QUANTIFIED ASSESSMENT OF THE METEOROLOGICAL VARIABLES FACILITATING THE ESTABLISHMENT OF THE KARAKORAM ANOMALY

by

Furrukh Bashir

____________________________
Copyright © Furrukh Bashir 2016

A Thesis Submitted to the Faculty of the

DEPARTMENT OF HYDROLOGY AND ATMOSPHERIC SCIENCES

In Partial Fulfillment of the Requirements

For the Degree of

MASTER OF SCIENCE

With a major in

HYDROMETEOROLOGY

In the Graduate College

THE UNIVERSITY OF ARIZONA

2016
STATEMENT BY AUTHOR

The thesis titled QUANTIFIED ASSESSMENT OF THE METEOROLOGICAL VARIABLES FACILITATING THE ESTABLISHMENT OF THE KARAKORAM ANOMALY prepared by Furrukh Bashir has been submitted in partial fulfillment of requirements for a master’s degree at the University of Arizona and is deposited in the University Library to be made available to borrowers under the rules of the Library.

Brief quotations from this thesis are allowable without special permission, provided that an accurate acknowledgment of the source is made. Requests for permission for extended quotation from or reproduction of this manuscript in whole or in part may be granted by the head of the major department or the Dean of the Graduate College when in his or her judgment the proposed use of the material is in the interests of scholarship. In all other instances, however, permission must be obtained from the author.

SIGNED: FURRUKH BASHIR

APPROVAL BY THESIS DIRECTOR

This thesis has been approved on the date shown below:

October 19th, 2016

Hoshin V. Gupta
Professor of
Hydrology and Atmospheric Sciences
Acknowledgements

I gratefully acknowledge the efforts made by Senator James William Fulbright that culminated as establishment of an international exchange program that allowed me to pursue education and research at Department of Hydrology and Atmospheric Sciences, University of Arizona, that would allow me to perform as a better scientist in the field of hydrometeorology, back at Pakistan and will benefit the public servicing sector as well.

I acknowledge the efforts made by United States Education Foundation Pakistan (USEFP), and its head Rita Akhtar, who are trying to educate the youth and building the capacity of the professional tirelessly for past several years.

I am grateful to Professor Hoshin V. Gupta, Professor Xubin Zeng, Professor Francina Dominguez, Professor Aveline F. Arellano, Professor Paul Sheppard, Professor James Shuttleworth and Professor Eric A. Betterton, who spent a lot of their personal time with me to teach me how to do research and kept me motivated and never gave up on me.

I would say thank you to my fellow students and to administrative and supporting staff of Department of Hydrology and Atmospheric Sciences, especially, to Sandy Holford, Lupe Romero, Sarah Warren, Mike Eklund and Matthew Jones for caring attitude and continuous support.

I would say special thanks to Sarah McCormink, Stephanie Sasz and Brian Diffley at Institute of International Education (IIE), and to Joanne Lagasse-Long at International Student Services who were always available to me in time of need.
Contents

Acknowledgements .................................................................................................................. 3
List of Figures .......................................................................................................................... 5
List of Tables ........................................................................................................................... 7
Abstract .................................................................................................................................. 9
  1. Introduction .......................................................................................................................... 11
  2. Material and Methods ........................................................................................................... 16
    I. Study Area ............................................................................................................................ 16
    II. Description of Variables ....................................................................................................... 16
    III. Trend Analysis .................................................................................................................... 21
    IV. Regional Mean Trend .......................................................................................................... 23
    V. Standardized Anomalies ....................................................................................................... 23
    VI. Upper Atmosphere Data ..................................................................................................... 24
    VII. Gridded and Reanalysis Data ............................................................................................ 24
  3. Results and Discussion ........................................................................................................... 26
  4. Conclusions ........................................................................................................................... 37

References .................................................................................................................................. 42
Figures ....................................................................................................................................... 46
Tables ........................................................................................................................................ 53
Appendix .................................................................................................................................... 58
List of Figures

Figure 1: Location of Study Area. .................................................................46

Figure 2: Standardized Anomalies of mean Maximum, Minimum, Mean Temperature (°C), Diurnal Temperature Range (°C), Dry Bulb Temperature (°C), Cloud Cover (Okta), Saturated Vapor Pressure (kPa) and Wind Speed (ms⁻¹) in morning (08 AM Local time) and afternoon (05 PM local time) of Gilgit Station for Summer Season from 1961-2011 with respect to mean and variance of the base period of 1961-1990. .................................................................47

Figure 3: Standardized Anomalies of mean Net-Radiation (MJm⁻²d⁻¹), seasonal total precipitation (mm), frequency of rainy days (#), Surface Pressure (kPa), Reference Evaporation (mm), and Climate Moisture Index (%) of Gilgit Station for Summer Season from 1961-2011 with respect to mean and variance of the base period of 1961-1990. ........................................48

Figure 4: Monthly Mean Vertical Temperature of the Atmosphere up to 6 Km reported by Radiosonde Launched at Kabul, Afghanistan in the Year 1983 and 2007 at 00-UTC (0430 AM, Local Time). The year 1983 and the year 2007 are selected as representatives of the decades of the 1980s and the 2000s, respectively, as there are no missing values in those years at 00-UTC. .................................................................................................................................49

Figure 5: Monthly Mean Vertical Temperature of the Atmosphere up to 6 Km reported by Radiosonde Launched at Kabul, Afghanistan in the Year 1980 and the 2009 at 12-UTC (0430 PM, Local Time). The year 1980 and the year 2009 are selected as representatives of the decades of the 1980s and the 2000s, respectively, as there are no missing values in those years at 12-UTC. .................................................................................................................................50
Figure 6: Comparison of summer seasonal trends of monthly mean maximum, minimum, and mean temperature and diurnal temperature range from 1961 to 2011 estimated by using in-situ instrumental meteorological data (circles) and CRU gridded data (contours).

Figure 7: Comparison of summer seasonal trends of monthly mean maximum, minimum, and mean temperature and diurnal temperature range from 1981 to 2011 estimated by using in-situ instrumental meteorological data (circles) and MERRA-2 reanalysis (contours).
List of Tables

Table 1: List of meteorological sites used in the study with their latitude, longitude, elevation and description of the data used in the study. .................................................................53

Table 2: List of observed and derived meteorological variables utilized to perform analyses. .........................................................................................................................54

Table 3: Comparison of mean seasonal and annual trends in temperatures from 1961 to 1999 as reported by [Fowler and Archer, 2006, p.4280] in their (Table 3) to the same from 1961 to 2011 with updated dataset computed from Dir, Drosh, Bunji, Gilgit, Astore, and Skardu. Positive trends are denoted with red color and negative trends are denoted with blue color. The increase in trend is indicated by the red up arrow and a decrease in trend is indicated by the blue down arrow. ........................................................................55

Table 4: Mean regional trends of different meteorological variables of HKH from 1961 to 2011 with a range of mean absolute difference are presented. Red colored box indicates absolute positive trend of the meteorological variable throughout the region, blue colored box indicates absolute negative trend of the meteorological variable throughout the region, and light red colored box indicates mean positive trend with some signal of negative trend in it and light blue colored box indicate negative trend with some signal of positive trend in it. Boxes with white background indicate mean neutral trend with both positive and negative signals.................................................................56

Table 5: Site-Elevation mean seasonal lapse rates of Maximum, Minimum and Mean Temperature and Diurnal Temperature Range along with temperature lapse rate estimated through radiosonde launched at Kabul Airport from 1966 to 2016 (decade of the 1990s and
some years from early 2000s are missing in this data) in both morning (0430 AM local
time) and afternoon (0430 PM local time). Temperature lapse rate is estimated for the
whole elevation range covered by the radiosonde, up to 15 km and up to 6 km to
differentiate the change in temperature in different layers of the atmosphere. Trends
significance is computed at 95% confidence interval and statistically significant trends are
presented with bold letters...
Abstract

Lofty Hindukush, Karakoram and Himalayan (HKH) mountain ranges centered in the Northern Pakistan are host to some of the world’s largest glaciers outside the Polar Regions and are a source of water for drinking and irrigation to the millions of people living downstream. With the increase in the global temperatures, glaciers are reported as retreating globally. However, some of the glaciers in the Karakoram mountain ranges are reported as surging with positive mass balance, especially since the 1990s. This phenomenon is described as “The Karakoram Anomaly”. Various efforts have been made to explain the state and fate of the HKH glaciers in the recent past. However, they are limited to quantification of the change in temperature, precipitation and river runoff, or through their impact on future climate projections. For the HKH region, temperature fluctuations have been out of the phase with hemispheric trends for past several centuries. Therefore, climate change in this region is not solely the temperature effect on melting as compared to other glaciated regions.

To identify the reasons for the establishment of the Karakoram Anomaly, monthly mean climatic variables for last five decades, reported from meteorological observatories at the valley floors in HKH region, are analyzed. In addition to the climatic variables of temperature and precipitation, monthly mean synoptic observations reported by meteorological observatories in both morning and afternoon, along with monthly mean radiosonde data are used. From these data, the role of different near-surface and upper atmospheric meteorological variables in maintaining the positive mass balance of the glaciers and the development of the Karakoram Anomaly can be explained. An overall warming in the region is observed. The trends in the summer temperatures, which were reported as decreasing a decade ago, are now found as increasing in updated time series. However, the overall gradient is still negative. The winter mean and maximum temperatures are increasing with accelerated trends. Both maximum and minimum temperatures in summer are not diverging anymore and the diurnal temperature range is decreasing in the most recent decade. The afternoon cloudiness is found as increasing throughout the year except for spring, which is indicative of an increase in convective
uplifting. Moreover, humidity is increasing all over the region; due to evaporation in the spring, from monsoon moisture advection in summer, and due to the recycling of monsoon moisture in autumn. Furthermore, near-surface wind speed and net radiation in the region are decreasing, explaining the decrease in the summer minimum temperature and the presence of the cloudy skies. The decrease in near-surface wind speed, and net radiation, and increase in water vapor pressure put a limit on the evapotranspiration process. In addition, winter and summer precipitation is increasing. The aridity index, which is based on the ratio of precipitation and reference evaporation, indicates that region is turning moisture surplus and energy deficient.

