Altitudinal variations of temperature, equilibrium line altitude, and accumulation-area ratio in Upper Indus Basin
Biswajit Mukhopadhyay and Asif Khan

ABSTRACT
Quantitative measures of adiabatic lapse rate, equilibrium line altitude (ELA), and accumulation-area ratio (AAR) are important to understand the hydrological processes and conduct hydrological modeling in a highly glaciated watershed. We present a detailed analysis of temperature data from 21 climatic stations, hypsometric analyses of glacier distributions, and a method to analyze ablation gradients and runoff curves concurrently to quantify these parameters for the watersheds of the Upper Indus Basin (UIB), with 15,062 km² of glacierized area and an elevation range of 361–8,611 m. We show that the ELA varies considerably from one watershed to another, implying a highly variable upper elevation limit up to which melting of snows and glaciers takes place throughout the basin. This is in sharp contrast to the assertions made by previous researchers. We show that the ELA is as low as 4,840 m in Astore watershed and it is as high as 6,200 m in Shyok watershed. In accordance with the variation of ELA, the AAR also varies considerably from one watershed to another. It is as low as 0.10 in Gilgit and as high as 0.65 in Zanskar watersheds. We ascribe 15–20% uncertainty to these estimates of ELA and AAR in UIB.

Key words | accumulation-area ratio, equilibrium line altitude, Karakoram glaciers, Upper Indus Basin, zero degree C isotherm

INTRODUCTION
In mountainous river basins, where melt water from seasonal and perennial snow and ice cover forms an integral component of the river discharge, an understanding of the hydrological processes and hydrologic modeling requires quantitative models of altitudinal variations of temperature. When such basins have significant glacierized areas, the equilibrium line altitude (ELA) is a useful parameter to depict the link between glacier mass balance and climate change. The ELA is widely used to infer present and past climatic conditions (e.g., Andrews 1975; Porter 1975, 1977) and is an important climate descriptor for a glacierized watershed’s summer air temperature and winter precipitation. Another parameter of great importance in the descriptions of hydrological processes on glaciers is the accumulation-area ratio (AAR), which is the ratio of the area of accumulation zone to the total glaciated area. This ratio is necessary to perform glacier mass balance calculations (e.g., see Appendix B in Mukhopadhyay & Khan 2015). In most cases, particularly in rugged and remote terrains, these parameters are largely unknown due to difficulties in obtaining sufficient field-based observations that lead to the estimations of these parameters. In this paper, we provide the estimates of these parameters for the Upper Indus Basin (UIB).

UIB is among the most melt water-dependent river basins worldwide. The Upper Indus River originates at an elevation of about 5,166 m above sea level (asl) in the remote region of western Tibet and flows in a general northwest direction between the Zanskar, Greater Himalayas and the Karakoram Mountains all the way to the Hindu Kush Mountains. Thus, UIB straddles two very high altitude mountain ranges that contain numerous peaks with elevations greater than...
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remarkably great. Archer (2004) has examined this problem. However, the data used in his study are mostly from the climatic stations that are located at valley floor elevations. Since then, further data have become available not only from these stations but also from several stations placed at higher elevations more recently. These datasets provide an opportunity to reexamine this problem, to better define the variation of temperature as a function of space, season, and altitude within this river basin, which in turn may help to better refine the snowmelt models of UIB developed previously (e.g., Mukhopadhyay & Dutta 2010; Tahir et al. 2011).

The ELA is best determined by careful measurements of specific balance quantities at many points on the glacier surface, so that reasonably reliable isolines of zero balance may be drawn (Braithwaite & Müller 1980). Such measurements have been carried out on relatively few glaciers in the Karakoram Range and, even then, for only a few years (e.g., Wake 1989; Hewitt 2005). In this paper, we develop an innovative approach using hydrological methods to estimate the ELA within various watersheds of UIB. We also provide the estimations of the AAR in these watersheds from estimates of glacierized areas and the ELA.

**DATA**

We have collected temperature data from the existing climatic stations maintained and operated by two official authorities of Pakistan, one in India and another in China (Table 1). Therefore, these data are from reliable sources and are of good quality as described below.

We have collected temperature data from 21 climatic stations (Table 1). The valley-based stations within the Pakistani part of the basin are maintained by the Pakistan Meteorological Department (PMD) and have a long period of record (>55 years), and the high altitude stations are maintained by the Water and Power Development Authority of Pakistan (WAPDA) for its snow and ice project, with a shorter period of record (Table 1). We have obtained daily maximum and minimum temperatures from PMD and WAPDA. PMD collects climatic data in accordance with the guidelines prescribed by the World Meteorological Organization (WMO). WAPDA has auto-recording climatic stations, and those data are collected in accordance with

