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# The sustainability of water resources in the High Mountain Asia in the context of recent and future glacier change

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## Abstract

High Mountain Asia contains the largest volume of glacier ice outside the Polar regions, and contain the headwaters of some of the largest rivers in central Asia. These glaciers are losing mass at a mean rate of between  $-0.18 \pm 0.04$  m and  $-0.5$  m water equivalent per year. While glaciers in the Himalaya are generally shrinking, those in the Karakoram have experienced a slight mass gain. Both changes have occurred in response to rising air temperatures due to Northern Hemisphere climatic change. In the Westerly influenced Indus catchment, glacier meltwater makes up a large proportion of the hydrological budget, and loss of glacier mass will ultimately lead to a decrease in water supplies. In the monsoon-influenced Ganges and Brahmaputra catchments, the contribution of glacial meltwater is relatively small compared to the Indus, and the decrease in annual water supplies will be less dramatic. Therefore, enhanced glacier melt will increase river flows until the middle of the 21<sup>st</sup> Century, but in the longer-term into the latter part of this century, river flows will decline as glaciers shrink. Declining meltwater supplies may be compensated by increases in precipitation, but this could exacerbate the risk of flooding.

## 1. Introduction

Millions of people rely on glaciers in the Himalaya, Karakoram and Hindu Kush mountains, collectively referred to as High Mountain Asia, as a water resource. These glaciers form the headwaters of the largest rivers in Asia, including the Indus, the Ganges and the Brahmaputra Rivers (Fig. 1) and as such the mountains are often referred to as the ‘water towers of Asia’ (Immerzeel *et al.* 2010). High Mountain Asia contains the largest glacierised area outside the

39 Polar regions (Bolch *et al.* 2012) and glaciers here are highly sensitive to climate change  
40 (Solomina *et al.* 2016). Glaciers in the Himalaya are predominantly shrinking (Kääb *et al.*  
41 2012), and rates of glacier mass loss, although spatially variable, have accelerated since the  
42 1990s (Bolch *et al.* 2012). If a constant rate of glacier mass loss after 1975 is assumed, then  
43 predictions indicate loss of 10–30 m water equivalent (w.e.) per year between 2010 and 2035,  
44 which is sufficient to result in the disappearance of many smaller glaciers across the  
45 mountain range (Cogley 2011). The catchments supplied by rivers draining High Mountain  
46 Asia are located in developing countries that use this water primarily for agriculture and  
47 hydroelectric power generation, and are extremely vulnerable to changes in their water  
48 supply (Kaser *et al.* 2010; Pritchard, 2017). Predictions are needed of how Asian water  
49 supplies are likely to change due to continued glacier mass loss in response to recent and  
50 future climate change. We therefore need to improve understanding of both the contribution  
51 of glaciers to the hydrological budgets of these large catchments, and discover how these  
52 glaciers are responding to recent and future climate change (e.g. Lutz *et al.* 2014; Brun *et al.*  
53 2017).

54

55 Although only 1–3% of the area of the Indus, Ganges and Brahmaputra catchments are  
56 glacierised, these densely-populated catchments rely on glacial meltwater (Immerzeel *et al.*  
57 2010). The contribution of glaciers to runoff varies regionally; from 18.8% in the Dudh  
58 Koshi catchment, which is a major tributary of the Ganges, up to 80.6% in the Hunza  
59 catchment that drains into the Indus (Lutz *et al.* 2014). The Indus and the Ganges provide  
60 important water supplies that are used to irrigate over 140,000 km<sup>2</sup> of agricultural land, and  
61 the largest irrigation network in the world is contained in the Indus catchment (Immerzeel *et al.*  
62 2010). In particular, the importance of glacier meltwater relative to other water sources  
63 (e.g. precipitation, snow melt, groundwater) for regional hydrological budgets has only  
64 recently been documented (Immerzeel *et al.* 2010, Lutz *et al.* 2014). In the monsoon-  
65 influenced Central and Eastern Himalaya the majority of annual precipitation occurs during  
66 the warm summer monsoon months (June to September) (Bookhagen & Burbank 2010). The  
67 high summer rainfall and snowfall roughly coincide with the timing of the majority of glacier  
68 ablation (Benn & Lehmkuhl 2000), so that the relative contribution of glacial melt to river  
69 flows is minimised compared to regions to the west where summers are generally dry (Kaser  
70 *et al.* 2010). In the Western Himalaya, Karakoram and Hindu Kush mountains, where the  
71 majority of precipitation occurs as winter snowfall, glacier melt plays a much more important  
72 role in regulating seasonal river flows, with a relatively larger proportion of the annual flow  
73 in the Indus resulting directly from glacier melt compared to that in the Ganges and  
74 Brahmaputra catchments (Lutz *et al.* 2014).

75

76 The impact of climate change on these vital river flows from the Himalaya is, however, not  
77 straightforward. For example, climatic warming may cause glaciers to lose mass and release a  
78 greater volume of meltwater each year, but may also result in increased orographic  
79 precipitation that could sustain or enhance flows and trigger a gradual or abrupt change in the  
80 seasonality of peak flows (Immerzeel *et al.* 2010). Understanding regional and catchment-  
81 scale hydrological budgets and predicting how they will vary under a changing climate  
82 therefore requires coupling our understanding of glaciers processes with climate model  
83 forecasts. The Intergovernmental Panel on Climate Change (IPCC) climate scenarios for the  
84 21<sup>st</sup> Century are used for this purpose, as they are compiled from comparison of an ensemble  
85 of state-of-the-art climate model outputs (Collins *et al.* 2013). Many regional hydrological  
86 models contain large uncertainties as they do not capture the processes that affect individual  
87 glaciers, and so detailed catchment-scale models validated by field data are also required to  
88 better constrain future hydrological changes (e.g. Ragetli *et al.* 2016).

89

90 Here we review recent and predicted glacier and hydrological change across High Mountain  
91 Asia (Fig. 1). Glaciers on the Tibetan Plateau are excluded from this review, as this region is  
92 strongly influenced by weather systems originating from the Arctic rather than the Asian  
93 monsoon, and show markedly different behaviour compared to these glaciers. The long-term  
94 response to climate change of Tibetan glaciers is described by Owen *et al.* (2008) and Owen  
95 (2009). We first summarise the current knowledge of the state of glaciers in High Mountain  
96 Asia, then discuss these changes in the context of observed longer-term glacier change since  
97 the late Holocene (the last 2,000 years). We then compare predictions of glacier change with  
98 current and future climate change and consider the likely impacts of these changes on  
99 regional hydrological budgets.

100

## 101 **2. The current state of glaciers in High Mountain Asia**

102 Glaciers in High Mountain Asia are discussed in terms of their location in sub-regional areas  
103 that are defined from the major regional climate controls (after Bolch *et al.* 2012; Fig. 1).  
104 From east to west these regions are: the monsoon-influenced Eastern and Central Himalaya,  
105 and the mid-latitude Western Himalaya, Karakoram and Hindu Kush ranges influenced by  
106 westerly weather systems. These areas follow the boundaries of the major river catchments  
107 within the high mountains, with the Western Himalaya, Karakoram and Hindu Kush draining  
108 into the Indus, the Central Himalaya into the Ganges, and the Eastern Himalaya and some of  
109 the Tibetan Plateau forming the headwaters of the Brahmaputra River.