Surface atmospheric pressure and 700 hPa geopotential height is increasing due to warming in the bottom layers of the troposphere. Nighttime inversion in the lower tropospheric layers is decreasing due to warming. Analysis of gridded observed and reanalysis datasets indicates that they are not presenting a signal of change in accordance with the instrumental record. Furthermore, it is found that meteorological conditions during the summer season are still favorable for the sustenance of glaciers whereas more melting may occur in the spring season that may increase the early season river flows and may affect lower lying portions of the debris-free glaciers.
1. Introduction

The mountain ranges of the Hindukush, Karakoram, and Himalaya (HKH) are known as the “Third Pole” of the planet earth [Soncini et al., 2015], as they host the world’s largest glaciers outside the Polar Regions. These mountain ranges also serve as a water tower for the region [Immerzeel et al., 2010], from where the rivers Indus, Ganges, and Brahmaputra originate. These rivers supply water for irrigation and drinking to satisfy the needs of around 800 million people [Immerzeel et al., 2010; Bolch et al., 2012]. The glacier coverage of the Himalaya-Karakoram is about 40,800 km² (Himalaya, ~22,800 km²; Karakoram, ~18,000 km²) [Bolch et al., 2012]. It is reported in World Glacier Monitoring Service, WGMS, [2012, p.71] that the glaciers formed during the Little Ice Age (LIA) are now retreating in mountain ranges all over the world. This retreat is declared as a clear indicator of the global climate change. However, a few studies reported the expansion of glaciers in the highest elevated watershed of the central Karakoram in the late 1990s [Hewitt, 2005, 2007]. Changes in regional climate, glacier nourishment, extreme vertical gradients and heavy debris cover [Smiraglia et al., 2007], steepness and the thermal characteristics [Hewitt, 2005, 2007; Quincey et al., 2011; Kapnick et al., 2014; Rankl et al., 2014] of the ice mass are considered as the factors that are facilitating glacier growth and/or surges. This peculiar response of the glaciers to climate change is termed as “The Karakoram Anomaly”. Western Himalayan and the Karakoram are located in the subtropics. Owing to the location of the central Karakoram and western Himalaya, it is influenced by three distinct weather systems: i) westerly storms that account for the two third of the high altitude snow accumulation [Hewitt et al., 1989; Wake, 1989; Hewitt, 2005; Bolch et al., 2012], ii) summer monsoon advances contributing to one-third of the high altitude snow accumulation [Wake, 1989; Archer, 2003; Winiger et al., 2005; Seong et al., 2007], and iii) the anticyclone clear weather conditions that affects the behavior of the former two and enables solar radiation to reach and melt the glacier [Hewitt, 2005]. Therefore, the glaciers residing in the region are intermediate between the “summer accumulation type glaciers” from the Greater Himalayan to the east and the “winter accumulation type glaciers” of the European
Alps to the west [Hewitt, 2005] with characteristic of both summer and winter accumulation types [Hewitt, 2014, p.96]. Since the second half of the twentieth century, some conflicting signals are reported for temperature from the Upper Indus Basin (UIB) for the western Himalayan and Karakoram [Fowler and Archer, 2006]. Using valley based meteorological stations, it is found that the mean and minimum summer temperatures are decreasing and winter mean and maximum temperatures along with diurnal temperature range (DTR) in all seasons are increasing. Moreover, an increase/decrease in the runoff of the Hunza River, a tributary of the Indus River, is reported for warming/cooling in summer temperature [Fowler and Archer, 2006; Tahir et al., 2011]. This indicates that snow and glaciated ice is highly sensitive to changes in temperature. In the ablation zone of the glacier, reduction in summer temperature, an increase in cloudiness, reduction in insulation and the presence of more moisture should have a cooling effect [Archer and Fowler, 2004; Fowler and Archer, 2006]. However, in the accumulation zone of the glacier, increased cloudiness and precipitation [Archer and Fowler, 2004; Nepal and Shrestha, 2015], thermal inputs through release of latent heat [Quincey et al., 2011], and increased temperature of summer snowfall facilitates the transfer of ice to the lower parts of the glacier through surges or avalanches [Copland et al., 2011; Quincey et al., 2011, 2015; Bolch et al., 2012; Rankl et al., 2014]. This can turn the mass balance of the glacier positive [Hewitt, 2005, 2007]. Another suggestion of the anomalous mass gain for the Karakoram glaciers is forwarded by Gardelle et al. [2012], who measured regional changes in ice elevation by estimating the difference between two digital elevation models (DEMs) from the year 2000 and 2008 on the stable terrain. They demonstrated that the mass balance of Karakoram glaciers is indeed anomalous in comparison to the global average [Cogley, 2012]. Cogley, [2016] suggested that Karakoram anomaly is a zonal feature extending well to the east of the Karakoram proper, whereas Mukhopadhyay et al., [2014] found evidence of the positive mass balance is confined to the western part of the Karakoram where river flows are declining in summer, while the river flows in the eastern Karakoram are increasing in summer due to increasing monsoonal precipitation and glacier melt.
It is established from the literature review that 'The Karakoram Anomaly' exists, especially in the western part of the Karakoram ranges [Hewitt, 2005; Fowler and Archer, 2006; Smiraglia et al., 2007; Armstrong, 2010; Quincey et al., 2011; Tahir et al., 2011; Bolch et al., 2012; Cogley, 2012, 2016; Gardelle et al., 2012; Immerzeel et al., 2012, 2015; Kapnick et al., 2014; Morgan, 2014; Mukhopadhyay et al., 2014; Rankl et al., 2014; Kumar et al., 2015; Soncini et al., 2015; Kozhikkodan Veettil et al., 2016]. However, this considerable evidence of climate change in the Karakoram is far from a simple matter of temperature "cause" and melting "effect" [Hewitt, 2014], especially since Karakoram temperature has remained out of phase with hemispheric temperature trends for past five centuries [Zafar et al., 2016]. It challenges the simplistic view in which 'global warming' has the same impact everywhere [Hewitt, 2014]. Moreover, the ratio of glacier meltwater to number of inhabitants dependent on them is highest for the river Indus. This motivates the need for a thorough investigation of the Karakoram anomaly so that its impact on the availability of water resources can be ascertained and policies may be framed. This will ensure the availability of water to the population dependent on it.

The interpretation of the Karakoram anomaly is complicated by the lack of long-term programs focusing on field mass balance of the glaciers, and the scarcity of near-glacier, up-to-date climate data. Nevertheless, identifying and separating the factors that lead to the development of the Karakoram Anomaly is important. In this regard, the climatic data observed on valley floors [around and below 2,000 meters above sea level (masl)] during the last decades of the twentieth century can provide the first clue [Gardelle et al., 2012].

As already mentioned above, in the ablation zone of the glacier, a decrease in summer temperature, increase in cloudiness, decrease in insulation and the presence of more moisture should have cooling effect that discourages melting. Likewise, in the accumulation zone of the glacier, increased cloudiness and precipitation, an increase in thermal inputs through release of latent heat, and increased temperature of summer snowfall facilitates the transfer of ice to the lower part of the glacier through surges or avalanches. This can turn the glacier mass balance positive. Long-term, in-situ observations
of these meteorological variables are limited to valley based stations that are mostly situated away from the glaciers. Rugged and inaccessible terrain with extreme weather conditions does not allow field campaigns on a regular basis. Among all the factors that control the inputs, outputs and storage of snow and glaciated ice, as mentioned above, only temperature, precipitation, and river inflows have been analyzed extensively [Archer, 2003; Archer and Fowler, 2004; Fowler and Archer, 2006; Tahir et al., 2011; Cook et al., 2013; Hasson et al., 2015; Zafar et al., 2016]. Detailed analyses of other meteorological variables such as humidity, cloudiness, surface pressure, geopotential heights of the upper atmospheric pressure surfaces, wind speed at the near-surface and in the upper atmosphere, number of rainy days and the atmospheric temperature lapse rate have received far less attention. Understanding the behavior of these additional meteorological variables over time can further reveal different aspects of the Karakoram anomaly and may help to identify the underlying mechanism and feedback processes. Ultimately, it will also help the scientific community to forecast the state and fate of Karakoram and Himalayan glaciers in different climatic forcing scenarios and its implications on the availability of water to the ever increasing population dwelling downstream. Therefore, this study aims to perform a detailed seasonal analysis of conventional climatic variables, along with analyses of synoptic and upper atmospheric meteorological variables that are not discussed in the available literature in particular perspective of the Karakoram Anomaly. In-situ monthly mean maximum and minimum temperature and their average and diurnal range, and total accumulated precipitation are the strong indicators of climate change in the region and most of the literate explaining the Karakoram Anomaly discusses these climatic variables to describe the causes and impacts of the Karakoram Anomaly. In addition to these variables, monthly mean of synoptic meteorological observations, recorded in the morning (08 AM local time) and in the afternoon (05 PM local time), such as dry bulb temperature, relative humidity, cloud cover fraction, wind speed, surface pressure is analyzed to argue the direction and magnitude of the climate change in the Hindukush-Karakoram-Himalaya (HKH) mountain ranges centered in Northern Pakistan.
To quantify the hydroclimate signal of the region effectively, net-radiation, water vapor pressure, and reference evapotranspiration are derived from the monthly mean synoptic meteorological variables to analyze their trends over the last five decades. In addition to these, trends in the frequency of the wet days of the region are also evaluated to assess the changing signals in the hydroclimate regime.

The role of upper atmospheric characteristics in the development of the Karakoram Anomaly are still unknown. Therefore, this study analyzes the change in regional temperature lapse rate, together with change in temperature, wind speed and geopotential heights of 700 hpa and 500 hPa pressure surfaces, so that, impact and role of all these meteorological variables in development of the Karakoram Anomaly can be analyzed and the future of the Karakoram Anomaly and regional hydrologic balance can be argued.
2. Material and Methods

I. Study Area

The study area is comprised of eastern Hindukush, western Karakoram and northwestern Himalayan mountain ranges that are juxtaposed in the center of the province of Gilgit-Baltistan, situated in Northern Pakistan. Location map of Pakistan and study area is presented in Figure 1. This area supports several concentrations of glaciers at high altitude. To analyze glaciers response to the prevailing climatic conditions, meteorological stations with long-term climatic records are selected. All such stations are operated by Pakistan Meteorological Department (PMD) and most of them are located on the valley floors. Latitude, longitude, and elevation of these sites are presented in Table 1.

II. Description of Variables

Monthly mean maximum and minimum temperatures, monthly accumulated precipitation, and frequency of wet days, along with monthly mean synoptic observations of dry bulb temperature, wind speed, cloud cover fraction and relative humidity observed at in-situ meteorological stations are used in this study. A list of observed and derived meteorological variables is presented in Table 2. Monthly mean maximum and minimum temperature are available for 12 sites, whereas the monthly means of synoptic observations are available from 6 sites as presented in Table 1.

Temperature is the most important climatic variable and therefore can be used to compare different climatic conditions on both spatial and temporal scales. Daily maximum and minimum temperature are robust measures of the variation of temperature on daily scales. Further, maximum and minimum temperatures are still used to estimate average temperature and diurnal temperature range. All these metrics are important as they can be used to evaluate temporal changes in warming for a given location and spatial changes between different locations. For the HKH region, centered in Northern Pakistan, the temperature is the most discussed climatic variable and it is used to explain changes in various phenomena, ranging from hydrologic balance to socioeconomic affairs. Therefore,
monthly mean maximum and minimum temperature and their derived metrics are used in this study.

Monthly mean maximum/minimum temperature is the average of all daily maximum/minimum temperatures recorded in a month and the monthly mean temperature is the average of the monthly mean maximum and minimum temperatures. Similarly, the monthly mean diurnal temperature range is the difference of monthly mean maximum and monthly mean minimum temperatures. As the maximum/minimum temperature is independent of fixed timing, therefore, it is challenging to compute hydrometeorological information as other meteorological variables are usually recorded at fixed hours throughout the day. Therefore, to overcome this challenge, here monthly averages of dry bulb temperature, wind speed, cloud cover fraction, surface pressure, and relative humidity observed in the morning at 08 AM local time and 05 PM local time are also analyzed to explain the role of different meteorological variables in development and sustenance of the Karakoram Anomaly.

