the reduction in flow at Setibeni (12%) was relatively minor compared to the 24 and 31% decreases in the Hunza and Sutlej Rivers respectively.

River flow from the west to east Himalaya is dominated by summer monsoon, with only the north-west, the Karakoram, receiving a greater contribution from melting snow and ice. Future runoff in monsoon-fed catchments will reflect total precipitation especially where levels of glacierisation are low. In the highly glacierised Karakoram, future river flow may rise if summer temperatures continue to warm, but will eventually be greatly reduced as glacierised areas decline.

Discharge records are available in the central Himalaya but not in the eastern Himalaya. With summer precipitation declining by 40% and 10% at Darjeeling and Shillong, annual discharge will probably have declined in the east, as for the Sutlej.

8. Conclusion

Climate warmed through the 20th century along the Himalayan arc, as indicated also in other studies which showed rising temperatures across the various catchments (Bhutiyani *et al*., 2007; Shrestha *et al*., 1999; Immerzeel, 2008). Annual and summer temperatures initially warmed in the first half of the twentieth century until around 1950, but cooled between 1950 and 1980-1990 as expected with the global dip in solar radiation (Huss *et al*., 2009). Most recently, temperatures reached their highest levels, but only just exceeding the earlier maxima of the 1950s. However, different regions vary in the timing of temperature change. Temperatures increased earlier from 1960 on the Tibetan Plateau, whilst warming in the eastern Himalaya was lagged until 1990. Summer temperatures are strongly correlated especially at stations at similar elevations, with timings of warm and cool periods closely related. With high mountains being particularly sensitive physical environments (Beniston, 2010), temperatures have shown greater warming at higher elevation stations from north-west to central Himalaya and Tibetan Plateau. At high altitude, climatic warming occurred sooner, increasing energy available for melting snow and ice.

Winter precipitation in north-west Himalaya has generally been stable, but declined by 19% from the 1970s at Srinagar. At Shimla winter precipitation declined by 40% between 1880 and 1987. In a warming climate, winter precipitation as snow has to be in high abundance to offset melting of snow and ice and maintain glacier dimensions. As remote sensing techniques have showed, glaciers lost mass between 1968 and 2006 in the north-west upper Ganges basin (Bhambri *et al*., 2011). However, it is difficult to identify the degree of change in glacier loss and the effect it is having on river flow with variations in meteorological conditions. With recent rising temperatures in north-west Himalaya, future river flows will reflect temperature variations and the amount of winter precipitation to melt. If warm summers continue to follow insufficient total winter precipitation, discharge will ultimately reduce.

During the 20th century summer monsoon precipitation significantly declined in the western Himalaya with around 50% loss around Sutlej and north-west Ganges river basins. Basistha *et al*., (2009) also observed drying trends in Uttarakhand. However, summer monsoon in the central Himalaya was stable since the 1950s with small increases since the 1980s. In the east, summer precipitation declined 40% at Darjeeling and 10% at Shillong from the late 1880s to 1990s. The Himalayan arc experienced drying trends in the western and eastern regions, but Nepal in the central Himalaya, the monsoon was steady since the 1950s. Inter-correlation of precipitation is weak with large differences between stations that are located in both dry and wet areas, probably as a result of small-scale variability. As ice free areas increase in warming monsoon-fed catchments, discharge will progressively reflect year-to-year variations of total summer precipitation. River flows will be reduced if summer monsoon continues to weaken, especially in western and eastern regions were precipitation has declined.

Discharge has been stable in the upper Indus since 1960s. However, flow in the Hunza at Dainyor Bridge declined 24% from 1960s to 1990s, whilst upper Sutlej discharge reduced 30% between 1970s and 2000s. The Indus region warmed since 1980, but runoff levels did not reflect the rise in temperature. Precipitation was insufficient to maintain river flows. Annual runoff across the central Himalaya has been stable between the 1960s and 2000s, with only discharge in the Kali Gandaki declining by 12%. As temperatures warm and precipitation declines, runoff cannot be sustained at a consistently high flow as glacier areas reduce. As climate change records are not entirely conclusive, more studies are needed to examine how meteorological conditions influence present and may modify future river flows. Warming temperatures have failed to increase snow and ice melt to offset runoff contributions as precipitation levels reduced, so that river flows declined due to diminishing precipitation amounts. Precipitation drives runoff across much of the Himalaya with the monsoon dominating wide expanses of the arc. Contributions from snow and ice to total discharge are failing to offset declining components from reducing summer precipitation.

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