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1 Review article: The hydrology of debris-covered glaciers - state of the
2 science and future research directions

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12 Abstract

13 Debris-covered glaciers (DCGs) are characterised by distinct hydrological systems that differ
14 fundamentally from those observed on clean-ice valley glaciers. To date, most studies of DCG
15 hydrology have focused on supraglacial hydrology, given that surface streams are broadly
16 accessible and repeat observations can lead to conceptual models of channel evolution. Few have
17 characterised englacial conduits and their layout, and none have directly investigated potential
18 subglacial drainage networks in any setting. In this review, we summarise the current state of
19 knowledge relating to DCG hydrology with a global focus, and present our own field observations
20 to illustrate the distinct nature of DCG landforms on a receding high-elevation glacier in the
21 Himalaya. We draw on recent work that has gone some way towards providing a process-based
22 understanding of the formation and evolution of englacial and subglacial hydrological pathways
23 and consider the role that DCG hydrology plays in regulating water supplies to downstream
24 communities, contrasting this information with clean-ice examples. We conclude by identifying
25 important knowledge gaps that might be considered priorities for future research into DCG
26 hydrology.

27 1. Introduction

28 Debris-covered glaciers (DCGs), also referred to as debris-mantled glaciers, moraine-covered
29 glaciers or ice-cored rock glaciers, are present in nearly all of Earth's glacierised regions, with a
30 particularly large concentration in the Himalayan mountain range (Bolch et al., 2012; Scherler et
31 al., 2011). Ice and snow melt from these DCGs represents the source for some of the world's largest
32 rivers: around 25% of Earth's population is dependent on glacier melt and seasonal snowpacks for
33 drinking water, irrigation and hydroelectric power supplies (Immerzeel et al., 2010). The recent
34 mass loss of glaciers in response to the warming climate is currently increasing river discharge and
35 sea-level contributions (Lutz et al., 2014; Radić et al., 2014; Shea and Immerzeel, 2016), but studies
36 simulating future scenarios are universal in predicting long-term reductions in flow, perhaps as

37 soon as 2050 in central Asia (Barnett et al., 2005; Bolch et al., 2012; Lutz et al., 2014; Ragettli et
38 al., 2016a; Sorg et al., 2012). This may threaten water security in many regions, particularly across
39 High Mountain Asia where most rivers source from glaciers in the Himalaya (Eriksson et al., 2009;
40 Hannah et al., 2005; Immerzeel et al., 2010; Winiger et al., 2005); these glaciers currently reduce
41 vulnerability to seasonal water shortages (Pritchard, 2017). A decreased discharge of the Indus and
42 Brahmaputra rivers alone is estimated to affect 260 million people (Immerzeel et al., 2010).

43   The long-term response of DCGs to changing climatic conditions is strongly non-linear and
44 reflects both spatial variability in debris concentration and climatic controls integrated over at
45 least several decades (Benn et al., 2012; Vaughan et al., 2013). Predictions of mass loss for
46 individual glacierised regions vary hugely. For example, in the Everest region of the Himalaya,
47 Rowan et al. (2015) predicted an 8-10% mass loss of glaciers by 2100, while Soncini et al. (2016)
48 calculated up to a 50% loss, and Shea et al. (2015) up to 99% loss in extreme scenarios (warming
49 of ~3°C). At a regional scale, model predictions also vary: Zhao et al. (2014) predicted a 22% total
50 loss of all glaciers in High Mountain Asia by 2050 (contributing 5 mm to sea-level rise); Chaturvedi
51 et al. (2014) found that up to 27% of glaciers in the Himalaya-Karakoram may have ablated
52 completely by 2080 under the most rapid warming scenario; and Kraaijenbrink et al. (2017) found
53 that 36% of glaciers in High Mountain Asia will be lost by 2100 with only a conservative 1.5°C global
54 temperature rise. Clearly, predictions such as these depend sensitively on the precise climate
55 scenario used, but a number of key knowledge gaps also exist concerning the character of DCGs
56 and the processes influencing them (Bolch et al., 2012; Huss, 2011). In particular, due to the
57 remoteness and inaccessibility of such glaciers, hydrological research has been severely limited.

58   In this review we consider the current state of knowledge of DCG hydrological systems, and
59 highlight key gaps as suggested topics for further research. While the review includes hydrological
60 research relating to DCGs located anywhere on Earth, it is noted that much of this research relates
61 to high-elevation Himalayan DCGs. First (Section 2), we discuss the formation and distribution of
62 DCGs. Next, we present a summary of existing research and understanding of the hydrological
63 systems of DCGs. This is considered in terms of four hydrological domains: supraglacial (Section 3),
64 englacial (Section 4), subglacial (Section 5), and proglacial (Section 6). Finally, in light of the above,
65 we propose several potential future research directions concerning the hydrology of DCGs (Section
66 7).