Precipitation is second most discussed climatological variables because of its importance as a source of life and sustenance. Due to topography, precipitation occurrence in the region is not uniform. Precipitation increases with elevation and the rate of increase is dependent on the slope and aspect of the terrain. Nevertheless, in this study, only monthly accumulated precipitation recorded at the valley floor are used as it is the most reliable information, that is available with long-term records.

Evapotranspiration is one of the important components in the hydrologic budgeting of a watershed. Though long-term records of precipitation and river discharge are available for different locations, such data for evapotranspiration is still not available. Therefore, water vapor pressure is estimated and reference evapotranspiration is evaluated here. This enables a better understanding of the regional hydrologic balance.

Monthly averages of dry bulb temperature, wind speed, cloud cover fraction, surface pressure, and relative humidity observed in the morning at 08 AM local time and 05 PM local time are used to compute vapor pressure and reference evapotranspiration as explained by [Shuttleworth, 2012].
Relationship between saturated vapor pressure and atmospheric temperature is explained by Clausius-Clapeyron and presented in equation (1), according to this relationship the equilibrium between water and water vapor is dependent on the ambient temperature as follows:

\[ \frac{1}{e_s} \frac{d e_s}{dT} = \frac{L_v}{R_v T^2} \tag{1} \]

Where ‘\(e_s\)’ is saturated vapor pressure of water, ‘\(L_v\)’ the latent heat of vaporization, and ‘\(R_v\)’ the gas constant for water, and ‘\(T\)’ the temperature in Celsius. Using empirically determined formula, the monthly mean vapor saturated vapor pressure ‘\(e_s\)’ is estimated using the monthly mean temperature reported in both morning and afternoon as follows:

\[ e_s = 0.6108 \exp \left( \frac{17.27T}{237.3 + T} \right) \quad \text{[kPa°C⁻¹]} \tag{2} \]

The gradient of the relationship between saturated vapor pressure and temperature represented by ‘\(\Delta\)’ is obtained by differentiating equation (2):

\[ \Delta = \frac{d e_{sat}}{dT} = \frac{4098 e_{sat}}{(237.3 + T)^2} \quad \text{[kPa°C⁻¹]} \tag{3} \]

Using monthly mean of relative humidity, recorded in both morning and afternoon, that is the ratio of the actual vapor pressure of the air to the saturated vapor pressure at air temperature and is expressed as a percentage, the actual vapor pressure is computed. Expression of actual vapor pressure ‘\(e\)’ is given by:

\[ e = \frac{R H \times e_{sat}}{100} \quad \text{[kPa]} \tag{4} \]

As evapotranspiration is strongly dependent on the net radiation that is the energy available for the process of evapotranspiration, it is imperative to evaluate the net radiation available in the region. The amount of solar radiation is considered constant at the top of the atmosphere i.e., 1367 \(\text{Wm}^{-2}\) (a.k.a solar constant \(S_o\)), however, it may vary due to earth’s elliptical orbit around the sun. The seasonal change in solar radiation is due to Sun-Earth distance. Earth revolves around the sun in an elliptical orbit, therefore, the distance between the two is not constant. However, this eccentricity can be parameterized
using a sinusoidal multiplicative factor which is called eccentricity factor ‘\(d_r\)’, that is the function of each day of the year in the following way:

\[
d_r = 1 + 0.033 \cos \left( \frac{2\pi}{365} D_y \right) \tag{5}
\]

Where ‘\(D_y\)’ is the Julian day of the year (that is taken as fifteenth day of each month over here) and the energy coming at the top of the earth’s atmosphere as solar radiation normal to the solar beam is calculated by:

\[
S_{top} = d_r S_o \tag{6}
\]

The amount of energy that is reaches the ground surface if there were no intervening atmosphere is dependent on the hour angle ‘\(\omega\)’, i.e., the angle representing time of day, is given by:

\[
\omega = \pi \left( \frac{12 - h}{12} \right) \text{ [radians]} \tag{7}
\]

Solar declination ‘\(\delta\)’ that is the angle between rays of the sun and the plane of the Earth’s equator, and is dependent on the day of the year, is given by:

\[
\delta = 0.4093 \sin \left( \frac{2\pi}{365} D_y - 1.405 \right) \text{ [radians]} \tag{8}
\]

The daily total incoming solar energy received on the ground integrated over daylight hours between time ‘\(t_1\)’ and ‘\(t_2\)’, the daily insolation ‘\(I_o\)’ is given by:

\[
I_o = \int_{t_1}^{t_2} S_o d_r (\sin \phi \sin \delta + \cos \phi \cos \delta \cos \omega) dt \tag{9}
\]

The values of ‘\(t_1\)’ and ‘\(t_2\)’ are best expressed in terms of the equivalent hour angle that defines both the beginning and end of the day. This angle is called the sunset hour angle, ‘\(\omega_s\)’, which can be calculated from:

\[
\omega_s = \arccos(-\tan \phi \tan \delta) \text{ [radians]} \tag{10}
\]

The length of the day, ‘\(N\)’, in hours as follows:

\[
N = \frac{24}{\pi} \omega_s \text{ [hours]} \tag{11}
\]

The total solar energy which would be received per unit area between sunrise and sunset on a horizontal surface at latitude ‘\(\phi\)’ is obtained by integrating equation (10) between sunset angle hours, as:
\[ S_0^d = 37.7 \, d_r (\omega_s \sin \phi \sin \delta + \cos \phi \cos \delta \cos \omega_s) \quad [MJ/m^2 d^{-1}] \]  

Where 37.7 = \( \frac{24 \times 3600 \times \text{sec}}{\pi} \left( \frac{1367 \, J/s}{1000000} \right) \), i.e., conversion of Wm\(^{-2}\) to MJm\(^{-2}\)d\(^{-1}\).

Atmospheric loss of solar radiation is parameterized using fractional cloud cover 'c', therefore the actual daily total solar radiation 'S\(^d\)', is given by:

\[ S^d = [a_s + (1 - c) b_s] S_0^d \]  

Empirical values of 'a\(_s\)' and 'b\(_s\)' would be derived locally by using measurements of insolation on overcast days to give 'a\(_s\)' and on days with continuous bright sunshine to give (a\(_s\) + b\(_s\)). Typical values derived this way are a\(_s\) = 0.25 and b\(_s\) = 0.5, corresponding to a 25% and 75% loss of energy in a clear sky and overcast conditions, respectively. These values of 'a\(_s\)' and 'b\(_s\)' are often assumed in the absence of any locally calibrated values.

Depending on albedo a portion of the incoming solar radiation is reflected back upon reaching earth surface. For HKH value of albedo 'a' is taken as 0.55 as suggested by [Ming et al., 2015]. Resultantly, the net daily solar radiation in short wave is less than incoming solar radiation and is given by:

\[ S_n^d = (1 - a) S^d \]

The temperature of the surface and air temperature of the lower atmosphere are related to each other and a great deal of the downward longwave radiation originates in the lower atmosphere comparatively close to the surface. Hence, both the upward and downward longwave radiation fluxes are linked to near-surface air temperature. A reasonably strong approximation of linear relationship exists between air temperature and both upward and downward longwave radiations. In the presence of clouds more of the outgoing longwave radiation is absorbed and returned to the surface, therefore, a simple empirical formula for the daily average net longwave radiation assessment is as follows:

\[ L_n^d = f \varepsilon' \sigma T_{air}^4 \]  

Where 'T\(_{air}\)' is the daily average air temperature (average of morning and afternoon temperature is used). 'a' is Stephen-Boltzmann constant. '\( \varepsilon' \)' is the effective emissivity that is dependent on the humidity content of the air and is estimated by:

\[ \varepsilon' = 0.34 + 0.14 \sqrt{e_d} \]
Where ‘\(e_d\)’ is the daily average vapor pressure, and factor ‘\(f\)’ in equation (15) is an empirical cloud factor which is the ratio of the estimated surface solar radiation calculated from equation (17) in ambient conditions to the value estimated for the same conditions for clear sky conditions. For arid conditions ‘\(f\)’ can be calculated using the following expression as explained by [Shuttleworth, 2012, p.62]:

\[
f = 1.35\left(\frac{s^d}{s^d_{\text{clear}}}\right) - 0.35
\]

(17)

Daily average net radiation flux, ‘\(R_n^d\)’ is given by:

\[
R_n^d = S_n^d + I_n^d
\]

\([MJ m^{-2} d^{-1}]\)  

(18)

Now reference evapotranspiration is calculated by utilizing some assumed constant parameters for a clipped grass reference crop with height of 0.12 m and a fixed surface resistance of 70sm\(^{-1}\), as follows:

\[
ET_0 = \frac{R_n + \gamma \frac{900}{T + 273} U_2 (e_s - e_a)}{\Delta + \gamma (1 + 0.34U_2)} \quad [mm d^{-1}]
\]

(19)

Where ‘\(U_2\)’ is wind speed observed from an anemometer that is 2-m high from the ground, and ‘\(e_a\)’ is actual vapor pressure, and Psychrometric Constant that relates the partial pressure of water in air to the air temperature is given as \(\gamma = \frac{C_F P}{\varepsilon \lambda}\).

Finally, a dimensionless Climatic Moisture Index (CMI) proposed by [Willmott and Feddema, 1992] is adopted to investigate the state of the moisture in the HKH region. The CMI is a ratio of precipitation ‘\(P\)’ to potential evapotranspiration ‘\(ET_0\)’, and is given as : 

\[
I_m = 100 \left[ \frac{P}{ET_0} - 1 \right]
\]

(20)

The CMI indicator ranges from -1 to +1. Where with wet climate showing positive CMI and dry climate negative CMI.