110

### 111 **2.1 Glacier extent and volume**

112 The Himalaya and Karakoram mountains contain 32,353 glaciers with a total glacierised area  
113 of about 41,000 km<sup>2</sup> equivalent to 6% of the global glacierised area (Arendt *et al.* 2015).

114 Until recently, relatively little was known about the total number and size of glaciers in the  
115 Himalaya because perennial snow cover, debris-covered ice and ice-cored moraine impeded  
116 identification of glacier outlines from satellite observations. Improvements in satellite remote  
117 sensing imagery have allowed identification of the majority of glacier outlines, which are  
118 compiled in the 6<sup>th</sup> Randolph Glacier Inventory (released in July 2017) and cover most of  
119 High Mountain Asia (Arendt *et al.* 2015). Global glacier inventories comprising glacier area  
120 boundaries drawn by the glaciological community are now sufficiently complete to estimate  
121 the glacierised extents, but data describing other important characteristics such as ice  
122 thickness are highly spatially variable and are limited by the small number of field  
123 observations. Glacier volume is more difficult to measure than area, as ice thickness is also  
124 unknown for most of the range. Estimated mean ice thickness for all glaciers in the Global  
125 Land Ice Measurements from Space database are low compared to typical values from  
126 individual glaciers derived from field data. Mean ice thickness is estimated to be 86 m for the  
127 Himalaya, and 172 m for the Karakoram (<http://glims.colorado.edu/glacierdata/> accessed on  
128 30/09/16), although the uncertainty associated with these values is undoubtedly large (Cogley  
129 2011; Frey *et al.* 2014) and likely biased by the majority of measurements being obtained  
130 from smaller glaciers.

131

132 More robust *in situ* ice thickness measurements have only been made for a handful of  
133 glaciers, using ground-penetrating radar or radio-echo sounding surveys. Access to large  
134 high-altitude glaciers can be challenging and so field observations are frequently made at  
135 more accessible glaciers. Accessible glaciers are generally at lower altitudes, smaller than the  
136 majority of the population, and have higher rates of mass loss than larger glaciers at higher  
137 altitude. These glaciers are not necessarily representative of regional-scale behaviour, and  
138 therefore field measurements often contain a bias that may skew understanding of regional  
139 mass balance trends (Fujita & Nuimura 2011). Ice thicknesses for three glaciers in Nepal in  
140 the Central Himalaya ranged from less than 20 m near the terminus to 440 m near the icefall for  
141 the largest, Khumbu Glacier (glacier area is 39.5 km<sup>2</sup>; Gades *et al.* 2000), 20–157 m for  
142 Lirung Glacier (13.5 km<sup>2</sup>; Kadota *et al.* 1997) to 51–86 m for the smallest, Glacier AX010  
143 (0.6 km<sup>2</sup>; Kadota *et al.* 1997). Ice thickness was 124–270 m for Chhota Shigri Glacier (15.7  
144 km<sup>2</sup>) in the Western Himalaya (Azam *et al.* 2012). These values represent the centreline ice  
145 thickness, and in each case the glacier cross-section thins towards the valley sides (*cf.* Azam  
146 *et al.* 2012).

147

148 Across the Himalaya, 14–18% of the glacierised area is debris covered (Kääb *et al.* 2012), the  
149 extent of which increases to the east to reach 36% in the Everest region of Nepal (Thakuri *et al.*  
150 *et al.* 2014). Supraglacial debris thickness typically increases down-glacier, as englacial and  
151 supraglacial debris transport concentrates sediment previously incorporated into the ice (Fig.

152 2) (Rowan *et al.* 2015). The thickness of supraglacial debris layers can exceed several metres  
153 (Nicholson & Benn 2013, Thakuri *et al.* 2014; Soncini *et al.* 2016). Supraglacial debris  
154 modifies glacier mass balance, either through enhancing ablation due to the decreased albedo  
155 of debris compared to ice, or by reducing ablation by insulating the glacier surface; feedbacks  
156 which are largely dependent on the thickness of the debris layer (Østrem 1959, Evatt *et al.*  
157 2015, Nicholson & Benn 2006). The threshold between the enhancement or attenuation of  
158 ablation by supraglacial debris occurs at a critical thickness of around 0.05 m, as  
159 demonstrated both from field (Rounce *et al.* 2015) and laboratory measurements  
160 (Reznichenko *et al.* 2010). The influence of variations in debris thickness on ablation was  
161 demonstrated at Khumbu Glacier, where rates of surface lowering are highest mid-glacier just  
162 below the icefall where the surface is either debris free or only thinly mantled compared to  
163 the heavily debris-mantled terminus where ablation rates are an order of magnitude lower  
164 (Nakawo *et al.* 1999, Adhikari & Huybrechts 2009, Owen *et al.* 2009).

165

## 166 **2.2 State of glacier mass balance and their Equilibrium Line Altitudes**

167 Glacier mass balance is highly variable across High Mountain Asia. This variability in mass  
168 balance has been identified directly using the traditional glaciological method where stakes  
169 are inserted into the glacier surface to measure ice ablation and snow accumulation (e.g.  
170 Soncini *et al.* 2016). Equilibrium Line Altitude (ELA; the point on a glacier at which  
171 accumulation and ablation are balanced) can be estimated from the areal extent and  
172 hypsometry combined with climate data (Benn & Lehmkuhl 2000). The ELA method has the  
173 advantage of allowing reconstructions of past glacier mass balance from geological evidence  
174 of glacier extent (Benn *et al.* 2005, Owen & Benn 2005). Regional glacier mass balance can  
175 be estimated from geodetic methods using multi-temporal satellite imagery that measures  
176 changes in glacier surface elevations (e.g. Bolch *et al.* 2011, Kääb *et al.* 2012). Remote  
177 sensing can also be used to measure snowline altitudes at the end of the ablation season, from  
178 which the ELA can be derived (e.g. Harper & Humphrey 2003). Mass balance is more  
179 difficult to measure for debris-covered glaciers than for clean-ice glaciers due to the rapid  
180 variations in rates of ablation that occur across the glacier surface, influenced by the thermal  
181 conductivity of the debris layer, which is controlled by factors including debris thickness,  
182 debris grain size, porosity and water content (Benn *et al.* 2012).

183

184 Mass balance has only been measured directly for a small number of glaciers (Fig. 3), and the  
185 longest of the continuous records cover only 10 years. Mass balance for Shaune Garang  
186 Glacier in the Himachal Pradesh in northern India was  $-0.36 \text{ m w.e. a}^{-1}$  between 1981 and  
187 1991 (Pratap *et al.* 2015). Measurements of mass balance for small glaciers in the Central  
188 Himalaya indicate the extreme sensitivity of the monsoon-influenced glaciers to air  
189 temperature. Mass balance measurements through the 1978 monsoon for Glacier AX010

190 indicate that a 0.5°C decrease in mean summer air temperature would result in a transition  
191 between positive and negative mass balance (Ageta *et al.* 1980). Three annual ablation stake  
192 surveys indicate a mass balance of  $-1.6 \text{ m w.e. a}^{-1}$  between 2003 and 2014 for Gangju La  
193 Glacier, a small clean-ice glacier in Bhutan in the Eastern Himalaya (Tshering & Fujita  
194 2016). Mass balance modelling for the partially debris-covered Langtang Glacier in the  
195 Central Himalaya simulated a mass balance of  $-0.11 \text{ m w.e. a}^{-1}$  between 1987 and 1997  
196 (Sharma & Owen 1996). Mean present-day ELA calculated from snowline elevations in the  
197 Annapurna region of western Nepal was  $\sim 5050 \text{ m}$  (Harper & Humphrey 2003) and in eastern  
198 Nepal the ELA ranged from 5300 m in the Langtang Valley to 5600 m in the Khumbu Valley  
199 (Kayastha & Harrison 2008).