2. Debris-covered glaciers

68   Kirkbride (2011) defined a DCG as 'a glacier where part of the ablation zone has a
69 continuous cover of supraglacial debris across its full width'. While this definition has been broadly
70 adopted, we do not necessarily determine that the full width of the glacier terminus must be
71 debris-covered; however, debris must cover a large enough portion of this area to distinguish it
72 from a broad medial moraine (Anderson, 2000). Therefore, we define a DCG herein as 'a glacier
73 with a largely continuous layer of supraglacial debris that covers most of the ablation area, typically
74 increasing in thickness towards the terminus'. Figure 1 shows many of the common features of a
75 typical DCG, both from above (Figure 1A) and obliquely (Figure 1B).

76   DCGs are present in nearly all glacierised regions, and occur extensively where high rates
77 of rock uplift provide large amounts of sediment through glacial erosion in young mountain ranges

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78   such as the Himalaya, Southern Alps and the Andes (Anderson and Anderson, 2016; Dunning et al.,
79   2015). Approximately 23% of all glaciers across the Himalaya-Karakoram have a debris cover
80   (Scherler et al., 2011), but due to the difficulties in mapping and accessing many glaciers, a global
81   map of DCGs has not yet been published (cf. Sasaki et al., 2016). DCGs have been mapped and
82   observed independently in the European Alps (Brock et al., 2010; Paul et al., 2004), Iceland (e.g.
83   Spedding 2000), Svalbard (e.g. Etzelmüller et al. 2001; Lukas et al. 2005), Scandinavia (e.g. Jansson
84   et al. 2000), Canada (e.g. Mattson 2000), Alaska (e.g. Kienholz et al. 2015), California (e.g. Clark et
85   al. 1994), the High Andes (Emmer et al., 2015; Janke et al., 2015; Racoviteanu et al., 2008), New
86   Zealand (e.g. Kirkbride & Warren 1999), Iran (e.g. Karimi et al. 2012), Caucasus (e.g. Stokes et al.
87   2007; Lambrecht et al. 2011) and Antarctica (e.g. Chinn & Dillon 1987; Levy et al. 2006; Mackay et
88   al. 2014). They are commonly found in areas of mountain permafrost (Schmid et al., 2015), while
89   permafrost-related patterned ground has also been observed on the debris layer of DCGs, for
90   example in Antarctica (Levy et al., 2006).

91   As well as on Earth, over 1,300 individual DCGs have been both identified (Baker, 2001;
92   Head and Marchant, 2003) and inventoried (Souness et al., 2012) in the mid-latitude regions of
93   Mars. Although currently colder and drier than Earth, Mars' so-called 'glacier-like forms' are
94   similarly lobate, debris-covered, deforming, and able to deposit debris to form bounding moraine
95   ridges. Both the detailed characterisation and broader dynamic glaciology of martian glacier-like
96   forms have been reported elsewhere (Hubbard et al., 2011, 2014), but given there are almost no
97   published data on their hydrological characteristics, planetary DCGs are not considered further
98   herein.

99   Debris is supplied to a glacier through avalanching, rockfalls and small landslides onto the
100   glacier surface (Figure 2A), thrusting from the bed, dust blown from exposed moraines or
101   solifluction from (ice-cored) moraines (Dunning et al., 2015; Evatt et al., 2015; Gibson et al., 2017a;
102   Hambrey et al., 2008; Kirkbride and Deline, 2013; Kirkbride and Warren, 1999; Rowan et al., 2015;
103   Spedding, 2000). Rockfall triggered by freeze-thaw processes (Nagai et al., 2013), landslides
104   (Hewitt et al., 2008) and permafrost degradation (Gruber and Haeberli, 2007) can also contribute
105   to the accumulation of debris on a glacier surface, and the frequency of such events appears to be
106   increasing with climate change (Gruber et al., 2004; Huggel et al., 2012). Where there is a supply
107   of debris in the accumulation zone, it is often advected into the ice and transported englacially
108   through the glacier along flowlines (Figure 2B); eventually being melted out at the surface in the
109   ablation area (Anderson and Anderson, 2016; Dunning et al., 2015; Evatt et al., 2015; Jansson et
110   al., 2000; Kirkbride and Deline, 2013). The surface of the accumulation zone is therefore commonly
111   largely free from debris, with a thin debris layer emerging at the surface near the equilibrium line
112   and increasing in thickness towards the terminus (Gibson et al., 2017b; Iwata et al., 2000). Debris
113   layers have been noted to develop more quickly, and to expand upglacier (or laterally from medial
114   moraines), during periods of glacier recession (Iwata et al., 2000). Debris can also be entrained
115   subglacially if it is frozen onto cold basal ice (Jansson et al., 2000) or where water is elevated under
116   pressure, triggering a switch from subglacial to englacial drainage and transporting debris up into
117   the glacier (Spedding, 2000).

118   A debris layer can range in thickness from a few millimetres, comprising scattered particles,
119   to several metres or more, comprising large rocks and boulders (Figure 2C & D) (Inoue and Yoshida,

120 1980; McCarthy et al., 2017). Direct measurements of thick supraglacial debris layers are difficult
121 to acquire, so published data are scarce. Gades et al. (2000) used radio-echo sounding on Khumbu
122 Glacier, Nepal Himalaya, to measure supraglacial debris up to 3 m thick; our own observations on
123 the same glacier suggest that in places the debris cover exceeds this thickness (Figure 2D). Satellite
124 imagery has also been used to approximate debris thickness in a variety of settings using debris
125 surface temperature measurements (Gibson et al., 2017b; Rounce et al., 2015): on Miage Glacier
126 in the Italian Alps, for example, the surface debris layer ranged from 0 to 0.6 m thick (Foster et al.,
127 2012; Mihalcea et al., 2008). Beneath the supraglacial debris layer, glacial ice can include entrained
128 debris (Figure 2B) or be debris-free (Figure 2C) (Schmid et al., 2015).