III. Trend Analysis

The trend of all the meteorological variables over the last five decades is evaluated through the least square method as explained by [Wilks, 2011]. Least square linear
regression is a maximum likelihood estimate i.e., given a linear model, what is the likelihood that this data set could have occurred in accordance with the model. Therefore, least squares linear regression evaluates the linear model that maximizes the likelihood of the representation. As geophysical phenomena approximate a normal distribution, especially when sample size increases, therefore, it is assumed that each data point has a measurement error that is independently random and normally distributed around the linear model with a variance. Now the probability events within the range of the data are the product of the probabilities at each point, given as:

\[
P \propto \prod_{i=1}^{N} \left\{ \exp \left[ -\frac{1}{2} \left( \frac{y_i-(a+bx_i)}{\sigma_i} \right)^2 \right] \Delta y \right\}
\]  

Maximization of this probability is equivalent to the minimization of the error, given by:

\[
\sum \left( \frac{y_i-(a+bx_i)}{\sigma_i} \right)^2
\]  

If the standard error at each point of the dataset is the same, then minimization is reduced to:

\[
\sum (y_i - (a + bx_i))^2
\]  

That can be solved by finding the partial derivatives with respect to “a” and “b” and substituting them equal to zero, and this give the best fit parameters for the regression constant and coefficient (α and β) as follows:

\[
\alpha = \frac{S_{xx}S_y - S_xS_{xy}}{\Delta} \quad \text{and} \quad \beta = \frac{NS_{xy} - S_xS_y}{\Delta}
\]  

Where \( \Delta = NS_{xy} - (S_x)^2 \) and \( S_x = \sum x_i, \ S_y = \sum y_i, \ S_{xy} = \sum x_iy_i, \ S_{xx} = \sum x^2 \)

To evaluate the robustness of the trend and change in both observed and derived meteorological variables, the significance is tested at 95%. For this, it is assumed that linear model is the best fit and all observations have the same variance throughout the duration under discussion. Moreover, it is assumed that the residuals are normally distributed around the linear model in following way:

\[
\sigma^2 = \frac{\sum (y_i - (a + bx_i))^2}{N - 2}
\]
Where \((N - 2)\) is referred to the degree of freedom as least square is a linear model based on two parameters of slope and intercept. Variance of \('b'\) can be given by:

\[
var(b) = \frac{\sigma^2}{\Sigma(x_i - \bar{x})^2}
\]  

\(26\)

Now, Student’s \(t\)-distribution can be used to define the multiplier \('t'\) for the confidence limits for the regression coefficients:

\[
b = \beta \pm t\sqrt{Var(b)}
\]  

\(27\)

The significance of trend of all observed and derived meteorological variables is evaluated within 95\% confidence interval as it is acceptable for climatic studies.

To assess the regional climate change all the analyses are performed on a seasonal scale. For this purpose, the whole year is divided into four seasons i.e., Winter, Spring, Summer, and Autumn. Winter season is comprised of the months of December, January, and February, and Spring is comprised of the months of March, April and May, and Summer season is comprised of the months of June, July and August, and Autumn season is comprised of the months of September, October and November.

**IV. Regional Mean Trend**

Regional mean trend of each climatic variable from 1961-2011 is estimated by taking an average of trends reported by all sites for the same variable in each season. Moreover, mean absolute difference (MAD) is evaluated to show the spread of trends around the mean trend.

**V. Standardized Anomalies**

It is common practice among climate data centers to update the climatic indices as anomalies and standardized anomalies relative to a 30-yr normal and variance. The World Meteorological Organization (WMO) suggests that the climatic normal should be based on a 30-yr period that starts at the beginning of each decade (1941-70, 1961-90, etc.). The base period of 1961-1990 is, therefore, selected to estimate the mean and variance of each season. Climatic variables are highly sensitive to latitude, surface type, elevation, aspect, proximity to large water bodies and cloud cover. Therefore, magnitude of different climatic
variables may differ from each other and to compare their changes they can be transformed into standardized anomalies with respect to the based period. This will preserve the general trend in the variable with an indication of positive and negative anomalies with their respective magnitudes. Therefore, seasonal anomalies of all climatic variables for the whole duration are computed by subtracting the climatological seasonal mean based on 1961-1990 from each year. Afterward, standardized anomalies are calculated by dividing the computed anomalies by the climatological seasonal standard deviation of the same base period. Resultant standardized anomalies are more informative about the magnitude of the anomalies because the influences of dispersion have been removed. It is not necessary that data have a specific distribution to express it in term of standardized anomalies.

VI. Upper Atmosphere Data

The radiosonde data of near most launching station i.e., Kabul (34.55° N, 69.22 ° E, 1791 masl, duration 1966-2016) is retrieved from the Integrated Global Radiosonde Achieve (IGRA), National Centers for Environmental Information (NCEI). IGRA maintains the radiosonde and pilot balloon observations at over 2,700 globally distributed stations. Daily vertical profile of temperature and wind speed with respect to elevation, temperature and wind speed at fixed pressure surfaces and height of pressure surfaces are available. All daily data is averaged on monthly scales to examine the relatively low-frequency variability.

VII. Gridded and Reanalysis Data

To demonstrate the strength of gridded observed and reanalysis datasets over glaciated areas of the Northern Pakistan, one dataset is selected as representative of each category. Gridded climate dataset from Climatic Research Unit (CRU) is selected as representative of the gridded observed data. CRU is a (0.5°×0.5° degree latitude/longitude) monthly dataset developed at School of Environmental Sciences, University of East Anglia, Norwich, UK and it is constructed from monthly observations at
meteorological stations across the world's land area [Harris et al., 2014]. It includes six mostly independent climatic variables i.e., mean temperature, diurnal temperature range, precipitation, wet-day frequency, vapor pressure and cloud cover. The latest updated version includes many new stations' data from the earlier record that have become available recently [Harris et al., 2014]. CRU is developed by merging monthly climate archives available through World Meteorological Organization (WMO) and National Oceanographic and Atmospheric Administration (NOAA) via National Climate Data Center (NCDC).

Modern-Era Retrospective analysis for Research and Application MERRA-2 is selected as representatives of the reanalysis datasets. MERRA-2 is a NASA atmospheric reanalysis that begins in 1980. It is upgraded from MERRA reanalysis using Goddard Earth Observing System Model, Version 5 (GOES-5) data assimilation system and Global Statistical Interpolation (GSI) analysis scheme. In comparison to European Centre for Medium-Range Forecast's (ECMWF) reanalysis product ERA-Interim, MERRA-2 is superior as its hourly diagnostics at a spatial resolution of 0.625° ×0.5° [Rienecker et al., 2011] are openly available to the public. MERRA-2 data are available online through the Goddard Earth Sciences (GES) Data and Information Services Center (DISC).
3. Results and Discussion

As already mentioned, climate change assessment in the Karakoram and Himalaya is not as simple as a change in one climatological variable may explain all the undergoing processes. Different meteorological variables offset each other with varying degree. This facilitates the development of the feedback mechanisms affecting the hydrological cycle. Recent climate change assessment from the instrument based observations report a different scenario from the assessment that was made a decade earlier [Fowler and Archer, 2006]. Overall warming in the region is observed and climatic variables which were reporting negative trends earlier are still reporting negative trends on longer scales. However, in very recent years, the sign of trend has changed.

Fowler and Archer, [2006] have analyzed instrumental records from the meteorological stations situated on valley floors in the Karakoram and Himalaya for the duration 1961-1999 and found increase in winter mean and maximum temperatures coupled with decrease in summer mean and minimum temperature (Table 3). Furthermore, they have found that summer temperature has an impact on the summer runoff of the Hunza River (i.e., an increase in summer seasonal temperature results in an increase in summer runoff). Moreover, a decrease in summer temperature and runoff is consistent with thickening and expansion of Karakoram glaciers as suggested by Hewitt, [2005] as well.

Comparison of seasonal trends of the maximum, minimum and mean temperature from the year 1961-2011 to that of Fowler and Archer, [2006] indicates that the thermal regime of the region has been changed from what was observed by them. Winter mean and the maximum temperature are now increasing with accelerated trends, whereas summer mean and minimum temperature are not cooling with the same rate as reported earlier. Table 3 compares the averages of the maximum, minimum and mean temperature trends from 1961-1999 as reported by Fowler and Archer, [2006] to that of 1961-2011 of same sites. Summer temperatures are still decreasing (−0.08 °C/Decade) though not as negatively accelerated as reported previously (−0.13 °C/Decade). Similarly, trend of winter and spring minimum temperature is no more negative as reported earlier. Winter
minimum temperature trend is changed from \(-0.02^\circ C/Decade\) to \(+0.04^\circ C/Decade\) and spring minimum temperature trend is changed from \(-0.24^\circ C/Decade\) to 0. Another interesting finding reported by the Fowler and Archer, [2006] was a consistent increase in Diurnal Temperature Range (DTR) in all seasons and annually, in direct contrast to both GCM projections and the narrowing of DTR globally. Analyses of updated data sets suggest that divergent trend of maximum and minimum temperature is not valid anymore, and DTR is narrowing in recent years. Though, long-term trends still suggest DTR broadening in all seasons and annually as presented in (Table 4).

Summer standardized meteorological variables observed at Gilgit station from 1961-2011 relative to mean and variance of 1961-1990 are presented in (Figure 2). Standardized anomalies of maximum temperature indicated frequent positive anomalies in the 1990s and negative anomalies in 2000s relative to the base period. Whereas negative standardized anomalies are found in minimum temperature from 1990 to the recent period though their magnitude is decreasing in most recent years. Converging trends of maximum and minimum temperatures, especially, since 2005 made DTR narrower that is in direct contrast to its long term trend [Hasson et al., 2015]. Mean temperature anomalies also suggest warming since the end of 1980s. Trends reported by updates time series (i.e., 1961-2011 in comparison to 1961-1999 as analyzed by [Fowler and Archer, 2006]) suggest a rise in maximum temperature in all seasons throughout the region. Except for the summer where a negative trend is reported \((-0.07 \pm 0.1^\circ C/Decade\) although some lower elevation sites did show a positive trend for this season as well (not shown here). The increased cloud cover fraction is responsible for the regional decrease in maximum temperature in summer season. On the other hand, in the spring the maximum temperature trend is positive and strongest among all the seasons \((0.43 \pm 0.12^\circ C/Decade)\). This results mainly due to a decrease in cloud cover fraction and increase in incoming insolation. Trends reported from updated time series indicates a decrease in summer minimum \((-0.35 \pm 0.17^\circ C/Decade)\) and mean temperature \((-0.21 \pm 0.1^\circ C/Decade)\) at regional scale. The strong positive maximum temperature trend in winter and spring causes an increase in mean temperature. Furthermore, positive trend in
DTR is reported in all seasons throughout the region with the highest trend in spring followed by autumn, summer and winter (see Table 4). Changes in the trends of temperature in the recent decade indicate that the region is ushered into the new climatic paradigm. However, it is not clear that the new paradigm is synonymous with the end of the Karakoram Anomaly or if it will further support the growth of glaciers as suggested by the Karakoram Anomaly. To investigate the role of the different climatic variable in the development of Karakoram Anomaly and its updated status, several synoptic variables reported by meteorological observatories in the morning and the afternoon are presented here.

The 2-meter dry bulb temperature (DBT) reported in the morning (08 AM local time) is very much related to minimum temperature in terms of its response. However, this variable shows a stronger warming as compared to the minimum temperature trend for all seasons except for autumn when its trend is stronger in negative direction (−0.31 ± 0.24 °C/Decade vs −0.2 ± 0.2 °C/Decade, respectively). Moreover, DBT reported in the afternoon (05 PM local time) is closely related to maximum temperature in its response. It suggests warming throughout the year (the warmest in the spring) except for the summer, same as maximum temperature. However, its trend in the autumn is not as strong as that of maximum temperature in the same season (Table 4). Standardized anomalies of the summer afternoon DBT have indicated cooling since the middle of the 1990s to the recent period with some positive anomalies of small magnitude (Figure 2). On the contrary, morning DBT anomalies indicate cooling throughout the same period with some decrease in very recent years. Accelerated warming is reported by all metrics of temperature in spring [Fowler and Archer, 2006; Hasson et al., 2015; Asad et al., 2016; Iqbal et al., 2016]. Variation in temperature is dependent on the amount of insolation at a particular latitude and season. Strongest control on the temperature is maintained by the cloud cover. During the day, the earth is heated by the sun and more solar insolation will reach to the earth’s surface and more sensible heat will be produced. Cloud cover intercepts the incoming solar insolation and reflects it back to the space, resultantly, less solar insolation will reach the earth and hence less warming will be produced. Considering such importance of the cloud...
cover, for the first time, in-situ observed cloud cover fraction in HKH is quantified as meteorological variable over here.