200

201 The complete mass budget of all glaciers in High Mountain Asia between 2000 and 2016 was  
202 recently calculated from remote topographic measurements as  $-0.18 \pm 0.04 \text{ m w.e. a}^{-1}$  (Brun  
203 *et al.* 2017). This value is slightly lower than that given by glaciological mass balance records  
204 (summarised by Bolch *et al.* 2012) which gave a regional mass budget of about  $-0.3 \text{ m w.e. a}^{-1}$   
205 <sup>1</sup> between the 1960s and 1990s, becoming increasingly more negative during the last two  
206 decades and similar to the global mean (around  $-0.5 \text{ m w.e. a}^{-1}$ ). Remote sensing studies have  
207 previously indicated a slightly more negative regional mass balance between 2003 and 2008  
208 of  $-0.21 \pm 0.05 \text{ m w.e. a}^{-1}$ , lower than the global average due to the slightly positive mass  
209 budget in the Karakoram (Kääb *et al.* 2012)—the so-called ‘Karakoram anomaly’ (Gardelle  
210 *et al.* 2012). Karakoram glaciers have recently gained mass due to rising air temperatures  
211 delivering more winter snowfall from the Arabian Gulf (Kapnick *et al.* 2014). A large  
212 proportion of surge-type glaciers are found in the Karakoram, and this dynamic behavior can  
213 also result in short-term mass gain (Quincey *et al.* 2011; Quincey *et al.* 2015).

214

### 215 **2.3 The “debris-cover anomaly”**

216 The ablation areas of many glaciers in the Himalaya are covered with rock debris, which is  
217 deposited on glacier surfaces as a result of erosion and mass wasting of the surrounding  
218 landscape. Supraglacial debris affects mass balance and complicates understanding of the  
219 response of these debris-covered glaciers to climate change (Scherler *et al.* 2011). There are  
220 four main sources of debris on the surface of Himalayan glaciers: (1) rockfall debris which is  
221 angular in character; (2) mixed rock- and ice-avalanche debris, which is texturally similar,  
222 but which is entrained as prominent debris layers within the glacier (Fig. 2d); (3) material  
223 resulting from collapse from over-steepened moraines which is characterised by sandy  
224 boulder gravel and is typically sub-rounded to angular; (4) debris derived from the base of the  
225 glacier that has been transported to the surface by thrusting or shear from the bed, such as at  
226 the base of an icefall. This debris is a mixture of silt, sand and gravel, with some boulders  
227 bearing striations. These four lithofacies become intimately mixed on the surface of debris-

228 covered glaciers due to local slope movements from uneven ablation (Figure 2c) (Hambrey *et*  
229 *al.* 2008). Many Himalayan glaciers are also bounded by prominent latero-terminal moraine  
230 systems. These moraines are comprised of a mixture of basally worked and rockfall debris,  
231 which texturally are typically sandy boulder-gravels. Downwasting of the glaciers since the  
232 Little Ice Age (LIA) about 1300–1600 (Rowan, 2017) have left ice-cored moraines up to a  
233 hundred metres above the glacier surface, which result in an unstable inner moraine face that  
234 is unstable and prone to collapse (Hambrey *et al.* 2008).

235

236 Debris-covered Himalayan glaciers tend to lose mass by surface lowering rather than  
237 terminus recession (Rowan *et al.* 2015, Quincey *et al.* 2009, Bolch *et al.* 2011). Surface  
238 lowering causes these debris-covered glaciers to develop very low or even reversed long-  
239 profile topographic gradients through their ablation areas, which promotes the formation of  
240 supraglacial water bodies. These ponds and lakes influence the seasonal transport of water  
241 through the glacial system, and can expand and coalesce to form substantial supraglacial or  
242 proglacial, moraine-dammed lakes that may eventually pose a potential flood hazard (Watson  
243 *et al.* 2016, Thompson *et al.* 2012). Such features are commonly bordered by steep, debris-  
244 free ice cliffs, which progressively backwaste, and, if connected to a supraglacial pond or  
245 lake, may undergo thermoerosional notch development at the ice-water interface, promoting  
246 the onset of calving processes (Fig. 2) (Hambrey *et al.* 2008, Thompson *et al.* 2016).

247

248 Satellite observations of glacier mass change suggest that debris-covered glaciers in the  
249 Himalaya and Karakoram may be losing mass at the same rate as those glaciers with clean-  
250 ice (debris-free) surfaces. This ‘debris-cover anomaly’ could be due to enhancement of  
251 ablation at ice cliff faces. Although the exposure of clean ice at these ice cliffs can  
252 dramatically enhance local ablation rates (Miles *et al.* 2016, Brun *et al.* 2016, Reid & Brock  
253 2014) field observations from Changri Nup Glacier in the Everest region suggest that, despite  
254 the presence of these ablation ‘hotspots’, a continuous or near-continuous mantle of  
255 supraglacial debris reduces net ablation, such that glacier-wide mass loss is less than would  
256 be the case for an equivalent clean-ice glacier (Vincent *et al.* 2016). To fully understand the  
257 effect of ice cliffs on ablation from debris-covered glaciers, these features and their evolution  
258 need to be incorporated into glacier-wide surface energy balance modelling (e.g. Buri *et al.*  
259 2016; Brun *et al.* 2016).

260



261

### 262 **3. Changes in glacier volume during the Late Holocene**

263 Changes in the areal extent and volume of glaciers over the last 2,000 years can be inferred  
264 from moraines that indicate the position of glacier margins, historical observations made by  
265 climbing expeditions, and field and satellite measurements of glacier geometries.

266

#### 267 **3.1 Late Holocene (2,000 years ago to present)**

268 Many glaciers in High Mountain Asia have advanced and receded two or three times during  
269 the last 2,000 years in response to climate change (Owen & Dortch 2014, Murari *et al.* 2014,  
270 Rowan 2017) and followed the global trend of glacier recession and shrinkage since about  
271 1850 (Thompson *et al.* 2006). The last period of regional glacier advance was the LIA which  
272 peaked between 1300 and 1600, although glaciers remained close to their LIA limits until the  
273 20<sup>th</sup> Century (Rowan 2017). These observations are based on geochronological data for  
274 moraines compiled from studies using radiocarbon (<sup>14</sup>C) dating (e.g. Muller 1961,  
275 Röthlisberger & Geyh 1986) and terrestrial cosmogenic nuclide dating (e.g. Owen 2009,  
276 Murari *et al.* 2014). More recent applications of these techniques generally provide more  
277 accurate results due to improvements in laboratory measurement protocols. Regional glacier  
278 volume change in the Himalaya over decadal to centennial timescales occurred in response to  
279 hemispheric changes in air temperature (Solomina *et al.* 2016, Rowan 2017). However,  
280 variations in the timing and extent of glacier volume change across this range are primarily  
281 driven by millennial-scale east–west and north–south variations in atmospheric circulation  
282 regimes (Vaux & Balk 2012). The characteristics of local weather systems, particularly  
283 precipitation distribution, are also important and probably governed by precession-scale  
284 insolation cycles (Thompson *et al.* 2006). Consequently, moraine ages indicate spatial  
285 variability in the amount and timing of glacier mass loss due to variations in the timing and  
286 intensity of monsoonal and Westerly snowfall across the region (Rowan 2017, Owen 2009).