6,000 m asl and several notable peaks in the range of 7,000-8,000 m asl such as Rakaposhi, Nanga Parbat, and K2 (Mount Godwin Austen). According to the elevation data derived from the Shuttle Radar Topography Mission (SRTM), launched by the National Aeronautics and Space Administration (NASA) in 2005, the elevation in the basin ranges from 361 m at the basin outlet (the Tarbela reservoir intake) to 8,569 m asl (Khan et al. 2014) with a mean elevation of 4,654 m asl. The highest point in the basin is the peak of Mount Godwin Austen (K2) at 8,611 m, the second highest peak in the world (the difference is due to errors in the SRTM data; see Mukhopadhyay & Singh 2011). The mean elevation refers to the elevation from where half of the basin area is above and the other half is below. In addition to the high altitude nature of the terrain, UIB is the abode of some remarkable glaciers such as Siachen, Masherbrum, Panmah, Baltoro, Biafo, Chogo Lungma, Batura, Hispar, and Rimo glaciers, among others, and surrounding snow covered mountain peaks and slopes. According to Bajracharya & Shrestha (2011), there are 11,415 glaciers in UIB with an area of 15,062 km². Consequently melt water from glaciers, and perennial and seasonal snow cover, forms the dominant constituent of the river flows in UIB.

Mukhopadhyay & Khan (2015) have shown that glacial melt far outweighs snowmelt in the rivers draining the Karakoram and Zanskar ranges. According to their estimates, in the Karakoram, the annual glacial melt proportion varies from 43 to 50% whereas snowmelt varies from 27 to 31%. On the other hand, snowmelt dominates over glacial melt in the rivers draining the western Greater Himalayas and the Hindu Kush. Here, the snowmelt percentage in river discharge varies from 31 to 53%, whereas that of glacial melt ranges from 16 to 30%. In the main stem of the Upper Indus River, the snowmelt fraction in most cases is slightly greater than the glacial melt fraction. In the main stem, the snowmelt percentage ranges from 35 to 44% whereas the glacial melt percentage ranges from 25 to 36%. The Upper Indus River just upstream of the Tarbela Reservoir carries annual flows constituted of 70% melt water, of which 26% is contributed by glacial melts and 44% by snowmelt.

In this paper, we present models of altitudinal variation of temperature in UIB where the variations of elevation are remarkably great. Archer (2004) has examined this problem.
WMO guidelines too. In addition, we have collected temperature data from the Indian Meteorological Department for one station (Leh) within the Indian part of the basin and also from the China Meteorological Administration for one station (Demchok) within the Tibetan part of UIB. The other datasets we have used include data related to topography, snow covered areas, glacier cover, and river flows. These data are described in Khan et al. (2014), Mukhopadhyay (2012, 2013), and Mukhopadhyay & Khan (2014a, 2014b, 2015).

### THE MAJOR WATERSHEDS OF UIB

One of the major tributaries of the Upper Indus is the Shyok River, which originates from the eastern Karakoram Range. Two other major rivers, namely the Nubra, originating from Siachen Glacier, and the Hushey, which flows from Masherbrum glacier in the Karakoram, feed the Shyok with melt water. The Zanskar River and Shingo River, both of which originate from the western ranges of the Greater Himalayas, nourish the main stem of the Indus, before its confluence with the Shyok. Downstream of the confluence of the Indus and the Shyok, the major rivers that contribute to the flows in the Indus are Shigar, carrying melt water from various glaciers in the central Karakoram; Hunza, carrying melt water from the western Karakoram; Gilgit/originating from the Hindu Kush; and Astore, originating from the Burzil Pass in the western extreme of the Greater Himalayas. Therefore, nine major watersheds constitute UIB (Figure 1). Table 2 provides the basic data on these

