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1 **The sustainability of water resources in High Mountain Asia in the context** 2 **of recent and future glacier change**

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16 17 **Abstract**

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

32 33 **1. Introduction**

34 Millions of people rely on glaciers in the Himalaya, Karakoram and Hindu Kush mountains,
35 collectively referred to as High Mountain Asia, as a water resource. These glaciers form the
36 headwaters of the largest rivers in Asia, including the Indus, the Ganges and the Brahmaputra
37 Rivers (Fig. 1) and as such the mountains are often referred to as the ‘water towers of Asia’
38 (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 thinning of 9–28 m water equivalent (w.e.) 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 km² 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 21st 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 monsoons, 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 km² 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 6th 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 near the terminus to 440 m near the icefall for
141 the largest, Khumbu Glacier (glacier area is 39.5 km²; Gades *et al.* 2000), 20–157 m for
142 Lirung Glacier (13.5 km²; Kadota *et al.* 1997) to 51–86 m for the smallest, Glacier AX010
143 (0.6 km²; Kadota *et al.* 1997). Ice thickness was 124–270 m for Chhota Shigri Glacier (15.7
144 km²) 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*
150 *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 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 glaciers since the Little
232 Ice Age (LIA) (Rowan, 2017) have left ice-cored moraines up to a hundred metres above the
233 glacier surface, which result in an unstable inner moraine face that is unstable and prone to
234 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 20th Century (Rowan 2017). These observations are based on geochronological data for
274 moraines compiled from studies using radiocarbon (¹⁴C) 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 (National Research Council 2012). The characteristics of local weather systems,
283 particularly precipitation distribution, are also important and probably governed by
284 precession-scale insolation cycles (Thompson *et al.* 2006). Consequently, moraine ages
285 indicate spatial variability in the amount and timing of glacier mass loss due to variations in
286 the timing and intensity of monsoonal and Westerly snowfall across the region (Rowan 2017,
287 Owen 2009).

288

289 **3.2 20th and 21st Centuries (1900 to present)**

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

299 accurately detected by comparing multi-temporal aerial and satellite topographic data,
300 including historical imagery from the Corona and HEXAGON satellites that date back to the
301 1950s (e.g. Bolch *et al.* 2011; Berthier *et al.* 2014). Measurements of the gravitational field of
302 the Earth's surface (the Gravity Recovery and Climate Experiment; GRACE; Tapley *et al.*
303 2004) combined with topographic data can be used to estimate changes in glacier mass across
304 a broad spatial area (e.g. Moiwu *et al.* 2011) but with large uncertainties (Gardner *et al.*
305 2013).

306

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

325

326 **4. Predictions of future glacier change**

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

337 regional glacier change in High Mountain Asia may also be enhanced when compared to
338 lower-altitude glacierised regions, as Northern Hemisphere warming is enhanced at altitudes
339 above 5000 m (Xu *et al.* 2016) where the ELAs of the many large glaciers are located (Benn
340 & Lehmkuhl, 2000). To better understand the impact of climate change on glacier mass
341 balance, meteorological variables on a smaller spatial scale 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 21st 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 melt glacier ice (Meehl *et al.* 2007). The mass balances of the majority of
367 glaciers in High Mountain Asia are out of equilibrium with present-day climate, as is the case
368 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 21st 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’ (Fig. 4) (Kaser *et al.* 2010). River flows will reach ‘peak water’ then decline as
382 glacier mass is dramatically reduced and rivers have a greater dependence on precipitation
383 and snow melt (Soncini *et al.* 2016). ‘Peak water’ in monsoon-influenced regions is predicted
384 to occur by the mid-21st Century, as identified by hydrological modelling of glacier
385 meltwater production (Lutz *et al.* 2014). In Nepal, glacier mass loss is predicted to increase
386 downstream water supplies during the first half of the 21st Century compared to 2001–2010,
387 as the additional meltwater released each year will boost river flows. Water supplies are then
388 either predicted to decline or remain stable depending on how the monsoon changes during
389 this period, as the predicted 10% decrease in meltwater runoff could be compensated by a
390 similar increase in precipitation (Ragettli *et al.* 2016). The contribution of glacier meltwater
391 to future river flows may increase slightly during the monsoon due to enhanced ice melt, but
392 decrease overall by 4% by 2050 as glacier mass rapidly declines (Soncini *et al.* 2016).
393 Hydrological modelling predicts a decline in the glacial meltwater contribution to catchment
394 hydrological budgets over the next century; by 2065 the change in mean catchment water
395 supply is likely to be –8% in the Indus, –18% in the Ganges, and –20% in the Brahmaputra
396 (Immerzeel *et al.* 2010). These decreases in meltwater supply are likely to be compensated, at
397 least partially, by increasing rainfall of +25% in the Indus and Brahmaputra and +8% in the
398 Ganges. However, these projections should be treated with caution, since changes in the
399 monsoon are currently difficult to represent in predictive climate models (Immerzeel *et al.*
400 2010).

401

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

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

419

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

431

432 **6. Improving understanding of glacier response to climate change**

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

443

444 ***6.1 Modification of glacier response to climate change by catchment geomorphology***

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

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

464

465 ***6.2 Sensitivity of debris-covered glaciers to climate change***

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

482

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

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

498

499 ***6.3 Modification of glacier response to climate by glacier dynamics***

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

516

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

527 formation on future glacier change is poorly understood. Although it seems intuitive that
528 lake-terminating glaciers should recede more rapidly than equivalent land-terminating
529 glaciers, few conclusive data are yet available to confirm this (e.g. King *et al.* 2017).

530

531 **7. Conclusions**

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

553

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

565 In the longer term, by the start of the 22nd Century, the predicted loss of over 50% of glacier
566 volume and the complete removal of smaller, lower-altitude glaciers across High Mountain
567 Asia is likely to result in a widespread decline in water supplies, which will have a dramatic
568 impact on the large populations relying on these glacier-fed rivers.

569

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571 We thank Tobias Bolch and Arthur Lutz for sharing regional and catchment boundaries and
572 model outputs used to draw Figures 1 and 5.

573

574

575 **Figure captions**

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

583

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

590

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

598

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

603

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

612

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

615

616

617

618

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