287

#### 288 **3.2 20<sup>th</sup> and 21<sup>st</sup> Centuries (1900 to present)**

289 Changes in glacier length and area during the early part of the 20<sup>th</sup> Century are described by  
290 historical accounts from early climbing expeditions. These records are based on visual  
291 comparison of the state of these glaciers to those in the European Alps, which has led to  
292 misinterpretation of ongoing glacier volume change (Grove 2004). Measurements of changes  
293 in length and area are of limited use for estimating the mass change of debris-covered  
294 glaciers that lose mass by surface lowering rather than terminus recession (e.g. Bolch *et al.*  
295 2011). Geochronological techniques such as <sup>14</sup>C and terrestrial cosmogenic nuclide dating do  
296 not currently operate at sufficient temporal resolution to describe the ages of moraines  
297 formed in the last 100 years. However, changes in glacier volume over small areas can be  
298 accurately detected by comparing multi-temporal aerial and satellite topographic data,

299 including historical imagery from the Corona and HEXAGON satellites that date back to the  
300 1950s (e.g. Bolch *et al.* 2011; Berthier *et al.* 2014). Measurements of the gravitational field of  
301 the Earth's surface (the Gravity Recovery and Climate Experiment; GRACE; Tapley *et al.*  
302 2004) combined with topographic data can be used to estimate changes in glacier mass across  
303 a broad spatial area (e.g. Ageta *et al.* 1980, Fujita & Nuimura 2011, Sarikaya *et al.* 2012) but  
304 with large uncertainties (Gardner *et al.* 2013).

305

306 Analyses of ice cores demonstrate a sharp decrease in accumulation on low-latitude (25–  
307 35°N) Himalayan glaciers, and an increase in ice volume at higher latitudes (35–70°N) on the  
308 Tibetan Plateau driven by variability in monsoon intensity and timing since 1950 (Hou *et al.*  
309 2002). However, glacier mass loss at high elevations has exceeded that which could be  
310 attributed to change in monsoon intensity alone (Mölg *et al.* 2012). Mass balance is most  
311 negative in the Eastern Himalaya, and becomes less negative to the north and in the northern  
312 and eastern parts of the Karakoram where some glaciers showed slightly positive mass  
313 balances between 1999 and 2008 (Gardelle *et al.* 2012). The opposing trends in glacier mass  
314 balance between the Karakoram and the Eastern Himalaya over the last 50 years are  
315 attributed to spatial variations in the rates of change in temperature and precipitation  
316 (Gardelle *et al.* 2012, Nakawo *et al.* 1999), as rising Northern Hemisphere air temperatures  
317 deliver winter snowfall from the Arabian Gulf further into the range (Kapnick *et al.* 2014).  
318 Climate warming appears to have accelerated the mass loss from glaciers in the Himalaya  
319 after 1995, reflecting the high sensitivity of the regional energy balance to small changes in  
320 climate (Cogley 2011). Over the same period, a slight gain in mass has been observed for  
321 glaciers in the Karakoram, attributed to a greater influence of Westerly winter snowfall  
322 (Gardelle *et al.* 2012, Yao *et al.* 2012), although not all glaciers in the Karakoram gained  
323 mass in the last 40 years (Sarikaya *et al.* 2012).

324

#### 325 **4. Predictions of future glacier change**

326 Predictions of how glaciers will continue to change from the present day requires  
327 quantifying, specifically: current glacier mass balances, the response time over which glaciers  
328 will reach equilibrium with climate, and how the climate will change over the period of  
329 interest. For the 21<sup>st</sup> and 22<sup>nd</sup> Centuries, climate model ensembles such as those produced by  
330 the IPCC (Collins *et al.* 2013) give a range of possible warming values for future emissions  
331 scenarios, which are useful for forcing meteorological and glacier modelling. Glacier models  
332 are often somewhat less sophisticated than these climate models and operate at different  
333 spatial scales, particularly in representing the dynamics of mountain glaciers, in which the  
334 processes controlling the flow of ice through steep rugged terrain and where feedbacks with  
335 often extreme topography and complex orographic meteorology are poorly documented. The  
336 rate of regional glacier change in High Mountain Asia may also be enhanced when compared

337 to lower-altitude glacierised regions, as Northern Hemisphere warming is enhanced at  
338 altitudes above 5000 m (Xu *et al.* 2016), the elevation range where the ELAs of the many  
339 large glaciers are located (Benn & Lehmkühl, 2000). To better understand better the impact  
340 of climate change on glacier mass balance, meteorological variables on a smaller spatial scale  
341 than the entire region are needed.

342

343 Predictions of glacier response to future climate change can be made either by extrapolating  
344 from observations of recent glacier change and present-day glacier characteristics, or by  
345 applying numerical glacier–climate models. These glacier models vary widely in their level  
346 of sophistication and complexity depending on the required application, but generally can  
347 either extrapolate from observed trends in the relationship between glacier mass balance and  
348 climate, or replicate the physical processes by which glacier change occurs and be forced by  
349 changing climate conditions. Numerical modelling of glacier mass balance forced by detailed  
350 simulations of mesoscale meteorology has been undertaken to better understand the  
351 atmospheric controls on Zhadang Glacier in central Tibet (Mölg *et al.* 2012), but is still in  
352 development for regional applications. Glacier-climate models can be used to make  
353 catchment-scale and regional-scale predictions of the contribution of glacial meltwater to  
354 hydrological budgets, and their contribution in the context of water supplied by precipitation  
355 or groundwater flow (e.g. Lutz *et al.* 2014). However, predictions based on numerical  
356 modelling must also consider the range of uncertainties associated with the data used to drive  
357 models. Many of these climatic and glaciological variables, such as the relationship between  
358 air temperature and ablation beneath supraglacial debris, or the subglacial conditions  
359 controlling ice flow, are poorly constrained both at present and in terms of future change in  
360 the Himalaya (Rowan *et al.* 2015).

361

362 Precipitation in the monsoon-influenced Eastern and Central Himalaya is predicted to  
363 increase by up to 10% during the 21<sup>st</sup> Century (IPCC scenario A2; Collins *et al.* 2013).  
364 Although this increase in precipitation would mean that widespread droughts are unlikely,  
365 with warmer air temperatures a greater proportion of precipitation will fall as rain rather than  
366 snow and will rapidly melt glacier ice (Meehl *et al.* 2007). The mass balances of the majority  
367 of glaciers in High Mountain Asia are out of equilibrium with present-day climate, as is the  
368 case for glaciers worldwide. A degree day model of the Eastern Himalaya based on a 20-year  
369 climate record demonstrated that loss of 25% of the glacierised area could occur with only  
370 1°C warming from present (Rupper *et al.* 2012). Under an extreme scenario of 2.5°C  
371 warming by the end of the 21<sup>st</sup> Century, the glacierised area of Bhutan would be reduced by  
372 50%, and the contribution of meltwater flux to annual hydrological budgets would become  
373 negligible (Rupper *et al.* 2012). Catchment-scale hydrological modelling of the Langtang  
374 catchment in Nepal, a typical high-altitude valley in the Central Himalaya, suggest a loss of

375 35–55% of the total glacierised area by 2100, with the contribution of areal loss from debris-  
376 covered glaciers only 25–33% over the same period (Ragettli *et al.* 2016).