<table>
<thead>
<tr>
<th>Station</th>
<th>Watershed</th>
<th>Altitude (m)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Period of record</th>
<th>No. of years</th>
<th>Source*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Astore</td>
<td>Astore</td>
<td>2,168</td>
<td>35.3665</td>
<td>74.8647</td>
<td>1954–2010</td>
<td>57</td>
<td>1</td>
</tr>
<tr>
<td>Rama</td>
<td>Astore</td>
<td>3,179</td>
<td>35.4548</td>
<td>74.7765</td>
<td>1999–2010</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Rattu</td>
<td>Astore</td>
<td>2,718</td>
<td>35.1615</td>
<td>74.7852</td>
<td>1999–2010</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Burzil</td>
<td>Outside</td>
<td>4,239</td>
<td>34.8994</td>
<td>75.0791</td>
<td>1999–2010</td>
<td>12</td>
<td>2</td>
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<tr>
<td>Gilgit</td>
<td>Gilgit</td>
<td>1,460</td>
<td>35.9206</td>
<td>74.3271</td>
<td>1951–2010</td>
<td>60</td>
<td>1</td>
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<tr>
<td>Ushkore</td>
<td>Gilgit</td>
<td>3,051</td>
<td>36.0271</td>
<td>73.4148</td>
<td>1999–2010</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Yasin</td>
<td>Gilgit</td>
<td>3,280</td>
<td>36.4505</td>
<td>73.2942</td>
<td>1999–2010</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Gupis</td>
<td>Gilgit</td>
<td>2,156</td>
<td>36.1789</td>
<td>73.4386</td>
<td>1955–2010</td>
<td>56</td>
<td>1</td>
</tr>
<tr>
<td>Hunza</td>
<td>Hunza</td>
<td>2,374</td>
<td>36.3220</td>
<td>74.6461</td>
<td>2007–2010</td>
<td>4</td>
<td>1</td>
</tr>
<tr>
<td>Naltar</td>
<td>Hunza</td>
<td>2,898</td>
<td>36.1679</td>
<td>74.1753</td>
<td>1999–2012</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Khunjerab</td>
<td>Hunza</td>
<td>4,630</td>
<td>36.8119</td>
<td>75.3321</td>
<td>1999–2013</td>
<td>12</td>
<td>2</td>
</tr>
<tr>
<td>Shigar</td>
<td>Shigar</td>
<td>2,367</td>
<td>35.6982</td>
<td>75.4814</td>
<td>1996–2010</td>
<td>15</td>
<td>2</td>
</tr>
<tr>
<td>Hushey</td>
<td>Shyok</td>
<td>3,075</td>
<td>35.3424</td>
<td>76.1395</td>
<td>1994–2010</td>
<td>17</td>
<td>2</td>
</tr>
<tr>
<td>Deosai</td>
<td>Upper Indus</td>
<td>4,149</td>
<td>35.003756</td>
<td>75.592112</td>
<td>1994–2010</td>
<td>17</td>
<td>2</td>
</tr>
<tr>
<td>Bunji</td>
<td>Upper Indus</td>
<td>1,372</td>
<td>35.6462</td>
<td>74.6292</td>
<td>1953–2010</td>
<td>58</td>
<td>1</td>
</tr>
<tr>
<td>Skardu</td>
<td>Upper Indus</td>
<td>2,317</td>
<td>35.2864</td>
<td>75.5630</td>
<td>1952–2010</td>
<td>59</td>
<td>1</td>
</tr>
<tr>
<td>Leh</td>
<td>Upper Indus</td>
<td>3,505</td>
<td>34.164</td>
<td>77.587</td>
<td>1876–1990</td>
<td>106</td>
<td>3</td>
</tr>
<tr>
<td>Demchok</td>
<td>Upper Indus</td>
<td>4,278</td>
<td>32.500</td>
<td>80.080</td>
<td>1979–2010</td>
<td>51</td>
<td>4</td>
</tr>
</tbody>
</table>

*Sources: 1 — Pakistan Meteorological Department; 2 — Water and Power Development Authority, Pakistan; 3 — Indian Meteorological Department (IMD) (at Leh, data for 1970–1977 are missing); 4 — China Meteorological Administration. All stations’ daily data (except Leh and Demchok) have been obtained, and a monthly average has been estimated from these datasets.
watersheds, including the extents of glacial covers in these watersheds as determined from the Randolph Glacier Inventory (RGI Version 3.2; http://www.glims.org/RGI/; Pfeffer et al. 2014). Table 3 gives the hypsometric data for each of the watersheds.

As noted above, UIB spans three mountain ranges, namely the Greater Himalayas, the Karakoram, and the Hindu Kush. According to Winiger et al. (2005), these three mountain ranges have to be divided into different climatic regions. Accordingly, three watersheds, namely Hunza, Shigar, and Shyok, cover the western, central, and eastern parts of the Karakoram; Gilgit watershed covers the Hindu Kush and four watersheds, namely Astore, Shingo, Zanskar, and Kharmong astride the western, central, west-central, and eastern sections of the Western Greater Himalayas. The remaining watershed, called Tarbela, lies along the valley between the ridges of the Karakoram, the Western Greater Himalayas, and the Hindu Kush.

### ALTITUDINAL VARIATION OF TEMPERATURE

Figure 2 shows the monthly average temperatures ($T_{Av}$) at the existing climate stations at different elevations, based on various periods of record (see Table 1 for elevations of the stations). Station temperature data show that July is the warmest month of the year at all locations, except at Hunza where $T_{Av}$ for August is slightly greater than that of July. Regardless, for all stations, $T_{Av}$ for July is approximately equal to $T_{Av}$ for August. Therefore, the maximum elevations up to which snow and ice melting occurs in the basin are reached during July and August. Peak discharge through all of the rivers in the basin also occurs during these 2 months (see hydrographs in Mukhopadhyay & Khan 2014a).

Altitudinal variations of temperatures within different watersheds in different months are modeled from calculations of adiabatic lapse rates ($\lambda_a$) based on daily temperatures records. We have calculated monthly average adiabatic lapse rates from observed daily temperatures at...
two stations at different elevations. We have used the differences in daily maximum and minimum temperatures at two different stations for a given month of a year, and the elevation difference between these two stations, to calculate maximum and minimum lapse rates. Then we calculated monthly average values from daily values. We have selected conjugate stations to calculate the value of the adiabatic lapse rate, given as:

\[
\lambda_a = \frac{T_{s2} - T_{s1}}{Z_{s1} - Z_{s2}}
\]

where, \(T_{s2}\) and \(T_{s1}\) represent the monthly average temperatures at station 1 and 2, respectively; and \(Z_{s1}\) and \(Z_{s2}\) represent the elevations of station 1 and station 2, respectively.