129   Several publications have reviewed the hydrology of 'clean-ice' glaciers (e.g. Fountain and
130 Walder, 1998; Hubbard and Nienow, 1997; Irvine-Fynn et al., 2011; Jansson et al., 2003) and ice
131 sheets (e.g. Greenwood et al., 2016). However, these reviews have omitted consideration of the
132 hydrology of DCGs, which is both under-investigated - and consequently very poorly understood -
133 and distinctive. This distinctiveness results from several characteristics, including: the presence of
134 supraglacial ponds that appear to interact intimately with near-surface englacial drainage; the
135 presence of a thick debris cover that influences patterns of surface melt and runoff; the possible
136 presence of cold ice advected from high elevation accumulation areas, influencing englacial and
137 subglacial drainage; the presence of a glacier tongue of low, or even reversed, surface slope that
138 would correspondingly influence the local hydraulic potential (Shreve, 1972); the common
139 presence of a substantial moraine-impounded proglacial lake that would also complicate near-
140 terminus englacial and subglacial water flows; and finally, the common location of DCGs in
141 monsoon-influenced areas, affecting temporal patterns of mass balance. Below, we summarise
142 the current information and understanding relating specifically to the hydrology of DCGs.

3. Supraglacial hydrology

144 Supraglacial hydrology includes meltwater generation, and meltwater transport through the
145 debris layer, supraglacial ponds, and supraglacial streams, eventually to be delivered to the
146 glacier's englacial drainage system or off the glacier margin directly to the proglacial zone. Here,
147 we discuss these flowpath components in sequence.

3.1 Meltwater generation

149 Similar to clean-ice glaciers, meltwater on DCGs is produced primarily through ablation of surface
150 ice and snow. However, the spatial pattern of the former is complicated by the presence of surface
151 debris over much of the ablation area of DCGs (Figure 2). Overall, the presence of thick debris
152 tends to suppress ablation. In the Caucasus, for example, debris layers were found to reduce melt
153 by an average of ~25% compared to clean ice (Lambrecht et al., 2011). This varies primarily with
154 the thickness and lithology of the debris layer (Figure 3). A debris layer thinner than a critical
155 thickness, typically of ~50 mm, will enhance albedo and thus increase the ablation rate compared
156 to debris-free ice. The ablation rate peaks at a debris thickness of ~2-5 mm, known as the
157 'effective' thickness (Adhikary et al., 2000; Evatt et al., 2015; Inoue and Yoshida, 1980; Juen et al.,
158 2014; Lejeune et al., 2013; Nicholson and Benn, 2006, 2013; Østrem, 1959; Singh et al., 2000;
159 Takeuchi et al., 2000). A debris thickness greater than ~50 mm instead insulates the ice from

160 incoming solar radiation, increasing the albedo but inhibiting the penetration of excess surface
161 energy to the ice and thus reducing the melt rate (Figure 3). The exact values of the critical and
162 effective thickness are strongly dependent on the thermal conductivity of the debris (Figure 3),
163 which can vary widely across a glacier surface and differs according to whether the debris is wet
164 or dry (Casey et al., 2012; Collier et al., 2014, 2015; Gibson et al., 2017a; Nicholson and Benn, 2013;
165 Pelto, 2000). For example, Kayastha et al. (2000) found that maximum ice ablation occurs beneath
166 a debris layer that is 3 mm thick on Khumbu Glacier. Variations in ablation according to these
167 factors represent an important first-order control on glacier surface morphology, and are partially
168 responsible for the characteristic hummocky topography of large mounds separated by troughs
169 superimposed upon a concave surface profile of debris-covered surfaces (Figure 1B).

170 Beneath a debris depth of 250 - 300 mm, the ice becomes almost fully insulated from short-
171 term surface energy fluxes, and the storage and conduction of heat through the debris layer plays
172 a much greater role in the ablation that occurs (Bocchiola et al., 2015; Brock et al., 2010; Conway
173 and Rasmussen, 2000; Nicholson and Benn, 2013; Østrem, 1959; Reid and Brock, 2010). For
174 example, a thick debris layer comprised of fine-grained particles with a low void space reduces the
175 rate of evapotranspiration driven by air flow at the debris-ice interface, resulting in more energy
176 available for melt (Evatt et al., 2015). Conversely, the presence of moisture in debris layers >1 m
177 thick has been found to decrease the efficiency of heat transfer by decreasing the thermal
178 diffusivity of the debris layer, thus reducing heat transmission and melt of the glacier ice below
179 (Collier et al., 2014). Although melt rates beneath thick debris layers are low, they are thus non-
180 negligible and hence need to be considered in glacier-wide surface energy-balance calculations
181 (Collier et al., 2015).