Morning and afternoon monthly mean synoptic observations suggest an increase in the cloud cover fraction in all seasons except spring where it is decreasing in both morning and afternoon (Table 4). Strong positive trend in cloud cover fraction is reported in the summer on the regional scale, and afternoons (0.15 ± 0.04% /Decade) are cloudier than mornings (0.11 ± 0.03 % /Decade). On the contrary, negative trends are reported in spring season and especially in the morning hours. Summer standardized anomalies of the cloud cover observed at Gilgit meteorological station report continuous positive trend in cloudiness since the 1980s with some occasional negative anomalies in the 1990s (Figure 2). Increased daytime cloudiness has a positive impact on the ablation zone of the glacier as more insolation will be intercepted and therefore glacier melt will be reduced [Hewitt, 2005, 2007, 2014; Fowler and Archer, 2006], that may turn the net mass balance of the glacier positive. On the other hand, more melting may occur on a cloudy night; therefore, cloudiness can be a negative forcing in the night. Though melting due to a cloudy night is minimal on high altitude because of the prevalence of very low temperature over there. Therefore, increased cloudiness in summer in the HKH region can affect snow and glacier melts seriously. More cloud cover in the afternoon as compared to the morning can be considered as a positive feedback on the ice bodies, as high sun insolation is intercepted in the afternoon, while the relatively smaller cloud fraction in the morning (or night) also has a positive influence and allows terrestrial radiation to escape into space, instead of intercepting it. Though, the sign of the cloud cover feedback is still a matter of uncertainty and depends on other cloud properties as well, nevertheless, increase in summer cloudiness is suggested by many [Archer, 2003; Hewitt, 2005, 2011, 2014; Fowler and Archer, 2006; Lau et al., 2006; Copland et al., 2011; Scherler et al., 2011; Cheema and Bastiaanssen, 2012; Zafar et al., 2016] and is related to the establishment of the Karakoram Anomaly. Increases in afternoon cloudiness throughout the years suggests an increase in convective uplifting on the regional scale. Especially, an increase in summer cloudiness indicates enhanced micro and mesoscale convective precipitation. Cloudiness is closely
related to the amount of humidity present in the atmosphere and vapor pressure is the best
description of the water vapor content in the atmosphere as it is independent of
temperature. Therefore, water vapor pressure needs to be quantified so that hydro-
meteorological signals can be interpreted correctly.

The increase in water vapor pressure (WVP) is reported in all seasons in both
morning and afternoon in the HKH region (Table 4). Highest increase is reported in the
summer and the autumn. For the summer afternoon larger increases are reported (0.07 ±
0.01 kPa/Decade) in comparison to the morning (0.05 ± 0.01 kPa/Decade). The
increase in WVP in the spring can be attributed to the increase in daytime temperature and
evaporation from the snow fields and soil. However, its increase in the summer is due to
penetration of monsoon moisture from the Bay of Bengal and the Arabian Sea. Moreover,
the increase in the autumn can be explained by the recycling of the moisture precipitated in
the summer. Summer standardized anomalies of the WVP (Figure 2) derived for Gilgit
meteorological station data indicate an ever-increasing trend since the 1970s that reached
to its maximum in mid-1980s and again in mid-1990s (> 2 standard deviations). The
increase in water vapor has profound effects on insolation by diffusion, and on terrestrial
radiation by absorption, and on energy partitioning by limiting evapotranspiration,
especially when wind speed is not removing the water vapor from a specific location,
effectively.

It goes without saying that the near surface wind speed is one of the most neglected
meteorological variables that is seldom discussed in the literature in the particular
reference of Karakoram and Himalaya. Wind speed along with vapor pressure and
radiation is an important meteorological variable that affects the partitioning of available
energy into sensible and latent heat for water balance assessment [Singh and Bengtsson,
2005; Winiger et al., 2005]. Near-surface wind speed in HKH region is decreasing over the
time and all around the year (Table 4). The decrease in the spring near surface wind speed
is higher than the winter and a larger decrease is reported in the afternoon in comparison
to the morning. Decreased wind speed in the morning is directly related to the decreased
minimum temperature (i.e. their linear coefficient of correlation ‘r^2’ is 0.47). A decrease in
wind speed can decrease the depth of the boundary layer. Therefore, the temperature of the lowest part of the troposphere decreases due to a decrease in the mechanical mixing of the sensible heat. Summer standardized anomalies of the morning and the afternoon wind speed (as presented in Figure 2) report a decrease in wind speed since the 1970s to late 1990s. Afterwards, negative anomalies are recorded for the morning near surface wind speed, and positive anomalies are observed in the afternoon near surface wind speed that indicates an increase in convective activity. Xu et al., [2006] explained that the East Asian summer monsoon wind speed is highly correlated with the incoming solar radiation. Hence, a decrease in wind speed coupled with an increase in water vapor and cloudiness implies a decrease in net radiation.

Net solar radiation of HKH is estimated using the albedo value as suggested by Ming et al., [2015] and cloud cover fraction as discussed in the methodology section. Incoming solar radiation is highly dependent on the cloud cover fraction as cloud intercepts incoming insolation and reflects it back into space. Therefore, incoming solar insolation is expected to decrease with the increase in the cloud cover fraction. The decrease in incoming insolation also reduces net radiation; the sum of incoming and outgoing short and longwave radiations (Equation 18).

The regional decrease in net radiation is found in all the seasons and annually, except for the spring where it is found to be increasing (+0.09 ± 0.03 MJm⁻² d⁻¹/Decade). The decrease in summer net radiation (−0.35 ± 0.07 MJm⁻² d⁻¹/Decade) is due to increased summer cloudiness. Summer standardized anomalies of the net radiation report negative anomalies since the late 1990s to the present (Figure 3). Moreover, a decrease in the summer near-surface wind speeds is also coexistent with a decrease in net radiation [Xu et al., 2006]. Wind speed, net radiation, and the amount of vapor pressure affect the capacity of the atmosphere to accommodate the addition of the water vapors from land surface through evapotranspiration. The decrease in the summer daytime temperature, wind speed, cloud cover fraction and net radiation coupled with an increase in vapor pressure favors the decrease in reference evaporation.
Reference evaporation is determined using thermodynamic relationships as presented in the materials and methods section. It is found decreasing throughout the year, especially in the summer and the autumn seasons. In the HKH region centered in Northern Pakistan, summer is the main season that facilitates the snowmelt and ablation from the glaciers, therefore a decrease in summer available energy and reference evapotranspiration hold a serious impact on the melting process. Summer standardized anomalies of the reference evapotranspiration indicates a negative trend since 1980 to the present (Figure 3). Evapotranspiration is an important component of the water balance complemented by precipitation and runoff. Though runoff is not the subject of this study, however regional change in precipitation is assessed. The total accumulated precipitation is found increasing in the winter and the summer seasons, whereas, it is decreasing in the spring and does not show much change in the autumn (Table 4). Another important metric of precipitation is its frequency on the monthly or seasonal scale. The frequency of wet days is found increasing in all seasons (Table 4). Moreover, trends in annual percentage precipitation indicate an increase in winter and summer and a decrease in spring and autumn. The increase in the frequency of rainy days with a decrease in accumulated precipitation and reduction of cloud cover in the spring can only be explained by an increase in afternoon convective activity.

Potential evapotranspiration and precipitation can be combined to estimate Aridity Index (A.I.) that is an indicator of the degree of dryness of the climate of a given location. Due to increase in the summer accumulated precipitation and decrease in potential evapotranspiration the A.I. indicates (Figure 3) that the region is turning more moist and energy deficient and that has serious implications for hydrological balance in the region.

It is reported that Elevated Heat Pump mechanism in the summer at western Tibetan Plateau (HKH) is due to absorbing aerosols transported from nearby deserts and local back carbon emissions [Lau and Kim, 2006]. The lapse rate of maximum, mean and minimum temperature and diurnal temperature range is determined using station elevation. Elevated sites are found more heated than lower ones in winter and autumn. However, they are cooler in summer in daytimes. In the night, elevated sites are getting
cooler throughout the study period that made diurnal temperature range broad at higher elevations in the study period as presented in (Table 5). Collectively it indicates that sites located at high altitude in the Karakoram are reporting stronger signal of change [Hasson et al., 2015].

Another most important meteorological variable that needs more attention, in the purview of Karakoram Anomaly, is the surface pressure. The surface pressure tendency is positive in all seasons and strongest in the summer and the autumn seasons. A stronger increase is noticed in the summer afternoon than its morning. Summer standardized anomalies of surface pressure recorded at Gilgit station indicate a steady increase in the surface pressure throughout the study period (Figure 3). As already discussed, that increase in surface pressure is not attributed to increasing in water vapor, or greenhouse gas as their contribution is small in magnitude. However, the main culprit is the warming in bottom most layers of the troposphere. In the hydrostatic atmosphere, warming in a layer increases pressure in that layer. The regional warming that occurs at lower elevations is reflected in an increase in surface pressure at the elevated sites [Toumi et al., 1999; Moore, 2012]. Thermodynamic equations that describe this phenomenon are discussed in the Appendix. Due to the warming in the lowermost layers of the troposphere the environmental lapse rate up to 6 km reported by radiosonde launched at Kabul station indicates positive trend in all the seasons, and stronger in night (00 UTC, 0430 AM Local time) than daytime (12 UTC, 0430 PM Local Time) in the summer and the autumn seasons, and vice versa in the winter season (Table 5). The stronger positive increase in the winter season daytime lapse rate can be attributed to the positive trend of maximum temperature at the low elevated sites (around 2000 masl). Whereas an increase in the summer nighttime lapse rate up to 6 km is mainly due to drop in temperature of troposphere above 5000 masl and an increase in temperature below 3000 masl. Moreover, nighttime inversion in recent years is not as strong as it was earlier. Lower troposphere vertical temperature profiles suggest strong inversion in the 1980s. Especially for the summer season that is changed in recent years. Figure 4 shows monthly mean vertical temperature profile of the lower troposphere in the year 1983 (blue line) and in the year 2007 (red dashed line) at
00-UTC, where inversion can be noticed in the lowermost layer. Temperature inversion in the bottom most layer in the summer season is mainly due to a low temperature at sites around 2000 masl, though inversion signature is still present in recent radiosonde profiles (year 2007). However, the temperature in the lowermost layer is increased and therefore inversion is not as strong as it was earlier. On the other hand, daytime summer temperature throughout the lower troposphere up to 6000 masl is either decreased or remained unchanged (mean monthly vertical temperature of the atmosphere from year 1980 and the year 2009 is compared in Figure 5) therefore a change in lapse rate is not as strong as nighttime (Table 5).