The pair that produces the most reasonable results is selected as the representative value for a particular watershed. Consistency in the calculated values, optimum fluctuations from day to day or from month to month, and values oscillating around the standard environmental lapse rate of 0.00649 \(\text{W C}^{-1}\) are the criteria used to judge the values that are most reasonable and consistent.

Lapse rates are calculated for the individual months and for the individual watersheds within the basin. Table 4 provides an internally consistent set of values of \(\lambda_a\) for different months within the major watersheds of UIB.

Typically, values of \(\lambda_a\) decrease from the dry adiabatic lapse rate of 0.0098 \(\text{C}^{-1}\) depending on the humidity, and are close to the global mean lapse rate of 0.00649 \(\text{C}^{-1}\), also known as the environmental lapse rate. However, in mountainous terrains, \(\lambda_a\) can show great diurnal and seasonal variations and can depart significantly from the

### Table 2 | Major watersheds of UIB

<table>
<thead>
<tr>
<th>Watershed</th>
<th>Watershed area (km²)</th>
<th>Glacial cover RGI-3.0 (km²)*</th>
<th>Glacier extents (range in m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shyok</td>
<td>33,470</td>
<td>7,696 ± 770</td>
<td>3,221–7,803</td>
</tr>
<tr>
<td>Hunza</td>
<td>13,734</td>
<td>3,843 ± 384</td>
<td>2,299–7,809</td>
</tr>
<tr>
<td>Shigar</td>
<td>7,040</td>
<td>2,121 ± 212</td>
<td>2,716–8,569</td>
</tr>
<tr>
<td>Gilgit</td>
<td>12,713</td>
<td>836.3 ± 84</td>
<td>2,750–6,992</td>
</tr>
<tr>
<td>Zanskar</td>
<td>14,861</td>
<td>1,189 ± 119</td>
<td>3,927–6,509</td>
</tr>
<tr>
<td>Shingo</td>
<td>10,502</td>
<td>733 ± 73</td>
<td>3,579–7,103</td>
</tr>
<tr>
<td>Astore</td>
<td>3,988</td>
<td>541 ± 54</td>
<td>2,897–8,020</td>
</tr>
<tr>
<td>Kharmong (UIB)*</td>
<td>44,667</td>
<td>601 ± 60</td>
<td>4,438–6,507</td>
</tr>
<tr>
<td>Tarbela (UIB)**</td>
<td>26,053</td>
<td>776 ± 78</td>
<td>2,602–8,069</td>
</tr>
</tbody>
</table>

*Kharmong (UIB): Watershed draining the headwater catchments of Upper Indus River and those along its course up to Kharmong (before its confluence with Shyok, Zanskar and Dras Nala watersheds are excluded).

**Tarbela (UIB): Watershed draining the catchments of the Upper Indus River along its course from Kharmong up to the point of inflow into the Tarbela Reservoir (all major tributary watersheds are excluded).

Data from the Randolph Glacier Inventory version 3.0, released in 2013 (http://www.glims.org/RGI/). These areas are estimated for glaciers only. The accuracy is assumed to be 90% (error bars at 10%).

The extents of glaciers in the elevation range have been derived from the SRTM DEM by overlaying the RGI 3.0 glacier maps on it.

### Table 3 | Percentages of surface areas in five elevation zones (hypsometric bands) for the nine major watersheds of UIB

<p>| Hypsometric bands (areas are in percentages of total areas) |
|---------------------------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|</p>
<table>
<thead>
<tr>
<th>Watershed</th>
<th>&lt; 1,000 m</th>
<th>1,000–2,500 m</th>
<th>2,500–3,500 m</th>
<th>3,500–5,300 m</th>
<th>5,300–6,500 m</th>
<th>&gt; 6,500 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tarbela</td>
<td>0.15</td>
<td>2.89</td>
<td>10.36</td>
<td>64.57</td>
<td>21.93</td>
<td>0.1</td>
</tr>
<tr>
<td>Shyok</td>
<td>0</td>
<td>0.84</td>
<td>3.73</td>
<td>44.73</td>
<td>50.2</td>
<td>0.5</td>
</tr>
<tr>
<td>Hunza</td>
<td>0</td>
<td>3.64</td>
<td>16.75</td>
<td>69.9</td>
<td>5.82</td>
<td>3.89</td>
</tr>
<tr>
<td>Shigar</td>
<td>0</td>
<td>4.26</td>
<td>8.52</td>
<td>65.34</td>
<td>19.89</td>
<td>1.99</td>
</tr>
<tr>
<td>Gilgit</td>
<td>0</td>
<td>4.64</td>
<td>14.24</td>
<td>78.58</td>
<td>2.44</td>
<td>0.1</td>
</tr>
<tr>
<td>Zanskar</td>
<td>0</td>
<td>0</td>
<td>1.48</td>
<td>69.98</td>
<td>28.26</td>
<td>0.28</td>
</tr>
<tr>
<td>Shingo</td>
<td>0</td>
<td>0</td>
<td>8.57</td>
<td>87.13</td>
<td>2.66</td>
<td>1.64</td>
</tr>
<tr>
<td>Astore</td>
<td>0</td>
<td>2.51</td>
<td>18.8</td>
<td>76.48</td>
<td>1.25</td>
<td>0.96</td>
</tr>
<tr>
<td>Kharmong (UIB)*</td>
<td>0</td>
<td>0.04</td>
<td>2.19</td>
<td>72.77</td>
<td>24.85</td>
<td>0.15</td>
</tr>
</tbody>
</table>