182 The ablation rate of DCGs is enhanced by the presence of supraglacial ponds (Section 3.3)
183 and ice cliffs (Figure 2B & D) that are generally absent from equivalent clean-ice valley glaciers.
184 Supraglacial ice cliffs can form through the slumping of debris from steep slopes, calving at the end
185 of supraglacial ponds, or the collapse of englacial voids (Section 4); all of which expose steep, bare
186 ice faces at the glacier surface (Benn et al., 2001, 2012; Sakai et al., 2002; Thompson et al., 2016).
187 Ice cliffs contribute a notable proportion of the ablation of DCGs (Brun et al., 2016; Buri et al.,
188 2016a; Han et al., 2010; Juen et al., 2014; Reid and Brock, 2014; Sakai et al., 2000, 2002; Thompson
189 et al., 2016), accounting for up to 69% of the total ablation of debris-covered areas whilst covering
190 as little as 2% of the total glacier area, exhibiting melt rates often 10-14 times higher than beneath
191 debris-covered ice (Immerzeel et al., 2014; Sakai et al., 1998).

192 Where ice cliffs are associated with supraglacial ponds, there is further potential for
193 increased melt rates through undercutting and calving processes (Brun et al., 2016; Buri et al.,
194 2016b; Miles et al., 2016a; Röhl, 2008; Thompson et al., 2016). Combined ice cliff and pond systems
195 have been found to contribute significantly to the surface lowering of DCGs (King et al., 2017;
196 Nuimura et al., 2012; Pellicciotti et al., 2015; Ragettli et al., 2016b; Thompson et al., 2016; Watson
197 et al., 2017) and, since it tends to be larger cliffs (and hence greater potential areas for melt) that
198 are associated with ponds (Kraaijenbrink et al., 2016a), this could lead to more rapid glacier surface
199 lowering and meltwater production (King et al., 2017). A decadal trend of surface lowering,
200 stagnation and glacier mass loss has already been observed on a large number of Himalayan DCGs
201 (Bolch et al., 2011, 2012; Kääb et al., 2012; Pellicciotti et al., 2015; Scherler et al., 2011) as a result

202 of warmer air temperatures and weaker monsoons (Pieczonka et al., 2013; Thakuri et al., 2014).
203 Surface lowering rates measured at glaciers in the Everest region were as high as 1.62 ± 0.14 m a^{-1}
204 for high-elevation land-terminating glaciers in the Pumqu catchment between 2000 and 2015
205 (King et al., 2017). Furthermore, surface lowering on DCGs is leading to an overall increase in debris
206 thickness (Gibson et al., 2017b) and an upglacier emergence of a thin supraglacial debris layer,
207 which will further increase albedo and surface meltwater production (and hence lowering,
208 potentially leading to a positive cycle until debris thickens sufficiently to insulate the surface)
209 (Kirkbride and Warren, 1999; Stokes et al., 2007), but may make observations of subsurface
210 hydrology even more difficult.

3.2 Debris layer hydrology

212 The occurrence of some ice ablation beneath even a thick debris layer implies that during much of
213 the ablation season, water must exist between the ice surface and the debris layer (McCarthy et
214 al., 2017), likely as a thin film. Subsequently, transport of this meltwater must occur, for example
215 as a saturated surface layer or – initially at least – as tiny rivulets. However, despite its importance
216 in contrasting with standard models of supraglacial hydrology based on research at clean-ice
217 glaciers, this process remains unexplored. This at least partly reflects the difficulty involved in
218 gaining non-influencing access to the ice-debris interface beneath thick surface debris. Despite the
219 absence of direct observations, meltwater transport through such a layer is likely to be slow and
220 inefficient, and water may be stored within the debris layer, introducing temporary delays in the
221 transport of meltwater through the system and thus affecting meltwater hydrochemistry (Tranter
222 et al., 1993, 2002), the development of other parts of the drainage network and the proglacial
223 discharge.

224 However, some parallels may be drawn from comparable systems, such as water flow
225 within debris above and the active layer of permafrost, moraines and talus fields, in order to
226 speculate how this transport may occur. For example, Hortonian overland (or infiltration excess)
227 flow is used to describe initial annual melt in permafrost regions, when frozen soils limit infiltration
228 producing a shallow saturated soil layer, above which overland flow is produced (Woo, 2012; Woo
229 and Xia, 1995). This has been observed within talus fields that are underlain by seasonally frozen,
230 and hence impermeable, ground (Liu et al., 2004) and within the active layer located beneath a
231 layer of debris, with meltwater infiltrating down to, then flowing downslope above the
232 impermeable permafrost table (Rist and Phillips, 2005). Where bedrock or debris is present, water
233 is transported through cracks or spaces between the rocks, but may also follow furrows between
234 linked depressions (Woo, 2012). A similar situation may hold on a smaller scale between
235 impermeable glacial ice and the overlying debris layer, with water flowing in runnels eroded either
236 down into the ice or between rocks in the debris layer. However, any such model remains to be
237 evaluated.