The inversion of temperature in lower troposphere can also be related to the decrease in wind speed. The decrease in wind speed reduces mechanical mixing of the air. Therefore, the temperature in the near surface air will further decrease due to the absence of mixing from the upper layer and strong inversion will occur. Equation (35) express the mechanical production due to the conversion of energy between mean flow and turbulent fluctuations and this mechanical production is proportional to the shear in the mean flow. Therefore, a decrease in minimum temperature is related to a decrease in near-surface wind speed that helps to establish strong inversion.

Trend analysis of 700 hPa pressure surface suggests that its geopotential height is increasing in both night and daytime in all seasons (Table 5). The highest increase in spring followed by summer and autumn indicates the impact of warming on the geopotential heights of the pressure surfaces in the troposphere. Moreover, temperature analysis of 700 hPa pressure surface also suggests a substantial increase in spring due to heating of bottom layers of the troposphere in HKH. The increase in 700 hPa geopotential height in the summer season coupled with not too much increase in its temperature suggests that increase in height is due to the expansion of lowest tropospheric layers that are below the mountain.

As mentioned before, the available literature on high mountain Asia is sometimes contradictory, merely reporting trends without significance and providing an incomplete picture due to their lack of full spatial and temporal coverage [Kapnick et al., 2014].
Meteorological records, for instance, provide data for the past 50 years and it is hard to trace meaningful change within the trends reported by temperature and precipitation that is sometimes non-significant. In case instrumental observations are spatially limited, gridded observed and reanalysis datasets can be considered as an alternate. Therefore, temperature trends from gridded climate data set from CRU are compared with those of instrumental record for the summer season as presented in (Figure 6). Negative trends of the summer maximum temperature reported by the instrumental record are not consistent with CRU dataset. The instrumental records are indicating more cooling at higher latitude sites in HKH in comparison to CRU, whereas instrumental record from low latitude sites indicates warming that is not presented by CRU. Similarly, all instrumental records are reporting cooling in mean and minimum temperature, whereas CRU is reporting warming significantly throughout the region. Finally, the instrumental record indicates the varying degree of increase in DTR on most of the sites. However, CRU is reporting a consistent decrease in DTR throughout the region. It is mainly because data from none of the stations located in HKH is incorporated in the development of CRU gridded data, especially in maximum, minimum and mean temperature products. Observations used to build the CRU TS 3.21 data sets as reported by Centre for Environmental Data Analysis (CEDA) indicates that precipitation records from Dir, Gilgit, Astore, Bunji, and Skardu are included for development of precipitation gridded data, however, temperature data is not incorporated. Therefore, trends of temperature in HKH region are not a true indicator of the region.

Similarly, MERRA-2 is selected as representative of the modern reanalysis equipped with an observation from satellite and data assimilation. Summer trends of maximum, minimum, mean temperature and diurnal temperature range reported by instrumental record and MERRA-2 from 1981 to 2011 are presented in (Figure 7). An increase in maximum temperature is reported by many sites in this period. However, MERRA-2 is indicating a decrease in western part of HKH centered in northern Pakistan and it is indicating increase in the eastern part of the region. For minimum temperature, instrumental record is reporting negative trend whereas MERRA-2 is reporting positive
trend. Mean temperature trends reported by both instrumental record and MERRA-2 are more in agreement. DTR trends are also not in agreement with each other.

Disagreement between the trends reported by the instrumental observations and the gridded observed and the reanalysis data sets indicates that one needs to be really careful in the use of such information as it is not representing the climate change occurring in the region. To overcome this discrepancy, it is suggested that extensive meteorological monitoring network should be established in the region so that reanalysis datasets can be improved by removal of biases on different elevations. Such improved products can be helpful for hydrologic budgeting of the region and to diagnose the change in a climate that may affect the snow and glaciated ice reserves.
4. Conclusions

In this study, quantified assessment of the change in in-situ meteorological variables contributing to the establishment of the Karakoram Anomaly is presented. Detailed analyses of in-situ maximum, mean, and minimum temperature, diurnal temperature range, humidity, cloudiness, precipitation and frequency of the rainy days, surface pressure, wind speed, incoming solar radiation, potential evapotranspiration, environmental lapse rate and the upper atmospheric temperature and geopotential height of the pressure surfaces on a regional scale are presented.

Overall warming in the region is observed and the negative trends in the summer mean and minimum temperatures that were reported a decade earlier as a potential reason for the establishment of the Karakoram Anomaly are no more valid. Fowler and Archer, [2006] reported an increase in the winter mean and maximum temperatures and a decrease in the summer mean and minimum temperatures coupled with an increase in DTR in all seasons that was in direct contrast to both GCM projections and the narrowing DTR globally. With the addition of twelve more years of the data, the analysis of same climatic variables from 1961-2011 reveals that winter mean and maximum temperatures are now increasing with accelerated trends, whereas summer mean and minimum temperatures are not cooling at the same rate as reported earlier. Although the long-term summer temperature trend is still negative, an increase is reported since the onset of the twenty-first century. The DTR trend that was reported positive previously, turned negative since the first decade of the twenty-first century.

Analysis of the monthly means of in-situ morning and afternoon synoptic observations suggests that cloudiness is increasing in all seasons except spring. The increase in summer cloudiness is higher than other seasons, where afternoons are cloudier than mornings. This may affect the mass balance of the glaciers at the regional scale. The increase in daytime cloudiness may be taken as a positive impact especially in the ablation zone of the glaciers, as more insulation will be intercepted and therefore glacier melt will be reduced. On the other hand, cloudy nights can facilitate more melting of the glaciated ice by intercepting outgoing terrestrial radiations. Nevertheless, this contribution is minimal
due to a very low nighttime temperature at elevated locations. Therefore, more cloudiness in the afternoon in comparison to the morning can be considered as positive impact as it will intercept incoming insolation and less cloudiness in morning or nighttime will help more terrestrial radiation to escape to space. Hence, the net effect is positive and potentially facilitates a positive mass balance of the glaciers. The increase in summer afternoon cloudiness also indicates enhanced micro- and mesoscale convective uplifting that may increase the occurrence of precipitation in the accumulation zone of the glaciers.

The increase in cloudiness is a natural corollary of the increase in humidity. Here, water vapor pressure is analyzed instead of relative humidity, as the former is independent of temperature. Water vapor pressure is increasing in the HKH region throughout the year. It is believed that the increase of water vapor pressure in spring is due to evapotranspiration from melting snow fields and from soil through plants, especially when spring cloudiness is decreasing and net radiation is increasing. The increase in summer's water vapor pressure is mainly due to the intrusion of monsoon moisture from Bay of Bengal and Arabian Sea, whereas the increase in the autumn's water vapor pressure is due to recycling of the water vapor from earlier season’s precipitation. Net radiation and water vapor both affects each other. The increase in net radiation can increase evapotranspiration that can increase water vapor pressure, whereas an increase in water vapor pressure may decrease the incoming insolation by diffusion, can absorb outgoing terrestrial radiation and may limit the evapotranspiration by saturating the lower atmosphere, especially if the wind speed is not removing it from the region.

Near surface wind speed in the HKH region is decreasing throughout the year, especially in the spring and the summer seasons, and with a larger decrease in the afternoon than morning. A significant correlation exists between the morning wind speed and minimum temperature. It indicates that an increase/decrease in near-surface wind speed is related to an increase/decrease in minimum temperature. The decrease in near-surface wind speed may affect the mechanical mixing within the boundary layer. Moreover, wind speed is also correlated to the net radiation [Xu et al., 2006].
A decrease in wind speed with an increase in water vapor pressure and cloudiness can seriously affect the net radiation by obstructing incoming insolation. Decreased net radiation is observed throughout the year except for the spring season when it is increasing. The decrease in summer net radiation is a long-term trend that is continued to the recent era. Decreased wind speed, net radiation, and increase in water vapor pressure is affecting evapotranspiration processes. Reference evapotranspiration were determined using standard thermodynamic relationships and it is found that it is decreasing throughout the year. However, its decrease is minimal in the winter and the spring and substantial in the summer and the autumn. The summer standardized anomalies of the Gilgit meteorological station indicates that negative reference evapotranspiration anomalies are coupled with negative anomalies in net radiation, minimum temperature and near surface wind speed, and positive anomalies in surface pressure, cloud cover fraction, and water vapor pressure. Reference evapotranspiration is decreasing since early 1980s and it remained below its normal of 1961-1990 with varying degree.

Precipitation is another important component of the hydrological budgeting of an area. An increase is found in the winter and the summer precipitation and it is decreasing in the spring. Whereas the percentage of seasonal precipitation to the total annual precipitation indicates that there is an increase in the winter and the summer, and decrease in the spring and the autumn. Moreover, the frequency of the rainy days is also found increasing throughout the year.

A.I. is an indicator of the degree of dryness of the climate of a given location that is made up of potential evapotranspiration and precipitation. It indicates that the HKH region is turning moist and energy deficient that may facilitate the positive mass balance of the glaciers.

The surface atmospheric pressure of the region is found increasing since the 1980s. An increase in the surface atmospheric pressure at mountainous location is due to the warming of bottom layers of the troposphere. Moreover, an increase in geopotential height of the lower atmospheric layers is also reported by radiosonde data and it is also because of warming of the bottom-most layers of the troposphere. Increase in lower troposphere
temperature lapse rate also indicates warming in the bottom layers. The positive trend in temperature lapse rate is noticed all around the year, which is stronger in the night than daytime in the summer and vice versa in the winter.

In the 1980s strong inversion in lowermost layers of the troposphere is noticed that is not as strong in 2000s as it was earlier. Temperature inversion in the 1980s is related to decrease in wind speed and thus decreased mechanical mixing of the boundary layer that facilitated an anomalous decrease in minimum temperature.

To evaluate the potential of gridded observed and reanalysis data sets to capture the signal of climate change within HKH region, temperature trends from CRU TS and MERRA-2 are compared to in-situ instrumental records for the same period and it is found that they are not in agreement with each other. Therefore, it is suggested that the scientific community needs to be very much careful in using information from these sources for the detection of climate change and its impact on hydrologic balance in HKH region.

To overcome the scarcity of the meteorological information from HKH region, the establishment of the extensive meteorological observation network covering different elevation heights and representative topography is recommended so that reanalysis datasets can be improved for the scientific investigation in the region.