*Kharmong (UIB): Watershed draining the headwater catchments of the Upper Indus River and those along its course up to Kharmong (before its confluence with Shyok, Zanskar and Shingo watersheds are excluded).
standard environmental lapse rate. For example, Archer (2003) has noted that within the northwestern part of UIB, the value of $\lambda_a$ ranges from 0.0065–0.0075 °C m$^{-1}$. Jain et al. (2008) observed $\lambda_a$ to vary from 0.0060–0.0074 °C m$^{-1}$ in the Sutlej River basin of the Western Himalayas. Singh (1991) observed that in the Western Himalayas, calculated
values of $\lambda_a$ also depend on the elevations of the stations from which measured temperatures are used in the estimation of $\lambda_a$. He observed a decrease in values of $\lambda_a$ with increasing station elevations. Kattel et al. (2015) show that the annual cycle of the $\lambda_a$ on the northern and southern slopes of the Himalayas exhibits distinct seasonal patterns controlled by the elevation-dependent moisture content of the atmosphere and solar radiation.

It should be noted at this point that lapse rates may not remain constant with time, given the evidence for elevation-dependent warming (EDW) with more rapid rise, historical and predicted, at higher elevations (Qin et al. 2009; Mountain Research Initiative (MRI) EDW Working Group 2015). MRI (2015) notes, for example, that over the past 20 years temperatures above 4,000 m asl have warmed nearly 75% faster than temperatures in areas below 2,000 m asl. However, Qin et al. (2009) note that ‘it is found... that the warming rate increases with respect to elevation below roughly 5,000 m ASL and such an increase tendency disappears for higher elevations (>5,000 m)’. This variation in tendency with respect to EDW may well also be reflected in lapse rates. More rapid increases in temperature at higher elevations, both historic and projected, have led or will lead to a decrease in lapse rates.

Our estimations indicate that lapse rates are lower in the Shyok watershed compared to those in the Shigar and Hunza watersheds. This is most likely, as noted above, due to a general decrease in elevation from eastern to western Karakoram. For this reason, these lapse rates, for the purpose of any hydrologic modeling, should be used relative to the elevations of the base stations used in their derivations. The highest lapse rates are in the eastern watersheds of the basin over the Western Greater Himalayas, likely due to greater aridity in this part of the basin compared to the northwestern parts (see Immerzeel et al. 2015). The lower autumn/winter lapse rate in the Shyok watershed compared to the Kharmong watershed is possibly due to the fact that the Shyok watershed straddles eastern Karakoram, where the general slope is south facing, whereas the Kharmong watershed spans the western Higher Himalayas, where the general slope is north facing. Therefore, during autumn and winter, altitudinal variation of temperature in the western Himalayas is greater than that in the eastern Karakoram. Kattel et al. (2015) have shown the control of the northern and southern slopes of the Himalayas on temperature lapse rates.

In general, within a watershed, the lapse rate increases from January to April/May and then remains fairly constant during the summer months, and then starts to decline from October with the commencement of autumn and continues to decline in the winter. In other words, the greatest altitudinal variation in temperature occurs during the snow and glacier melting season. Immerzeel et al. (2015) show the spatial variation of precipitation over UIB. According to their analysis, the highest precipitation is found along the monsoon-influenced southern Himalayan arc, while in the north of the Himalayan range the precipitation decreases quickly towards the vast dry area in the northeastern part of UIB (Shyok watershed). On the other hand, the northwestern part of the basin (the Hunza and Shigar watersheds) receives considerable amounts of precipitation at high altitude. Thayyen & Dimri (2014) demonstrate that manifestations of the presence or absence of moisture is the single most important factor determining the temperature distribution along the higher Himalayan slopes, driven by the orographic forcing. These authors also suggest that ‘Diverse slope environmental lapse rate of temperature for different altitudinal sections of monsoon and cold-arid systems of the Himalaya amply demonstrate that the use of environmental lapse rate or any other temperature lapse rate values arbitrarily for extrapolating the temperature to the Himalayan cryospheric systems is not appropriate’.

The elevation where freezing temperature (0 °C) prevails in a month is designated as the freezing level altitude (FLA). This elevation for a watershed is calculated as:

$$FLA = \frac{T_{0 C}^S}{\lambda_a} + Z_S$$

(2)

where, FLA is the elevation of the 0 °C isotherm, $\lambda_a$ is the monthly average temperature lapse (used with a positive sign) rate in that watershed, $Z_S$ is the elevation of a base station within that particular watershed, and $T_{0 C}^S$ is the monthly average temperature as measured at this base station.