238 In general, water flow within and below the supraglacial debris layer and across the
239 impermeable supraglacial ice surface would be expected to be directed downglacier towards the
240 terminus and lateral margins (Winter, 2001). Clean-ice glaciers typically have a convex supraglacial
241 geometry, producing clearer watersheds and drainage routes. On DCGs, this pattern is complicated
242 by the presence of hummocky topography and a concave surface profile that commonly results
243 from the reversed mass balance gradient (Bolch et al., 2011), interrupting and complicating these

244   drainage routes (Benn et al., 2017; Miles et al., 2017). Although relatively unexplored, these factors
245   can lead to multiple scales of superimposed hydrological units, from a single supraglacial
246   depression to the full watershed hydrological unit (Winter, 2001).

247   Moraine-talus features in proglacial environments may also provide a shallow subsurface
248   flow system comparable to water within the supraglacial debris layer on DCGs. Investigation of a
249   moraine-talus feature containing buried ice at Opabin Glacier in the Canadian Rockies
250   demonstrated a system of small channels flowing over the buried ice within the moraine, through
251   the bedrock and talus field beyond the moraine (Roy and Hayashi, 2009). Langston et al. (2011)
252   reported that subsurface ice at the same glacier acted as an impermeable layer causing relatively
253   fast and shallow groundwater flow towards depressions within the proglacial moraine. The water
254   accumulated within these depressions, saturating sediments or surface water features and
255   enhancing the melt of the subsurface ice (Langston et al., 2011). Both of these situations could be
256   plausible within the debris layer of a DCG: meltwater contained within the layer could augment
257   the melt of glacier ice below, or it could initiate and contribute to supraglacial hydrological features
258   such as supraglacial ponds and streams.

3.3 Supraglacial ponds

260   Supraglacial ponds (Figure 4), a term here used to also encompass larger water bodies elsewhere
261   referred to as lakes, are extremely common and important features on DCGs, particularly those
262   with recent surface lowering. Ponds are generally absent from clean-ice valley glaciers, but are
263   prevalent on low-gradient areas of glaciers draining ice sheets; close to the margin the surface is
264   too steep for water to accumulate (Chu, 2014; Sundal et al., 2009). Similarly on a DCG, given a
265   water supply, the most important control on the location of supraglacial pond formation is the
266   slope of the glacier surface, with ponds being most prominent in areas with the lowest gradients
267   (Miles et al., 2016b; Quincey et al., 2007; Reynolds, 2000; Sakai, 2012; Sakai et al., 2000; Sakai and
268   Fujita, 2010; Salerno et al., 2012). A surface gradient of 2° or less promotes the development of
269   larger lakes; at slopes greater than this threshold, smaller isolated and transient ponds are more
270   likely (Miles et al., 2016b; Quincey et al., 2007; Reynolds, 2000). Salerno et al. (2012) additionally
271   found that the upglacier slope has an influence on pond formation, being inversely correlated to
272   the total area of lakes downglacier.

273   Glacier velocity and motion type exert less important controls over the location of
274   supraglacial ponds. An increase in lake concentration was reported towards the terminus of DCGs,
275   which is also characterised by low or very low surface velocities (Kraaijenbrink et al., 2016a; Miles
276   et al., 2016b; Quincey et al., 2007; Sakai, 2012; Salerno et al., 2012, 2015). A decrease in velocity
277   towards the glacier terminus, as well as ice inflow at flow unit confluences (Kraaijenbrink et al.
278   2016b), causes longitudinally compressive flow, which tends to close transverse crevasses and
279   englacial conduits and force water back to the surface, as well as limiting drainage from the glacier
280   surface (Kraaijenbrink et al., 2016a; Miles et al., 2016b). The thinning and stagnation of DCG
281   termini may additionally have resulted in enhanced melting beneath the debris layer, further
282   promoting the formation of ponds (Salerno et al., 2015; Thakuri et al., 2016).

283   Initial supraglacial pond growth occurs through subaqueous melting at the base of any
284   slight depression (Chikita et al., 1998; Mertes et al., 2016; Miles et al., 2016a; Stokes et al., 2007;

285 Thompson et al., 2012). Once water has accumulated and been warmed by incoming solar
286 radiation, the pond becomes warmer than the surrounding ice. For example, Chikita et al. (1998)
287 measured a maximum temperature of ~5°C at the surface of a supraglacial lake on Trakarding
288 Glacier, Nepal Himalaya. Excess energy is thus available for further ablation both vertically and
289 laterally where the pond water is in contact with ice, increasing the pond size, steepening marginal
290 slopes and mobilising debris to expose bare ice (Stokes et al., 2007). Xin et al. (2012) observed on
291 Koxkar Glacier, Tien Shan mountains, that meltwater at 0°C flowing into a pond initially cooled the
292 surface layer, but gradually mixed with warmer, deeper layers and warmed to ~4°C. This increased
293 the layer's density, causing it to sink and therefore move the warmer water towards the base of
294 the pond, providing greater potential for additional subaqueous melting. In addition, wind-driven
295 currents promote water circulation and vertical transfer of heat downwards, further enhancing
296 basal melt of the pond (Chikita et al., 1998).