377

### 378 **5. Impacts on water resources with future glacier change**

379 Until 2050, if only the contribution of glaciers to the hydrological budget is considered, river  
380 flows are likely to rise during the monsoon an effect called the ‘deglaciation discharge  
381 dividend’ (Kaser *et al.* 2010). River flows will reach ‘peak water’ then decline as glacier  
382 mass is dramatically reduced and rivers have a greater dependence on precipitation and snow  
383 melt (Soncini *et al.* 2016). ‘Peak water’ in monsoon-influenced regions is predicted to occur  
384 by the mid-21<sup>st</sup> Century, as identified by hydrological modelling of glacier meltwater  
385 production (Lutz *et al.* 2014). In Nepal, glacier mass loss is predicted to increase downstream  
386 water supplies during the first half of the 21<sup>st</sup> Century compared to 2001–2010, as the  
387 additional meltwater released each year will boost river flows. Water supplies are then either  
388 predicted to decline or remain stable depending on how the monsoon changes during this  
389 period, as the predicted 10% decrease in meltwater runoff could be compensated by a similar  
390 increase in precipitation (Ragettli *et al.* 2016). The contribution of glacier meltwater to future  
391 river flows may increase slightly during the monsoon due to enhanced ice melt, but decrease  
392 overall by 4% by 2050 as glacier mass rapidly declines (Soncini *et al.* 2016). Hydrological  
393 modelling predicts a decline in the glacial meltwater contribution to catchment hydrological  
394 budgets over the next century; by 2065 the change in mean catchment water supply is likely  
395 to be –8% in the Indus, –18% in the Ganges, and –20% in the Brahmaputra (Immerzeel *et al.*  
396 2010). These decreases in meltwater supply are likely to be compensated, at least partially, by  
397 increasing rainfall of +25% in the Indus and Brahmaputra and +8% in the Ganges. However,  
398 these projections should be treated with caution, since changes in the monsoon are currently  
399 difficult to represent in predictive climate models (Immerzeel *et al.* 2010).

400

401 Short-term increases in river flow as rainfall becomes a more important constituent of the  
402 hydrological budget are likely to increase the risk of regional flooding (Ragettli *et al.* 2016).  
403 However, the magnitude and timing of peak flows relative to the present day are generally  
404 unknown (Fig. 4). The expansion and coalescence of supraglacial melt ponds to form larger  
405 supraglacial or proglacial lakes bounded by terminal and lateral moraines presents an  
406 additional risk in the form of the hazard posed by potentially catastrophic glacial lake  
407 outburst floods (Benn *et al.* 2012). These sudden-onset floods generally arise from the  
408 breaching of an impounding moraine, and are capable of generating peak flood discharges  
409 that can exceed seasonal high flow floods by over an order of magnitude (Cenderelli and  
410 Wohl, 2001). Large glacial lakes may also be considered as an intermediate storage  
411 component in the hydrological cascade of glacierised (and generally deglaciating) catchments  
412 and effectively regulate the downstream transmission of glacial meltwater (Carrivick and

413 Tweed, 2013). An anticipated increase in the number and extent of glacial lakes as a result of  
414 climate change is of concern, especially when considered in the context of the rapidly  
415 expanding Asian hydropower sector, which is likely to become increasingly exposed to  
416 climatically controlled glacial flood hazards (Schwanghart *et al.* 2016), and modified  
417 hydrological regimes.

418

419 Beyond 2050, sustained glacier mass loss will result in declining water supplies and possible  
420 shifting of seasonal river flows, as spring meltwater will no longer sufficiently compensate  
421 for the pre-monsoon dry season (Immerzeel *et al.* 2010). River flow will decline most  
422 dramatically in the Indus basin where a significant proportion of the annual hydrograph is  
423 derived from glacier melt (Fig. 5) (Lutz *et al.* 2014). Future river flows will depend on  
424 changes in the amount and timing of precipitation, and highly seasonal river flows are likely  
425 to change their timing compared to the present day, possibly resulting in enhanced spring  
426 flows (Immerzeel *et al.* 2010). Total hydrological budgets are likely to decline dramatically  
427 by 2100, with extreme scenarios predicting a 26% decrease in flow predicted by 2100 for the  
428 Everest region, due to glaciers losing over 50% of their volume compared to 2012–2014  
429 (Soncini *et al.* 2016).

430

## 431 **6. Improving understanding of glacier response to climate change**

432 Whilst there are large uncertainties about the current state of and the ongoing changes  
433 experienced by glaciers, conclusions can nevertheless be drawn about important controls on  
434 their response to climate change to make predictions of their future state (Bolch *et al.* 2012).  
435 These predictions often contain large uncertainties, due to a lack of available data with which  
436 to validate models, and the suitability of existing glacier models which have often been  
437 developed for application to Polar ice sheets rather than glaciers flowing through steep,  
438 mountainous terrain. Many factors control the response of mountain glaciers to climate  
439 change (Fig. 6), and spatial and temporal variability in these controls can be challenging to  
440 represent in numerical models. Some key areas for possible future research to reduce these  
441 uncertainties are described here.

442

### 443 **6.1 Modification of glacier response to climate change by catchment geomorphology**

444 The dynamics and hydrology of individual glaciers are governed by characteristics such as  
445 glacier aspect, size, altitude, and hypsometry, collectively known as morphometry. These  
446 morphometric factors exert a significant control on the dynamics and mass balance of  
447 mountain glaciers (Quincey *et al.* 2009). Moreover, the pronounced interaction between high  
448 topography and atmospheric circulation systems such as the Indian summer monsoon results  
449 in distinctive local mesoscale meteorological patterns (Bookhagen & Burbank 2010). This  
450 interaction between the atmosphere, landscape and cryosphere can produce catchment-scale

451 variations in energy and mass balance that cause adjacent glaciers to exhibit different  
452 responses to the same change in climate (e.g. Glasser *et al.* 2009). For this reason, a coupled  
453 mesoscale–energy balance modeling approach represents an important advance in the  
454 understanding of glacier–climate interactions in High Mountain Asia (Mölg *et al.* 2012).  
455 Robust dynamic or statistical methods are needed to downscale climate model outputs to a  
456 scale relevant to glacier mass balance that accounts for mountainous topography (e.g.  
457 Reichert *et al.* 2002). Furthermore, these relationships may need to be reconsidered for  
458 application using future climatologies (Meehl *et al.* 2007). The degree-day model  
459 applications, such as that used by Rupper *et al.* (2012) are useful to predict regional  
460 glaciological and hydrological changes with climate variations. However, the modification of  
461 glacier–climate relationships by factors such as surface debris cover and glacier morphometry  
462 requires further exploration and the acquisition of field data for model validation.