Figure 3 shows the seasonal variation of the FLAs in the different watersheds of UIB. The maximum altitudinal recession of the FLA occurs in July and August. The FLA during
the month of July ranges from 5,239 to 6,263 m, and during August it ranges from 5,315 to 6,187 m.

**ESTIMATION OF ELAS**

Normally, accumulation increases and ablation decreases with increasing elevation. As defined by Cogley et al. (2011), the ELA is a spatially averaged line on the glacierized surface where accumulation equals ablation and specific mass balance is zero (Figure 4). Note that during the summer melting season, FLA > ELA (Figure 4). From Figure 4, it can be seen that ablation gradient ($\alpha_g$), a concept introduced by Haefeli (1962), is given by:

$$\alpha_g = \frac{x_m}{FLA - Z}$$  \hspace{1cm} (3)

where, $x_m$ is the ice melt water (in terms of water equivalent) produced at an elevation $Z$ (<FLA) on a glacierized surface.

The ablation gradient can be measured from detailed field studies. Unfortunately, such studies are lacking in UIB. The only glaciers for which the ablation gradient has been established from field investigation are the Biafo glacier (Hewitt et al. 1989) and the Baltoro glacier (Mihalcea et al. 2006). According to the results of the study conducted by Hewitt et al. (1989), an estimate of $\alpha_g \approx 1/0.005$ m (1V:xH), as noted by Yu et al. (2013). In the present study, we test three values of $\alpha_g$: 1 m : 0.0025 m; 1 m : 0.005 m, and 1 m : 0.01 m since it is largely an unknown parameter.

Mukhopadhyay & Khan (2014a) have developed a hydrograph separation technique to differentiate three components of river discharge within UIB. With this methodology, they have separated the base flow (which includes rain derived runoff including high altitude monsoonal precipitation in July and August), mid-altitude (M1), and high-altitude (M2) melt components of an annual hydrograph obtained at a stream gauging station. Figure 5 is a schematic representation of the general system under consideration.

Figure 6 shows the separated hydrographs of the long-term average annual hydrographs as given by Mukhopadhyay & Khan (2014a). The annual contribution of the M2 component to total annual flow volume (m$^3$) varies from 41 to 54% along the main stem of the Indus, upstream of the Himalayan foothills. The contribution of M1 melt water varies from
In the tributaries, the annual contributions of M2 vary from 37% to as high as 65%. Similarly, the annual contributions of M1 in the tributaries vary from as low as 12 to 54%. Thus, the relative importance of high altitudes far outweighs those originating from the mid altitudes in river runoff within UIB.

Here we introduce the concept of the runoff curve for the M2 component, a mixture of snow and glacial melts, given as:

$$\int_{Z_{M2}}^{Z_{f2}} q(Z) \, dZ = Q_{M2}$$  \hspace{1cm} (4)

where, $Z$ denotes elevation, $Z_{u}^{M2}$ and $Z_{l}^{M2}$ represent the upper and lower limits, respectively, of the elevation band from which M2 water is generated, $q(Z)$ is specific runoff, and $Q_{M2}$ is the total discharge ($m^3$) at the outlet of the M2 zone for a specific month. From the calculated values of $Q_{M2}$ for the month of July and August, we calculate runoff curves for a watershed. We estimate $Z_{u}^{M2}$ and $Z_{l}^{M2}$ by examining the distributions of the glacierized areas as a function of elevation within various watersheds of UIB and the calculated values of the FLA for the months of July and August, as discussed above. We have determined the altitudinal distributions of glaciers by simultaneous analyses of the maps of the glacierized surfaces prepared from the RGI V3.0 database and the DEM prepared from the SRTM data. We define the altitudinal zone of M2 as the elevation range, stretching from the third band of glacierized area defined in the glacier hypsometry and FLA (Figure 7). We discard the first two bands of glacier hypsometry because the percentages of snow and ice covered areas that lay in these two bands are very low and when uncertainties in the RGI database (estimated to be 10%) and the SRTM data are considered, particularly for delineation of the lower limits of glacierized surfaces from Landsat images, these two lower bands become insignificant.

Figure 8 shows the ablation curve calculated by assuming a value of $a_{g}$ as discussed above, and the runoff curve calculated for the M2 zone of the Astore watershed, as an example. Let at any elevation the amount of runoff generated be denoted by $q_{x}$ and the amount of melt water produced be denoted by $x_{m}$. Mostly due to loss in the ground and due to evaporation, $q_{x} < x_{m}$. Let an ablation curve and the runoff curve intersect at elevation $Z_{E}$ (see Figure 8). The following relationships are noted:

for, $Z > Z_{E}$, $q_{x} > x_{m}$, which is not possible, but for $Z < Z_{E}$, $q_{x} < x_{m}$, and at $Z = Z_{E}$, $q_{x} = x_{m}$.