297   Many supraglacial ponds are surrounded by ice cliffs (Figure 4) where ponds can expand by
298 subaerial melting and backwasting of the bare ice face (Röhl, 2008). Pond stratification and wind-
299 driven currents may further enhance the subaqueous melt expansion of supraglacial ponds by
300 triggering calving of the ice cliffs. The warm surface layer of the pond is disrupted by wind-driven
301 currents, and where it come into contact with glacier ice, can undercut the cliff beneath the
302 waterline. Progressive undercutting and thermo-erosional notch development may then lead to
303 calving of the ice cliff face (Chikita et al., 1998; Kirkbride and Warren, 1997; Mihalcea et al., 2006;
304 Miles et al., 2016a; Röhl, 2006, 2008; Sakai et al., 2009). Ice cliff calving occurs when the
305 subaqueous melt rate exceeds the ice cliff melt rate; this is noted to be effective when the fetch is
306 greater than 20 m and the water temperature is 2-4°C, though is possible at lower values (Sakai et
307 al., 2009). Calving expansion is particularly effective at larger ponds (Röhl, 2008).

308   Calving events cause further mixing of pond layers, driving warmer surface water towards
309 the base and again enhancing basal melting. Thompson et al. (2012) reported that the largest
310 deepening rates of a supraglacial pond on Ngozumpa Glacier, Nepal Himalaya, occurred adjacent
311 to the highest calving ice cliffs. Furthermore, when debris that has been heated by solar radiation
312 falls into a pond, it contributes to the energy available for melt around the pond base (Thompson
313 et al., 2012). Although shallowing of ponds can occur by sedimentation from inflowing water, this
314 tends to be outstripped by growth caused by ablation (Thompson et al., 2012).

315   A pattern of supraglacial pond evolution has been observed on DCGs, primarily based on
316 observations in the Himalaya. According to this model, supraglacial ponds form as 'perched ponds'
317 that lie above the englacial drainage network (Benn et al., 2012). As these ponds increase in area
318 and depth, they evolve from perched to base-level features, where the base-level is determined
319 by the height at which water leaves the glacial system (usually the elevation of a spillway through
320 the moraine at the glacier terminus or even the bed if water is transported there) (Mertes et al.,
321 2016; Thompson et al., 2012). However, differing sub-catchments may have differing base-levels
322 defined by other hydrological features such as moulins, which can result in a stepped hydrological
323 cascade based on these local base-levels; alternatively, the presence of a groundwater system can
324 produce a regional base-level. Where glaciers are in recession, an increasing number of
325 supraglacial ponds will form and grow over time, creating a chain of terminus-base-level lakes that
326 coalesce as each individually increases in area (Figure 1) (Sakai, 2012; Salerno et al., 2012). The

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growth of base-level lakes is not limited by periodic drainage, and so such lakes can potentially increase exponentially in area, particularly through calving processes (Benn et al., 2001; Sakai, 2012; Thompson et al., 2012). If meltwater cannot escape from the system, lake expansion and coalescence will eventually lead to the formation of a single base-level moraine-dammed lake at the terminus (Mertes et al., 2016), that will then continue to expand both upglacier and downwards by ice melt. Water can escape the system by permeating through or flowing over the terminal moraine in a proglacial outlet spillway as the lake fills with sediment, or in rare instances the moraine dam may fail causing the lake to drain (Benn et al., 2001, 2012; Chikita et al., 2001; Sakai, 2012).

The progression of supraglacial pond evolution can currently be observed at various stages on many Himalayan glaciers. Several regions have experienced an increase in supraglacial pond area and proglacial lake formation in recent decades, assumed to be in response to a warmer climate and glacier surface lowering, for example glaciers in the Tian Shan mountains (Wang et al., 2013), Bhutan Himalaya (Ageta et al., 2000; Komori, 2008), Nepal Himalaya (Benn et al., 2000; Watson et al., 2016), New Zealand (Kirkbride and Warren, 1999; Röhl, 2008) and the Andes (Harrison et al., 2006; Rivera et al., 2007). Within the Hindu-Kush Himalaya, a clear divide has appeared between glacial lakes in the East, where there are a greater number of larger lakes that have grown progressively between 1990-2009 to become increasingly proglacial, compared to the western Himalaya, where smaller supraglacial lakes have generally been decreasing in area (Gardelle et al., 2011).

As isolated perched ponds widen and deepen, they can become connected to the englacial system by deepening to a point where they intersect englacial drainage channels and drain rapidly (Benn et al., 2001; Qiao et al., 2015; Röhl, 2008; Watson et al., 2016; Wessels et al., 2002), which temporarily halts the process of pond expansion (Mertes et al., 2016). Drainage then occurs periodically in a cycle of expansion and englacial connection, unlike larger, permanently hydraulically connected ponds, which tend to be more stable due to inputs of meltwater from streams and other ponds farther upglacier (Benn et al., 2001; Miles et al., 2017; Wessels et al., 2002). An abundant supply of meltwater from the ice surface or the wider drainage system is indicated by ponds with a high suspended sediment concentration (SSC); these ponds may also expand more rapidly due to the increased presence of warmer water for ablation of the pond walls (Takeuchi et al., 2012). Narama et al. (2017) observed a seasonal pattern of supraglacial pond filling and drainage, with 94% of their observed ponds over seven glaciers draining between 2013-2015. Pond seasonality has also been noted by Miles et al. (2016b), who found the maximum ponded area of five glaciers in the Langtang Valley, Nepal Himalaya, to occur during June for the study period 1999-2013. Larger ponds were also observed to partly drain and separate into multiple, smaller ponds, and later refill to form one large pond (Benn et al., 2001; Miles et al., 2016b; Wessels et al., 2002). Warmer temperatures during the spring months have been noted to correlate with a greater number of drainage events later in the same year, potentially due to the subsurface drainage system becoming increasingly connected from a greater amount of meltwater earlier in the year (Qiao et al., 2015).