463

#### 464 **6.2 Sensitivity of debris-covered glaciers to climate change**

465 As glaciers lose mass, debris accumulates on their surfaces, and as a result, the debris-  
466 covered glacierised area worldwide is increasing. Ablation under a supraglacial debris layer  
467 is primarily controlled by its thickness, but to a lesser extent by debris properties including  
468 lithology, moisture content and porosity (Benn *et al.* 2012). These parameters are spatially  
469 and temporally variable, due to variations in input, transport and exhumation of debris to the  
470 glacier surface in space and time (Anderson & Anderson, 2016; Gibson *et al.* 2017). Much  
471 recent work has focused on determining spatial variability in debris thickness, either remotely  
472 using thermal satellite imagery (e.g. Mihalcea *et al.* 2008; Foster *et al.* 2012; Rounce &  
473 McKinney, 2014) or directly using ground-penetrating radar (e.g. McCarthy *et al.* 2017) and,  
474 in some cases, the impact of this spatial variation on glacial hydrology (Minora *et al.* 2015;  
475 Soncini *et al.* 2016). Many of these inputs are validated with minimal field measurements of  
476 debris thickness, but such validation would greatly extent the scope of predictions that could  
477 be made considering debris-covered glacier change. Few studies currently consider the  
478 influence of spatial variation in moisture content (e.g. Collier *et al.* 2014), porosity (e.g. Evatt  
479 *et al.* 2015), albedo (e.g. Nicholson & Benn, 2013) or aerodynamic roughness length (e.g.  
480 Rounce *et al.* 2015; Miles *et al.* 2017; Quincey *et al.* 2017).

481

482 Predictions of glacier mass balance modified by debris cover requires distributed surface  
483 energy balance models that consider variations in debris cover across the glacier surface and  
484 through time, and ideally simulate the interaction of free air and moisture with the porosity of  
485 debris layers (e.g. Evatt *et al.* 2015, Collier *et al.* 2014). However, suitable models are few,  
486 often only consider one important variable (e.g. debris thickness or porosity), and are mainly  
487 driven by empirical relationships derived from limited field data (e.g. Mihalcea *et al.* 2008).  
488 Therefore, to comprehensively understand the influence of a debris layer on ablation, further

489 field data is needed to quantify the extent of variations in debris parameters and to develop  
490 understanding of heat flux through supraglacial debris. It is potentially possible to use field  
491 measurements of debris thickness to calibrate and validate a method of classification of this  
492 variable from remote observations, which would greatly extent the scope of predictions that  
493 could be made considering debris-covered glacier change. Furthermore, as field-based  
494 research tends to focus on relatively accessible, often smaller lower altitude glaciers, a remote  
495 calibration method for debris thickness has the potential to dramatically advance knowledge  
496 of how these glaciers behave in response to climatic forcing.

497

### 498 **6.3 Modification of glacier response to climate by glacier dynamics**

499 Feedbacks between glacier mass balance and dynamics control the magnitude and timing of  
500 the response of individual glaciers to climate change. Ice flow also drives processes that  
501 affect glacier mass. For example, the transport of debris to the glacier surface from englacial  
502 or subglacial storage can cause the supraglacial debris layer to thicken and thereby reduce  
503 ablation. In contrast, ice flow stagnation may promote the development and expansion of  
504 supraglacial or moraine-dammed lakes, which promotes widespread calving of the lake-  
505 terminating glacier tongue and thereby accelerating mass loss (Gardelle *et al.* 2011). Debris  
506 cover frequently causes glacier tongues to lose mass by surface lowering rather than terminus  
507 recession, which in turn affects dynamics, since ice flow tends to stagnate with the loss of  
508 driving stress (Quincey *et al.* 2009). Commonly, supraglacial or proglacial lake formation  
509 coincides with this process, and once a lake crosses a threshold of depth of about 80 m deep  
510 the lake-marginal glacier ice starts to calve (Quincey *et al.* 2007, Robertson *et al.* 2012) with  
511 potentially dramatic consequences for glacier dynamics and mass balance. Under IPCC 21<sup>st</sup>  
512 Century climate change scenarios, proglacial lakes are increasingly likely to pose a potential  
513 hazard to human life through an increased risk of sudden-onset glacial lake outburst floods.  
514 However, the timing and magnitude of lake formation and growth are difficult to predict.

515

516 Rates of recent proglacial lake expansion have been quantified from satellite imagery and are  
517 particularly well-studied in the Khumbu Himal in the Central Himalaya (Watson *et al.* 2016)  
518 and the Lunana region of Bhutan in the Eastern Himalaya (Fujita *et al.* 2008). The primary  
519 process by which lakes expand appears to be via the subaerial loss of mass at the active  
520 calving front rather than the ablation of subaqueous ice, including ice at the lake bottom  
521 (Fujita *et al.* 2009). Controls on calving rates for glaciers terminating in freshwater rather  
522 than marine settings differ, and are dominated by wave fetch and lake temperature (Sakai *et al.*  
523 2009). However, prediction of future change in these variables and the impact on lake  
524 development has not been investigated. Proglacial lake growth has typically accelerated with  
525 rapid glacier recession since the 1960s (Bajracharya & Mool 2010), but the impact of lake  
526 formation on future glacier change is poorly understood. Although it seems intuitive that

527 lake-terminating glaciers should recede more rapidly than equivalent land-terminating  
528 glaciers, few conclusive data are yet available to confirm this (e.g. King *et al.* 2017).

529

## 530 **7. Conclusions**

531 Glaciers in High Mountain Asia are changing rapidly in response to recent climate change,  
532 with many glaciers losing mass at accelerated rates since the 1990s. Sustained glacier mass  
533 loss is predicted to continue through the 21<sup>st</sup> and 22<sup>nd</sup> Centuries even in the absence of further  
534 climate warming from the present day. In the monsoon-influenced Eastern and Central  
535 Himalaya, the majority of glacier mass loss occurs during the warm summer months (June to  
536 September), which coincides with the timing of maximum precipitation at high altitudes. As a  
537 result, the glacier meltwater contribution to the Ganges and Brahmaputra catchments is  
538 relatively less important than for the Indus catchment, which drains the Westerly influenced  
539 Western Himalaya, Karakoram and Hindu Kush and has a smaller rainfall component of the  
540 annual hydrological budget. Seasonal river flows in the Eastern and Central Himalaya are  
541 therefore unlikely to decrease during the next 50 years and may even increase slightly, the  
542 ‘deglaciation discharge dividend’, as climate models predict increased monsoon precipitation  
543 with warming during this century, and because accelerated glacier melt will provide  
544 additional water over the same period. In the Western Himalaya, Karakoram and Hindu  
545 Kush, river flows are much more dependent on glacier mass change than in the catchments  
546 further to the east. The future of these glaciers is a less clear, as those in the Karakoram  
547 appear to have recently experienced a slightly increase in ice mass which may be due to the  
548 increased extent of winter snowfall resulting from warming air temperatures. However, in the  
549 Western Himalaya, glacier mass loss appears similar to that elsewhere in the mountain range,  
550 suggesting that the hydrological budget of the Indus is likely to be severely affected by  
551 climate change.

552

553 Beyond the mid-21<sup>st</sup> Century, when large volumes of glacier ice have been lost, the  
554 hydrological budget of catchments in the monsoon-influenced regions will be more  
555 dependent on the timing and availability of monsoon precipitation than glacier melt.  
556 Therefore, water availability is unlikely to decrease in the short term, as a warming climate  
557 will result in decreasing glacier runoff compensated by increased monsoon precipitation, with  
558 little change in the seasonality of river flows. In the Westerly influenced regions, glacier  
559 mass loss will likely lead to decreased river flows, and although glaciers in the Karakoram  
560 are at present showing slightly positive mass balances, this trend is unlikely to continue with  
561 sustained climate warming in the longer-term. As rainfall becomes a more important  
562 component of the total hydrological budget, the risk of flooding is predicted to increase as  
563 rainfall is transferred much more rapidly into rivers than snow that accumulates as glacier ice.  
564 In the longer term, by the start of the 22<sup>nd</sup> Century, the predicted loss of over 50% of glacier



565 volume and the complete removal of smaller, lower-altitude glaciers across High Mountain  
566 Asia is likely to result in a widespread decline in water supplies, which will have a dramatic  
567 impact on the large populations relying on these glacier-fed rivers.