Therefore, only up to elevation $Z_{E}$ is melt produced that contributes to runoff. In other words, $Z_{E}$ closely represents the upper limit of the ablation zone. Therefore, we set

$$ELA = Z_{E}$$  \hspace{1cm} (5)

It should be pointed out that the linear ablation gradient, as obtained from application of Equation (3), is an idealized representation. In reality, the gradient may have a curvature in the lower elevation range due to debris cover over the tongues of the low and extended glaciers. Since the section of the curve that bears importance in the analysis presented above is in the higher elevation range, the curvature in a real ablation curve introduces little error in the calculations presented above. Furthermore, Kaab et al. (2012), Gardelle et al. (2012, 2013), Mihalcea et al. (2006) and Minora et al. (2015) have noticed nearly the same melting rate for debris-covered ice compared to clean ice. Therefore, in this study we have ignored the effect of different classes of glaciers, but have used the same melting rates for both debris-covered ice and clean ice areas.
We have calculated the ablation and runoff curves for each of the major watersheds of UIB (Figure 1) to derive the ELA within the respective watersheds for both of the months of July and August. Table 5 provides the results of the calculations. The lower and upper limits of the ELA for a month correspond to the lower and upper limits of the values of $\alpha_g$ selected (which are 1:0.0025 m and 1:0.01 m, respectively). From the calculations of the ELA for the months of July and August, we take the averages of the lower and upper limits of the ELA as given in Table 5.

Different researchers have also estimated various values of ELA in this terrain (Table 6) using other methods. For example, Gardelle et al. (2013) used approximately 2000 Landsat images to estimate the ELA in the Karakoram and Hindu Kush mountains. Our calculations are in general agreement with the values listed in Table 6. However, our estimates of ELA for each of the individual watersheds within UIB show that with the exception of the Shyok watershed, the ELA is in the general range of 5,000–5,500 m. In the case of the Shyok watershed, the ELA is in the range of 6,000–6,200 m. This higher range of ELA in this watershed, compared to the other watersheds of the basin, can be ascribed to the greater aridity of this watershed compared to the other watersheds of the basin (Immerzeel et al. 2015, Figure 5).

Figure 6 | Separated hydrographs showing M1, M2 and base flow components and total annual hydrographs developed from the long-term averages of the flow records at the gauging stations near the outlet of the watersheds (for locations of the gauging stations and period of records, see Mukhopadhyay & Khan 2014a). The M2 peaks of July and August are used to calculate the runoff gradients. (Continued.)
ESTIMATION OF AARS

We have extracted AARs for various watersheds, using lower and upper limits of ELAs, RGI v 3.0 glacier inventory, and the SRTM DEM. First, we have selected the lower limit of the ELA in a watershed and estimated the glacier area above it, using RGI and DEM data, and then divided it by the inventory-based total glacierized area of the same watershed. This computation has resulted in defining the upper limit of AAR. Similarly, we have used the upper limit of the ELA to estimate the lower limit of the AAR. Table 7 gives the list of AAR values within the individual watersheds.

DISCUSSION

In this paper we have developed an innovative approach using hydrological methods to estimate the ELA for the principal watersheds of UIB. We have developed this method after establishing the FLA in the watersheds from calculations of adiabatic lapse rates and using the concepts of the ablation gradient and runoff curve. Due to the presence of intrinsic uncertainties in the input data, we ascribe 15–20% errors to the estimates provided in this study. No analytical expression can be developed linking the independent and dependent variables. Therefore, straightforward application of an error propagation formula cannot be
applied to estimate these errors directly. For this reason, these error estimates are approximations based on assumed errors in the input variables that include topographic data from the SRTM, the glacierized areas from RGI database, temperature estimates, and so on.

The estimations of the ELAs and AARs, presented in this study, can be used in improved hydrological modelling and glacier mass balance studies in UIB. Due to the absence of regional ELA estimates in the UIB, some of the previous studies have adopted ELA estimates from other western Himalayan regions for glacial mass balance analysis and future forecasts of climate change impact on the Karakoram glaciers (e.g., Chaturvedi et al. 2014). However, our estimations show that Karakoram, the Hindu Kush and the Himalayas have different ELA and AAR values. Therefore, use of the estimates given in the present study can be useful for estimating uncertainties in future climate impact analysis due to variability in ELAs and AARs. Using the ELA and AAR values provided in the current study will provide more improved and accurate results in future glacier mass balance studies and hydrologic modeling to make predictions on water availability within UIB.

The ablation gradient and runoff curve have inverse relationships in an $X$-$Z$ space. The ablation gradient is a theoretical curve that is mostly a rate function of temperature that increases with decreasing elevation. The runoff curve, on the other hand, is the curve that is based on the observed runoff amount, which increases with increasing elevation. Greater runoff is generated at higher elevation because of two principal reasons. First, the glacierized

![Figure 7](image-url)  
Figure 7 | Hypsometric data for distribution of glacierized areas within the individual watersheds of UIB. Note that the cumulative glacierized area increases with elevation. (Continued.)
surface within a watershed increases with elevation (Figure 7). Secondly, even though a principal form of energy that causes snow and ice melting is atmospheric temperature, it is a poor proxy to the total energy budget on snow and glacier surfaces due to its decreasing strength with elevation. Kattel et al. (2015) have shown that a considerable amount of solar radiation at higher elevations causes the atmospheric warming in the Himalayas. Therefore, solar radiation is the prime source of energy to drive the melting process at elevations < the FLA. In the following discussion, we show that in general solar radiation increases with elevation in UIB.