Pond drainage is promoted in zones of higher local surface velocity and hence strain rates, creating a greater connectivity between the supraglacial and englacial drainage networks; more

369   frequent drainage in such regions results in smaller-sized ponds (Miles et al., 2016b). However, as
370   noted earlier in this section, ponds are more likely to form in areas with lower surface velocities.
371   Ponds may also drain by preferentially exploiting inherited structured weaknesses such as
372   sediment-filled crevasse traces, crevasses and englacial channels that have been forced closed in
373   regions of longitudinal compression, allowing drainage by hydrofracture (the penetration of a
374   water-filled crevasse through an ice mass assisted by the additional pressure of the water at the
375   crevasse tip) (Benn et al., 2009, 2012, 2017; Gulley and Benn, 2007; Miles et al., 2016b).
376   Alternatively, perched ponds may drain by overspilling, when a channel is melted into the
377   downstream end of a pond; if, during drainage, this channel incises faster than the pond level
378   decreases then unstable and potentially catastrophic drainage can result (Qiao et al., 2015;
379   Raymond and Nolan, 2000). Analyses on Lirung Glacier, Nepal Himalaya, provided strong evidence
380   of continuous inefficient drainage of supraglacial ponds, likely into debris-choked englacial
381   conduits (Miles et al., 2017).

382   Supraglacial ponds are responsible for a large proportion of the melt from DCGs. Sakai et
383   al. (2000) estimated that ponds on Lirung Glacier absorb seven times more heat than the ice
384   beneath the debris-covered area, with at least 50% of this released with the melt output from the
385   pond. Miles et al. (2016a) found that subaqueous melt rates can keep pace with the backwasting
386   of ice cliffs, enabling these systems to propagate, and enabling the ice cliff to persist and backwaste
387   stably (Brun et al., 2016; Buri et al., 2016b). Both Sakai et al. (2000) and Miles et al. (2016a) inferred
388   that ponds have a strongly positive surface energy balance, and the warm water they discharge
389   contributes to internal melting along englacial conduits. This in turn leads, in some cases, to roof
390   collapse and the formation of new ponds (Benn et al., 2012; Miles et al., 2016a; Sakai et al., 2000),
391   resulting in a net glacier-wide increase in ablation rate. Salerno et al. (2012) stated that the
392   increasing presence of ponds is the clearest indicator of the effect that climate change is having
393   on DCGs.

394   3.4 Supraglacial streams

395   Supraglacial streams are commonly difficult to discern in debris-covered or crevassed regions of
396   the glacier surface, and therefore are rarely recorded in the literature. Large streams can be traced
397   in the upper reaches of some glaciers in satellite imagery, but small surface streams and diffuse
398   flows are less easily located and thus their prevalence remains unreported. For streams to form
399   and grow, a large catchment is required (Benn et al., 2017; Gulley et al., 2009a) and the rate of
400   stream downcutting must outpace the rate of surface lowering (Marston, 1983). Such conditions
401   may be promoted beneath thick debris with the ability to suppress surface ablation (Benn et al.,
402   2017). However, the presence of supraglacial streams has been recorded, ranging from small
403   temporary incisions to large perennial channels (Figure 5A-C). Stream growth and downcutting is
404   driven by thermal erosion (Marston, 1983) and can be marked by grooves down the side of the
405   channel showing previous high water-levels (Figure 5C); in extreme cases, ice cliffs form either side
406   of the stream. While such cliffs form on clean-ice glaciers, the relief of those on DCGs appears to
407   be more pronounced (Figure 5C & D), probably due to the debris-related suppression of surface
408   lowering away from surface streams.

409   Supraglacial streams have been noted originating in the upper ablation area of Khumbu
410   Glacier, for example beneath the ice fall, with at least one perpetual feature visible in several years

411 of satellite imagery (Gulley et al., 2009a). These streams seldom intersected with supraglacial
412 ponds, instead progressively eroding into the debris layer with notable rates of downcutting: one
413 stream was 5-10 m deep when it reached the lower ablation area where the debris layer is
414 substantially thicker (Gulley et al., 2009a; Iwata et al., 1980). In this region, streams enter the
415 glacier's interior; the nature of the entry point is unknown (Iwata et al., 1980). This is supported
416 by our observations of a large supraglacial stream on Khumbu Glacier (Figure 5A & B) which begins
417 to downcut into the glacier surface in the upper-mid ablation area and eventually disappears and
418 becomes englacial in the mid-ablation area (Figure 5D), although the exact point of transition
419 cannot be seen due to the presence of supraglacial debris and layers of older relict channels.