In conclusion, it is found that different climatic variables are offsetting each other in increasing and decreasing patterns since last five decades. Summer temperature that were indicating cooling previously now indicates warming in the region. However, other meteorological variables like an increase in cloud cover fraction, precipitation and wet days frequency, and intrusion of monsoon moisture, and a decrease in net radiation and near surface wind speed still favors the positive mass balance of the high elevated glaciers. On the other hand, anomalous warming is reported in the spring season, which is coupled with decrease in cloud cover fraction and precipitation, and with increase in net radiation. All these conditions favor the early melt from snow field and debris free glaciers that will increase the inflows in River Indus in the spring season. Finally, the Karakoram Anomaly is the response of the high-altitude orography to various climatic conditions on the regional scale that is not synchronous with the climate change patterns exhibited in lower elevated
neighboring areas. The change in the spring and the summer hydroclimate signal suggests that region is ushered into a new climatic paradigm where glaciers mass balance will be settled according to the new configurations. However, other meteorological indicators still suggest that signal of positive mass balance may continue to prevail through the year except the spring season.
References


Hasson, S., J. Böhner, and V. Lucarini (2015), Prevailing climatic trends and runoff response from


Immerzeel, W. W., L. P. H. van Beek, and M. F. P. Bierkens (2010), Climate change will affect the Asian water towers., Science, 328(5984), 1382–5, doi:10.1126/science.1183188.


Singh, P., and L. Bengtsson (2005), Impact of warmer climate on melt and evaporation for the rainfed,


Toumi, R., N. Hartell, K. Bignell, and G. Zi (1999), change p • o Mg zv fo • Mg dz ) Lhasa, , 26(12), 1751–1754.


Figure 1: Location of Study Area.
Location of Pakistan in South Asia and a list of sites that are used in the study with their elevation in meters are presented in the sub-panels.
Figure 2: Standardized Anomalies of mean Maximum, Minimum, Mean Temperature (°C), Diurnal Temperature Range (°C), Dry Bulb Temperature (°C), Cloud Cover (Okta), Saturated Vapor Pressure (kPa) and Wind Speed (ms-1) in morning (08 AM Local time) and afternoon (05 PM local time) of Gilgit Station for Summer Season from 1961-2011 with respect to mean and variance of the base period of 1961-1990.
Figure 3: Standardized Anomalies of mean Net-Radiation (MJm-2d-1), seasonal total precipitation (mm), frequency of rainy days (#), Surface Pressure (kPa), Reference Evaporation (mm), and Climate Moisture Index (%) of Gilgit Station for Summer Season from 1961-2011 with respect to mean and variance of the base period of 1961-1990.
Figure 4: Monthly Mean Vertical Temperature of the Atmosphere up to 6 Km reported by Radiosonde Launched at Kabul, Afghanistan in the Year 1983 and 2007 at 00-UTC (0430 AM, Local Time). The year 1983 and the year 2007 are selected as representatives of the decades of the 1980s and the 2000s, respectively, as there are no missing values in those years at 00-UTC.
Figure 5: Monthly Mean Vertical Temperature of the Atmosphere up to 6 Km reported by Radiosonde Launched at Kabul, Afghanistan in the Year 1980 and the 2009 at 12-UTC (0430 PM, Local Time). The year 1980 and the year 2009 are selected as representatives of the decades of the 1980s and the 2000s, respectively, as there are no missing values in those years at 12-UTC.
Figure 6: Comparison of summer seasonal trends of monthly mean maximum, minimum, and mean temperature and diurnal temperature range from 1961 to 2011 estimated by using in-situ instrumental meteorological data (circles) and CRU gridded data (contours).
Figure 7: Comparison of summer seasonal trends of monthly mean maximum, minimum, and mean temperature and diurnal temperature range from 1981 to 2011 estimated by using in-situ instrumental meteorological data (circles) and MERRA-2 reanalysis (contours).
# Tables

Table 1: List of meteorological sites used in the study with their latitude, longitude, elevation and description of the data used in the study.

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation</th>
<th>Data Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Astor</td>
<td>35.3</td>
<td>74.9</td>
<td>2167</td>
<td>☀/☁</td>
</tr>
<tr>
<td>Bunji</td>
<td>35.6</td>
<td>74.6</td>
<td>1372</td>
<td>☀/☁</td>
</tr>
<tr>
<td>Chillas</td>
<td>35.6</td>
<td>74.1</td>
<td>1250</td>
<td>☀/☁</td>
</tr>
<tr>
<td>Chitral</td>
<td>35.8</td>
<td>71.8</td>
<td>1499</td>
<td>☀</td>
</tr>
<tr>
<td>Dir</td>
<td>35.2</td>
<td>71.8</td>
<td>1375</td>
<td>☀</td>
</tr>
<tr>
<td>Drosh</td>
<td>35.5</td>
<td>71.7</td>
<td>1463</td>
<td>☀</td>
</tr>
<tr>
<td>Gilgit</td>
<td>35.9</td>
<td>74.3</td>
<td>1459</td>
<td>☀/☁</td>
</tr>
<tr>
<td>Gupis</td>
<td>36.2</td>
<td>73.4</td>
<td>2156</td>
<td>☀/☁</td>
</tr>
<tr>
<td>Garhi Dupatta</td>
<td>34.1</td>
<td>73.6</td>
<td>812</td>
<td>☀</td>
</tr>
<tr>
<td>Kotli</td>
<td>33.5</td>
<td>73.9</td>
<td>613</td>
<td>☀</td>
</tr>
<tr>
<td>Muzaffarabad</td>
<td>34.2</td>
<td>74</td>
<td>701</td>
<td>☀</td>
</tr>
<tr>
<td>Skardu</td>
<td>35.3</td>
<td>75.6</td>
<td>2317</td>
<td>☀/☁</td>
</tr>
</tbody>
</table>

Symbol of ☀ indicates that monthly mean of maximum, mean and minimum temperature and accumulated precipitation data from that site is utilized and symbol of ☁ indicates that synoptic data reported by meteorological station in morning and afternoon is utilized for the study.
Table 2: List of observed and derived meteorological variables utilized to perform analyses.

<table>
<thead>
<tr>
<th>Variable Name</th>
<th>Units</th>
<th>Retrieval Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum Temperature</td>
<td>°C</td>
<td>Observed</td>
</tr>
<tr>
<td>Minimum Temperature</td>
<td>°C</td>
<td>Observed</td>
</tr>
<tr>
<td>Mean Temperature</td>
<td>°C</td>
<td>Derived</td>
</tr>
<tr>
<td>Diurnal Temperature Range</td>
<td>°C</td>
<td>Derived</td>
</tr>
<tr>
<td>Precipitation</td>
<td>mm</td>
<td>Observed</td>
</tr>
<tr>
<td>Dry Bulb Temperature</td>
<td>°C</td>
<td>Observed</td>
</tr>
<tr>
<td>Wind Speed</td>
<td>m/s</td>
<td>Observed</td>
</tr>
<tr>
<td>Cloud Cover Fraction</td>
<td>%</td>
<td>Observed</td>
</tr>
<tr>
<td>Surface Pressure</td>
<td>kPa</td>
<td>Observed</td>
</tr>
<tr>
<td>Frequency of Rainy Days</td>
<td>#</td>
<td>Observed</td>
</tr>
<tr>
<td>Relative Humidity</td>
<td>%</td>
<td>Observed</td>
</tr>
<tr>
<td>Vapor Pressure</td>
<td>kPa</td>
<td>Derived</td>
</tr>
<tr>
<td>Net Radiation</td>
<td>MJ m²d⁻¹</td>
<td>Derived</td>
</tr>
<tr>
<td>Reference Evaporation</td>
<td>mm</td>
<td>Derived</td>
</tr>
<tr>
<td>Aridity Index</td>
<td>%</td>
<td>Derived</td>
</tr>
</tbody>
</table>
Table 3: Comparison of mean seasonal and annual trends in temperatures from 1961 to 1999 as reported by [Fowler and Archer, 2006, p.4280] in their (Table 3) to the same from 1961 to 2011 with updated dataset computed from Dir, Drosh, Bunjí, Gilgit, Astore, and Skardu. Positive trends are denoted with red color and negative trends are denoted with blue color. The increase in trend is indicated by the red up arrow and a decrease in trend is indicated by the blue down arrow.

<table>
<thead>
<tr>
<th>Duration</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tx</td>
<td>+0.25</td>
<td>+0.3↑</td>
<td>+0.03</td>
</tr>
<tr>
<td>Tn</td>
<td>-0.02</td>
<td>+0.04↑</td>
<td>-0.24</td>
</tr>
<tr>
<td>Tm</td>
<td>+0.11</td>
<td>+0.17↑</td>
<td>-0.18</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Autumn</th>
<th>Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tx</td>
<td>+0.21</td>
<td>+0.2↓</td>
</tr>
<tr>
<td>Tn</td>
<td>-0.46</td>
<td>-0.26↑</td>
</tr>
<tr>
<td>Tm</td>
<td>-0.13</td>
<td>-0.03↑</td>
</tr>
</tbody>
</table>