Dubayah & Rich (1995) and Fu & Rich (2002) have developed geometric solar radiation models by taking into account variability in elevation, slope, slope orientation (azimuth or aspect), and shadowing to calculate the amount and gradients in solar radiation incident at a location on the Earth's surface. We have selected a set of points on each of the major glaciers such that each glacier is represented by two points, one located near the lower limit (elevation ∼3,000–3,500 m) and the other located near the upper end (elevation ∼4,500–6,000 m) of the general zone of ablation. Subsequently we have calculated values of solar radiation (insolation) at each of these points according to the geometrical solar radiation model (Dubayah & Rich 1995; Fu & Rich 2002). Altitudinal variations of insolation ($R_s$) within two example watersheds are shown in Figure 9. Note that, even though there are...
considerable scatters in the plots because $R_s$ greatly depends on aspect and slope, in general it increases with elevation. Thus, an increased glacierized surface combined with increased energy inputs at higher elevations causes the total volume of melt water generated at higher elevations to be greater than that generated at relatively lower elevations.

Our calculations show that the ELA varies considerably from one watershed to another within UIB. This in turn implies that the upper elevation limit up to which melting of snows and glaciers takes place is not constant throughout the basin. This is in sharp contrast to the assertions made by previous researchers. For example, Young & Hewitt (1993) have considered 4,800 m as the upper elevation limit of the zone of melt production. On the other hand, Hewitt (2005) has considered this elevation to be 4,650 m. Yu et al. (2010) have set 5,000 m as the highest elevation from which melt water originates to supply to river runoff within UIB. We show that the upper elevation limit of production of M2 melt water can be as low as 4,840 m in the Astore watershed to as high as 6,200 m in the Shyok watershed. In the western and central Karakoram, this elevation ranges from 5,080 to 5,480 m. Thus, in general, the ELA is in the range of ~5,000–5,500 m.
In accordance with the variation of ELA, the AAR also varies considerably from one watershed to another. This finding has a significant bearing on the determination of loss or gain of glacial storage from the slopes of the trend lines of August flows (see Mukhopadhyay & Khan 2014b; Mukhopadhyay et al. 2015). For example, Mukhopadhyay & Khan (2014b) assumed that 30–40% of the glacierized area of the Shigar watershed belongs to the zone of ablation. Our present estimates show that 57–70% of the glacierized area of this watershed is in the zone of ablation. Therefore, the loss of glacier storage in Shigar watershed is actually much higher than the estimates given by Mukhopadhyay & Khan (2014b). Our calculations indicate that in general, the AAR in UIB can range from as low as 0.10 ± 0.04 in Gilgit watershed to as high as 0.65 ± 0.07 in Zanskar watershed.

The variations in the ELA and AAR are essentially due to the spatial variations of the FLA during the summer ablation season. These variations are principally caused by variations of altitude (terrain characteristics) and differential atmospheric moisture availability in different parts of the river basin.

CONCLUSIONS

Adiabatic lapse rates, ELA, and accumulation-area ratio are three important parameters for an understanding of the hydrological processes and hydrological modeling of mountainous watersheds with a significant presence of glaciers. UIB is such a river basin, where elevation and glacierized areas are highly significant. Thus far, quantitative measures of these parameters of this river basin have been sketchy or lacking. Here we provide a detailed analysis of temperature data from 21 climatic stations, hypsometric analyses of glacier distributions, and a new hydrological method to analyze the ablation gradient and the runoff curve concurrently to quantify these parameters for the constituent watersheds of the basin.

The 0 °C isotherm or FLA in all of the watersheds is at its lowest elevation in January (1,896–2,735 m). It starts to recede to higher elevations from February and attains the maximum elevation in July in all watersheds except Astore, where the maximum elevation is attained in August. The maximum FLA ranges from 5,239 to 6,263 m. After attainment of maximum elevation in either July or August, the FLA starts to descend to lower elevations with the onset of autumn. Accordingly, there is a great spatial variation of FLA in the basin, possibly controlled by the humidity and terrain characteristics in the different parts of the basin.

Peak discharge in the rivers occurs in July and August when the ablation zone attains its maximum extent. The elevation of the upper limit of the ablation zone or ELA during July and August ranges from 4,840 to 6,200 m. Thus, there is a great altitudinal range within the basin up to which melting of snows and glaciers occurs. This is in contrast to the assumption used in several previous investigations (e.g., Young & Hewitt 1993; Yu et al. 2013) where the upper limit of the zone of ablation has been considered to be in the range of 4,800–5,000 m. The fraction of glacierized area that is subject to ablation also varies greatly from one watershed to another as, shown by the variation in the AAR. In general, the AAR ranges from 0.27 to 0.65.
However, in two watersheds, namely Gilgit and Shyok, the AAR is relatively low.

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