568

### 569 **Acknowledgements**

570 We thank Tobias Bolch and Arthur Lutz for sharing regional and catchment boundaries and  
571 model outputs used to draw the figures.

572

573

### 574 **Figure captions**

575 Figure 1. High Mountain Asia, showing the Himalaya, Karakoram and Hindu Kush regions  
576 defined by Bolch *et al.* (2012) and major rivers. The location of glaciers for which  
577 glaciological mass balance records exist and the length of the record in years (Bolch *et al.*  
578 2012; Pratap *et al.* 2016; Soncini *et al.* 2016; Tshering & Fujita 2016), and Vincent *et al.*  
579 (2016), and the location of automatic weather stations (AWS) collecting data at high altitudes  
580 are also indicated. The names of individual glaciers or glacierised regions referred to in the  
581 text are highlighted in bold.

582

583 Figure 2. The surface features of Khumbu Glacier in the Everest region of Nepal, showing (a)  
584 the debris-covered ablation area, looking to the south, (b) the Khumbu Icefall marking the  
585 transition between the clean-ice accumulation area and the debris-covered ablation area, (c  
586 and d) typical ice cliffs and supraglacial ponds in the ablation area showing englacial debris  
587 layers within the ice, likely resulting from ice-rock avalanching (note figures circled in red  
588 for scale).

589

590 Figure 3. Direct measurements of mass balance for six glaciers in the Himalaya between  
591 1992 and 2012, showing; (a) annual mass balance, (b) cumulative annual mass balance, and  
592 (c) cumulative mass balance normalised by glacier terminus altitude. Note that the data for  
593 Kangwure Glacier between 1994 and 2008 (dashed line) are reconstructed mass balance  
594 values derived from meteorological data using the relationship between *in situ* mass balance  
595 measurements made in other years. Redrawn from Yao *et al.* (2012). See Figure 1 for the  
596 locations of these glaciers.

597

598 Figure 4. Schematic diagram showing hypothetical changes in glacier volume and meltwater  
599 release from mountain glaciers in response to regional climate warming over a period  
600 equivalent to their last advance and recession, from the Little Ice Age maximum through the  
601 present day and 21<sup>st</sup> Century.

602

603 Figure 5. Annual hydrographs for headwaters of the major Himalayan catchments produced  
604 using a hydrological model for a present-day reference period of 1998–2007, showing the  
605 contribution to the total hydrological budget from glacier melt, snow melt, rainfall–runoff  
606 and base (groundwater) flow [redrawn from Lutz *et al.* (2014)], for (a) the Indus in the  
607 westerly influenced Western Himalaya, (b) the Ganges in the transition between the westerly  
608 and monsoon-influenced Central Himalaya, and (c) the Brahmaputra in the monsoon-  
609 influenced Central and Eastern Himalaya. (d) shows the catchment boundaries used to make  
610 these calculations, with topographic imagery from Google OpenLayers.

611

612 Figure 6. Climatic, glaciological and landscape space–time controls on glacier and climate  
613 change in the Himalaya and Karakoram.