- Srinagar is excluded to evaluate mean trend of Tm
Table 4: Mean regional trends of different meteorological variables of HKH from 1961 to 2011 with a range of mean absolute difference are presented. Red colored box indicates absolute positive trend of the meteorological variable throughout the region, blue colored box indicates absolute negative trend of the meteorological variable throughout the region, and light red colored box indicates mean positive trend with some signal of negative trend in it and light blue colored box indicate negative trend with some signal of positive trend in it. Boxes with white background indicate mean neutral trend with both positive and negative signals.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Unit</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Autumn</th>
<th>Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tx</td>
<td>°C</td>
<td>+0.29 ± 0.16</td>
<td>+0.43 ± 0.12</td>
<td>-0.07 ± 0.1</td>
<td>+0.18 ± 0.15</td>
<td>+0.21 ± 0.14</td>
</tr>
<tr>
<td>Tn</td>
<td>°C</td>
<td>+0.02 ± 0.11</td>
<td>+0 ± 0.06</td>
<td>-0.35 ± 0.17</td>
<td>-0.2 ± 0.2</td>
<td>-0.14 ± 0.13</td>
</tr>
<tr>
<td>Tm</td>
<td>°C</td>
<td>+0.15 ± 0.13</td>
<td>+0.22 ± 0.07</td>
<td>-0.21 ± 0.1</td>
<td>-0.01 ± 0.08</td>
<td>+0.04 ± 0.06</td>
</tr>
<tr>
<td>DTR</td>
<td>°C</td>
<td>+0.26 ± 0.15</td>
<td>+0.43 ± 0.12</td>
<td>+0.28 ± 0.18</td>
<td>+0.38 ± 0.21</td>
<td>+0.36 ± 0.21</td>
</tr>
<tr>
<td>DBT 08 AM</td>
<td>°C</td>
<td>+0.08 ± 0.09</td>
<td>+0.1 ± 0.05</td>
<td>-0.3 ± 0.12</td>
<td>-0.31 ± 0.24</td>
<td>-0.12 ± 0.12</td>
</tr>
<tr>
<td>DBT 05 PM</td>
<td>°C</td>
<td>+0.18 ± 0.09</td>
<td>+0.32 ± 0.07</td>
<td>-0.18 ± 0.13</td>
<td>+0.03 ± 0.09</td>
<td>+0.08 ± 0.05</td>
</tr>
<tr>
<td>Cloud Cover 08 AM</td>
<td>(%)</td>
<td>0 ± 0.01</td>
<td>-0.06 ± 0.02</td>
<td>+0.11 ± 0.03</td>
<td>+0.02 ± 0.02</td>
<td>+0.02 ± 0.02</td>
</tr>
<tr>
<td>Cloud Cover 05 PM</td>
<td>(%)</td>
<td>+0.02 ± 0.04</td>
<td>-0.01 ± 0.02</td>
<td>+0.15 ± 0.04</td>
<td>+0.1 ± 0.03</td>
<td>+0.06 ± 0.04</td>
</tr>
<tr>
<td>Rainy Days</td>
<td>(#)</td>
<td>+0.75 ± 0.13</td>
<td>+0.94 ± 0.49</td>
<td>+0.95 ± 0.53</td>
<td>+0.43 ± 0.35</td>
<td>+3.03 ± 1.35</td>
</tr>
<tr>
<td>Precipitation</td>
<td>(mm)</td>
<td>+3.92 ± 1</td>
<td>-1.6 ± 2.53</td>
<td>+4.33 ± 1.72</td>
<td>+0.35 ± 0.58</td>
<td>+6.69 ± 2.18</td>
</tr>
<tr>
<td>Vapor Pressure 08 AM</td>
<td>kPa</td>
<td>0</td>
<td>+0.02</td>
<td>+0.05 ± 0.01</td>
<td>+0.02</td>
<td>+0.02</td>
</tr>
<tr>
<td>Vapor Pressure 05 PM</td>
<td>kPa</td>
<td>+0.01 ± 0.01</td>
<td>+0.02</td>
<td>+0.07 ± 0.01</td>
<td>+0.03 ± 0.01</td>
<td>+0.03 ± 0.01</td>
</tr>
<tr>
<td>Surface Pressure 08 AM</td>
<td>kPa</td>
<td>+0.19 ± 0.12</td>
<td>+0.13 ± 0.1</td>
<td>+0.29 ± 0.09</td>
<td>+0.21 ± 0.07</td>
<td>+0.19 ± 0.09</td>
</tr>
<tr>
<td>Surface Pressure 05 PM</td>
<td>kPa</td>
<td>+0.2 ± 0.14</td>
<td>+0.08 ± 0.11</td>
<td>+0.34 ± 0.2</td>
<td>+0.21 ± 0.06</td>
<td>+0.18 ± 0.13</td>
</tr>
<tr>
<td>Wind Speed 08 AM</td>
<td>ms⁻¹</td>
<td>-0.08 ± 0.03</td>
<td>-0.09 ± 0.06</td>
<td>-0.08 ± 0.04</td>
<td>-0.08 ± 0.01</td>
<td>-0.08 ± 0.02</td>
</tr>
<tr>
<td>Wind Speed 05PM</td>
<td>ms⁻¹</td>
<td>-0.08 ± 0.06</td>
<td>-0.14 ± 0.15</td>
<td>-0.17 ± 0.14</td>
<td>-0.14 ± 0.09</td>
<td>-0.13 ± 0.16</td>
</tr>
<tr>
<td>Net Radiation</td>
<td>MJm⁻²d⁻¹</td>
<td>-0.02 ± 0.04</td>
<td>+0.09 ± 0.03</td>
<td>-0.35 ± 0.07</td>
<td>-0.13 ± 0.03</td>
<td>-0.1 ± 0.065</td>
</tr>
<tr>
<td>Reference Evaporation</td>
<td>mmd⁻¹</td>
<td>-0.02 ± 0.01</td>
<td>-0.01 ± 0.05</td>
<td>-0.34 ± 0.06</td>
<td>-0.17 ± 0.03</td>
<td>-0.14 ± 0.08</td>
</tr>
<tr>
<td>Aridity Index</td>
<td>(%)</td>
<td>+0.09 ± 0.01</td>
<td>-0.01 ± 0.02</td>
<td>+0.02 ± 0.01</td>
<td>+0 ± 0.01</td>
<td>+0.03 ± 0.01</td>
</tr>
</tbody>
</table>
Table 5: Site-Elevation mean seasonal lapse rates of Maximum, Minimum and Mean Temperature and Diurnal Temperature Range along with temperature lapse rate estimated through radiosonde launched at Kabul Airport from 1966 to 2016 (decade of the 1990s and some years from early 2000s are missing in this data) in both morning (0430 AM local time) and afternoon (0430 PM local time). Temperature lapse rate is estimated for the whole elevation range covered by the radiosonde, up to 15 km and up to 6 km to differentiate the change in temperature in different layers of the atmosphere. Trends significance is computed at 95% confidence interval and statistically significant trends are presented with bold letters.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Autumn</th>
<th>Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site –Elevation Lapse Rate ( \text{°C/Km} )</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( T_x )</td>
<td>-0.09</td>
<td>0</td>
<td>0.01</td>
<td>-0.13</td>
<td>-0.05</td>
</tr>
<tr>
<td>( T_n )</td>
<td>0.01</td>
<td>0.03</td>
<td>0.16</td>
<td>0.13</td>
<td>0.08</td>
</tr>
<tr>
<td>( T_m )</td>
<td>-0.04</td>
<td>0.02</td>
<td>0.08</td>
<td>0</td>
<td>0.01</td>
</tr>
<tr>
<td>DTR</td>
<td>-0.1</td>
<td>-0.02</td>
<td>-0.14</td>
<td>-0.27</td>
<td>-0.13</td>
</tr>
<tr>
<td>Lapse Rate 0430 ( \text{°C/Km} )</td>
<td>-0.49</td>
<td>-0.97</td>
<td>-1.2</td>
<td>-1.3</td>
<td>-0.9</td>
</tr>
<tr>
<td>Lapse Rate 1630 ( \text{°C/Km} )</td>
<td>-0.16</td>
<td>-0.76</td>
<td>-1.76</td>
<td>-1.27</td>
<td>-0.78</td>
</tr>
<tr>
<td>Lapse Rate 0430 ( \text{°C/Km} ) &lt; 15km</td>
<td>0.11</td>
<td>0.1</td>
<td>0.12</td>
<td>-0.03</td>
<td>0.14</td>
</tr>
<tr>
<td>Lapse Rate 1630 ( \text{°C/Km} ) &lt; 15km</td>
<td>0.72</td>
<td>0.27</td>
<td>0.16</td>
<td>0.11</td>
<td>0.33</td>
</tr>
<tr>
<td>Lapse Rate 0430 ( \text{°C/Km} ) &lt; 6km</td>
<td>0.37</td>
<td>0.68</td>
<td>1.77</td>
<td>1.57</td>
<td>0.87</td>
</tr>
<tr>
<td>Lapse Rate 1630 ( \text{°C/Km} ) &lt; 6km</td>
<td>3.36</td>
<td>0.77</td>
<td>0.15</td>
<td>0.14</td>
<td>3.03</td>
</tr>
<tr>
<td>( T_{@700hpa} ) 0430 ( \text{°C} )</td>
<td>8.54</td>
<td>13.64</td>
<td>0.95</td>
<td>8.83</td>
<td>13.23</td>
</tr>
<tr>
<td>( T_{@700hpa} ) 1630 ( \text{°C} )</td>
<td>11.95</td>
<td>2.94</td>
<td>0.35</td>
<td>1.81</td>
<td>11.58</td>
</tr>
<tr>
<td>( T_{@500hpa} ) 0430 ( \text{°C} )</td>
<td>2.12</td>
<td>9.86</td>
<td>-5.24</td>
<td>6.06</td>
<td>2.07</td>
</tr>
<tr>
<td>( T_{@500hpa} ) 1630 ( \text{°C} )</td>
<td>-0.63</td>
<td>1.06</td>
<td>-6.64</td>
<td>0.81</td>
<td>4.67</td>
</tr>
<tr>
<td>( gph_{@700hpa} ) 0430 ( \text{m} )</td>
<td>5.86</td>
<td>11.33</td>
<td>8.74</td>
<td>7.99</td>
<td>7.69</td>
</tr>
<tr>
<td>( gph_{@700hpa} ) 1630 ( \text{m} )</td>
<td>3.05</td>
<td>6.47</td>
<td>5.7</td>
<td>2.6</td>
<td>4.81</td>
</tr>
<tr>
<td>( gph_{@500hpa} ) 0430 ( \text{m} )</td>
<td>11.24</td>
<td>21.87</td>
<td>8.34</td>
<td>17.58</td>
<td>11.02</td>
</tr>
<tr>
<td>( gph_{@500hpa} ) 1630 ( \text{m} )</td>
<td>5.95</td>
<td>12.05</td>
<td>5.28</td>
<td>5.15</td>
<td>15.51</td>
</tr>
<tr>
<td>( U_{@700hpa} ) 0430 ( \text{m/s} )</td>
<td>-1.88</td>
<td>-0.5</td>
<td>3.54</td>
<td>-1.77</td>
<td>0.64</td>
</tr>
<tr>
<td>( U_{@700hpa} ) 1630 ( \text{m/s} )</td>
<td>3.27</td>
<td>-2.85</td>
<td>1.78</td>
<td>0.28</td>
<td>1.51</td>
</tr>
<tr>
<td>( U_{@500hpa} ) 0430 ( \text{m/s} )</td>
<td>-5.56</td>
<td>-2.33</td>
<td>-8.62</td>
<td>-9.5</td>
<td>-8.2</td>
</tr>
<tr>
<td>( U_{@500hpa} ) 1630 ( \text{m/s} )</td>
<td>8.56</td>
<td>7.68</td>
<td>5.54</td>
<td>11.5</td>
<td>4.07</td>
</tr>
</tbody>
</table>
Appendix

In hydrostatic equilibrium, an increase in pressure can be approximated by the integral of warming below the site [Toumi et al., 1999]. The hydrostatic relation is given by:

$$\frac{dp}{dz} = -\rho g \quad (28)$$

Where \( \rho \) is mean density of the layer of atmosphere, using ideal gas law and geopotential height it can be expressed as:

$$\frac{dp}{P} = -\frac{g}{RT} dz \quad (29)$$

As environmental lapse rate is expressed as:

$$\Gamma = -\frac{dT}{dz} \quad (30)$$

By using lapse rate, hydrostatic relation as be expressed as follows:

$$\frac{dP}{P} = \frac{g}{R} \frac{dT}{\Gamma T} \quad (31)$$

Integrating equation (31) can yield following:

$$P_z = P_o \left( \frac{T_z}{T_o} \right)^{\frac{g}{R\Gamma}} \quad (32)$$

As \( T_z = T_o - z\Gamma \), therefore:

$$P_z = P_o \left( 1 - \frac{z\Gamma}{T_o} \right)^{\frac{g}{R\Gamma}} \quad (33)$$

And geopotential height increases with increase in mean temperature of the layer as given by:

$$\phi(Z_2) = \phi(Z_1) + RTln\left(\frac{P_1}{P_2}\right) \quad (34)$$
The thickness of layer between fixed pressure surfaces increases due to increase in mean temperature of the layer.

Mechanical Production

\[
MP = -\bar{u}'w' \frac{\partial \bar{u}}{\partial z} - \bar{v}'w' \frac{\partial \bar{v}}{\partial z}
\]  (35)