614

615

616

617

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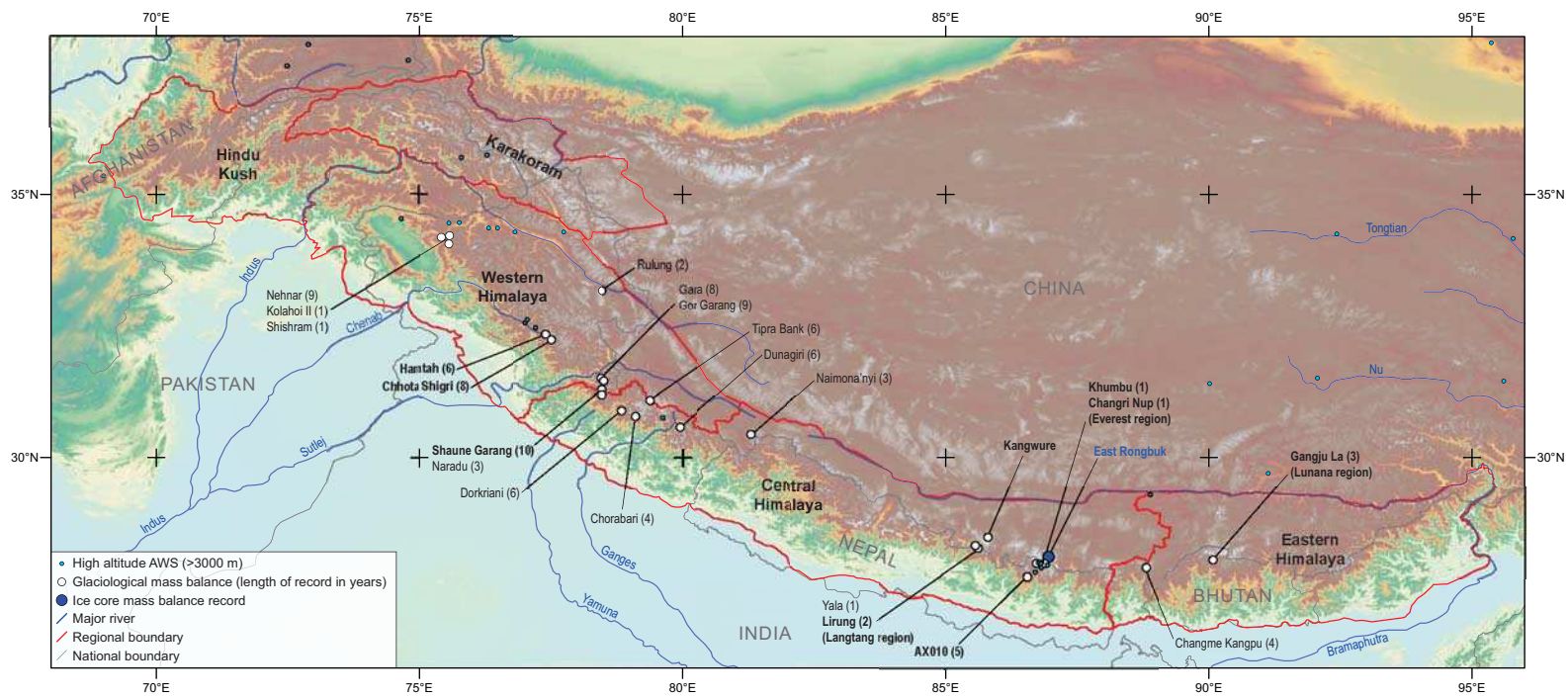
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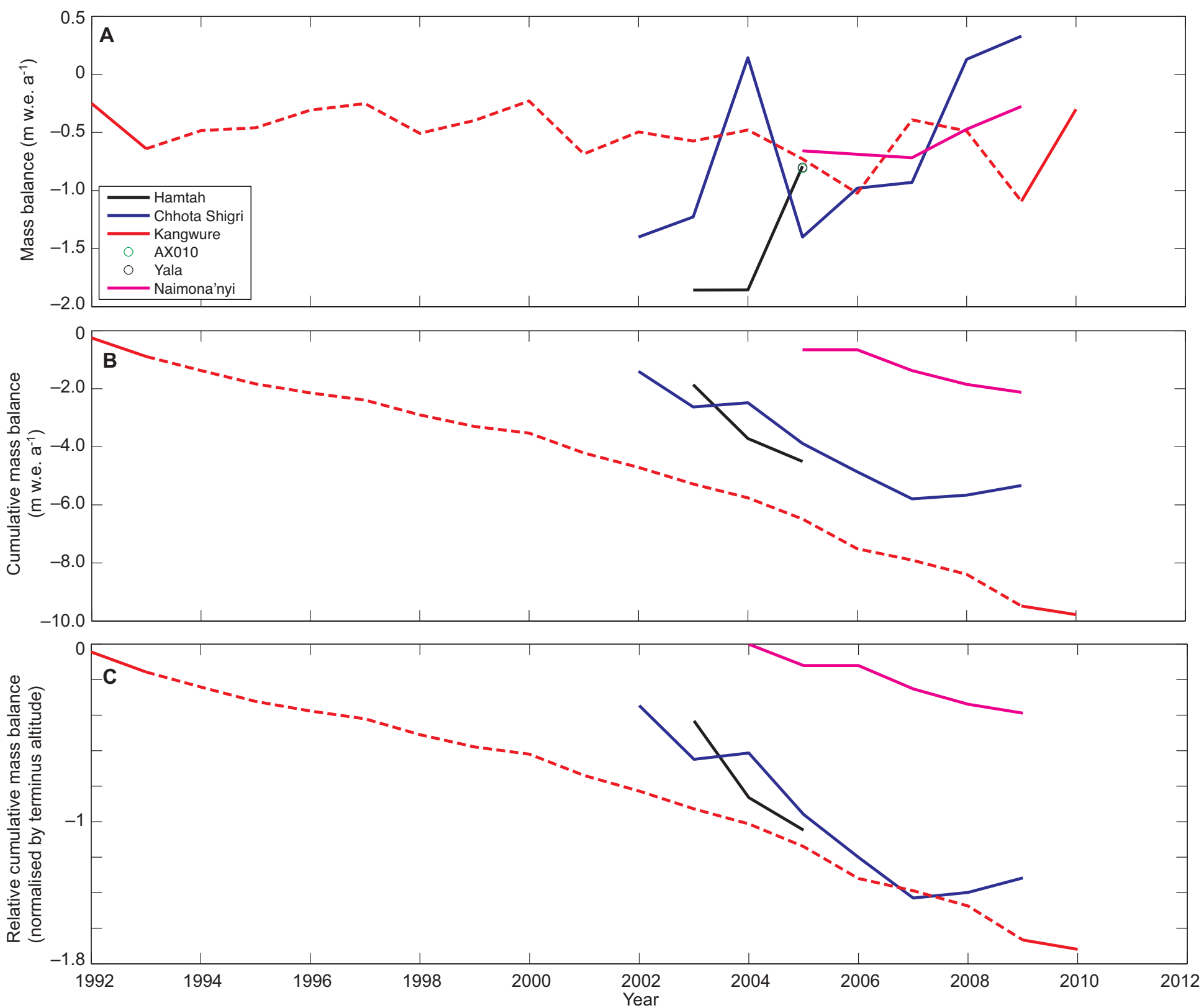
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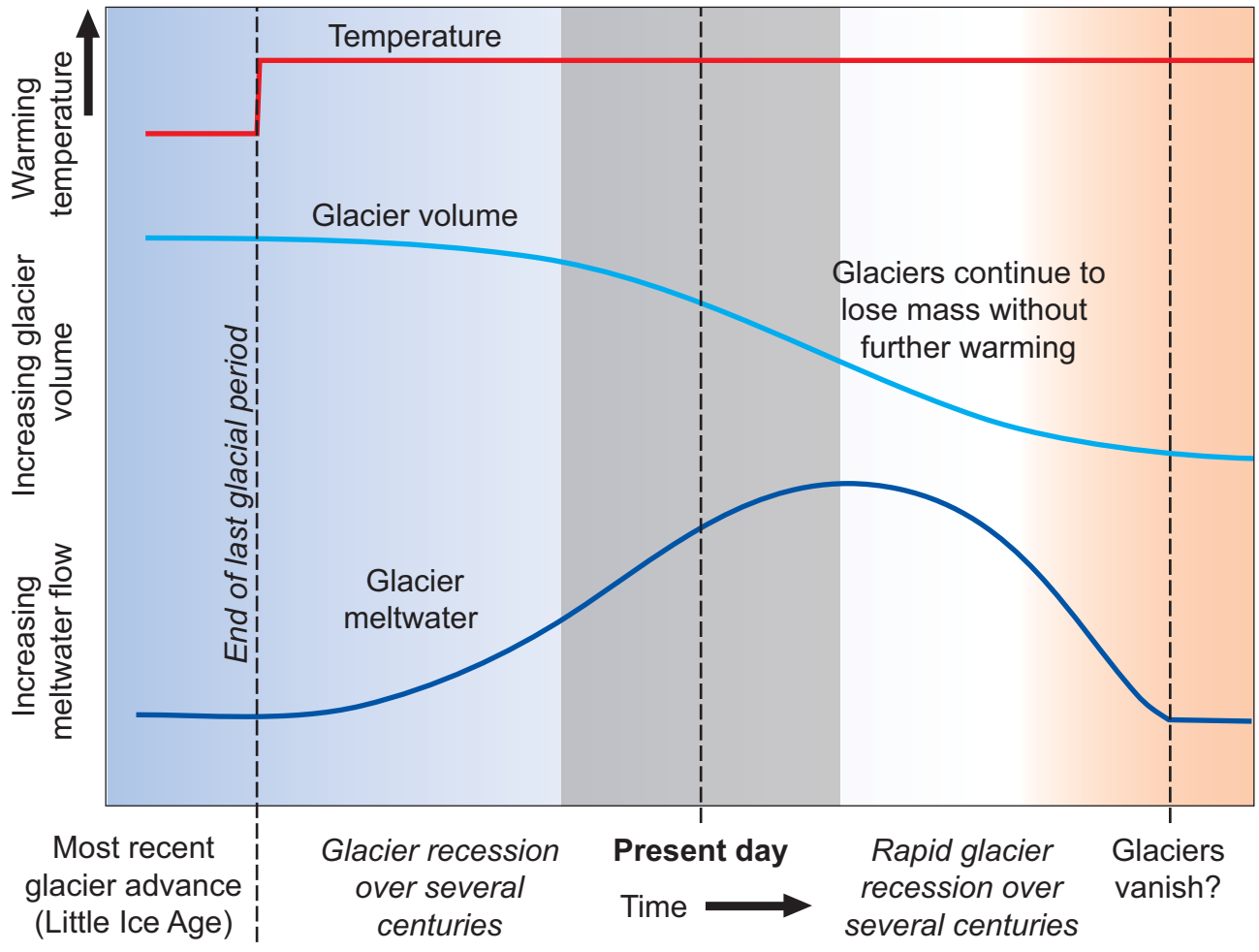
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Warming temperature ↑

Increasing glacier volume

Increasing meltwater flow

Most recent glacier advance (Little Ice Age)

Glacier recession over several centuries

Present day

Time →

Rapid glacier recession over several centuries

Glaciers vanish?

Temperature

Glacier volume

Glacier meltwater

End of last glacial period

Glaciers continue to lose mass without further warming