420 Similar to regions of clean-ice, supraglacial streams may drain into DCGs through crevasses
421 or moulins (Gulley et al., 2009a; Iwata et al., 1980), or incise to a depth that they develop a closed
422 roof through snow and debris accumulation combined with ice creep (Gulley et al., 2009a; Jarosch
423 and Gudmundsson, 2012). However, supraglacial streams on DCGs differ from those on clean-ice
424 glaciers due to the former's commonly reversed surface profile (Section 3.1). Such features are
425 therefore often interrupted by crevasses or hummocky topography, and may not persist a long
426 way along the glacier (Benn et al., 2017). Low surface gradients, low strain and longitudinal
427 compression reduce the capacity for crevassing in the lower ablation area of heavily debris-
428 covered tongues, where crevasses are therefore rarely encountered. Farther upglacier, often
429 under conditions of strong longitudinal extension associated with ice falls, open crevasses are
430 common and may suppress supraglacial stream development (Benn et al., 2017).

4. Englacial hydrology

432 There are relatively few observations of englacial drainage systems within DCGs, either directly or
433 indirectly inferred. However, as most DCGs have a steep and variable surface gradient in their
434 accumulation and upper ablation zones, crevassing can be prevalent, providing a route for
435 supraglacial stream water to access the interior of the glaciers.

436 A glacier's thermal structure determines the water content of englacial ice, thereby
437 exerting a primary control on the ability of an englacial hydrological system to form. Glacial ice can
438 be defined as: cold (ice temperature below the pressure melting point); warm (temperature at the
439 pressure melting point); or polythermal (zones of warm and cold ice). Glaciers with a polythermal
440 structure can be further subdivided into several categories depending on the location of the
441 boundary between the warm and cold ice (Blatter and Hutter, 1991). Glaciers with a higher warm
442 ice content are more likely to contain a defined englacial hydrological system, as cold ice near the
443 surface can limit, but not necessarily completely preclude, the penetration of meltwater into the
444 glacial drainage system (Irvine-Fynn et al., 2011). Unfortunately, very few studies have determined
445 the thermal structure of DCGs, and therefore little is known about whether and how water may
446 route through these ice masses. Mae et al. (1975) measured an ice temperature of -5.3°C at 2.7 m
447 depth within a borehole in the upper ablation area of Khumbu Glacier. By assuming that the ice
448 temperature would increase with depth, they estimated the ice would reach pressure melting
449 point at 16 m depth, and below this be warm-based to the bed (Mae, 1976; Mae et al., 1975).
450 Similar assumptions were made for Rongbuk Glacier, Tibet, where ice temperatures were
451 measured to a depth of 10 m (Academica Sinica, 1975). At a depth of 3 m, the ice temperature was

452   -4°C and continued to increase with depth. However, since none of these studies was able to
453   measure temperature at a depth beyond that influenced by seasonal variations in air temperature
454   (~10-15 m), the influence cannot be isolated. The assumption of a continued temperature increase
455   to the pressure melting point with depth may also not be valid.

456   Techniques involving proglacial water properties have allowed some inferences to be
457   drawn relating to the existence of englacial drainage systems within DCGs. Hydrological studies of
458   surface mass balance components of Biafo Glacier, Karakoram Himalaya, allowed Hewitt et al.
459   (1989) to infer water storage within the glacier at the start of the melt season, between the time
460   of initial meltwater production and the subsequent reactivation/development of the drainage
461   system. Hydrogeochemical analyses, particularly based on meltwater electrical conductivity (EC),
462   have also been used to infer drainage pathways. Englacially-transported water has a lower
463   sediment and ionic chemical content than subglacially-transported water, which entrains particles
464   and solutes during contact with freshly-eroded basal debris, and therefore displays higher ionic
465   concentrations, and hence EC values (Kumar et al., 2009). Consequently, studies have utilised this
466   binary classification of low-EC supraglacially- and englacially-routed water as opposed to high-EC
467   subglacially-routed water to attribute, via a mixing model, the proportions of water flowing
468   through each system. For example, Hasnain & Thayyen (1994) used such a method to determine
469   and differentiate between the englacial and subglacial components of the proglacial discharge of
470   Dokriani Bamak Glacier, Garhwal Himalaya. Englacially, they found an efficient system that was
471   active through the melt season, and that the amount of meltwater transport was proportional to
472   supraglacial water production, implying a direct link between these systems. However, despite the
473   evident utility of this approach, the assumption underpinning such mixing models has been
474   questioned. Glacier drainage systems are inherently more complex than comprising only two
475   principal pathways, and the solute content and degree of subglacial weathering can vary at a
476   number of timescales (Sharp et al., 1995).

477   Englacial water storage and transport has been inferred from measurements of
478   supraglacial pond water-levels, and an assumed connection between the pond and an englacial
479   channel. Thakuri et al. (2016) measured a constant water-level in Imja Tsho lake on Imja Glacier,
480   Nepal Himalaya, after the melt season, despite reduced precipitation and air temperatures,
481   implying decreasing meltwater production. The authors attributed this to a lake recharge from
482   englacially- and subglacially-stored water that was being progressively released over time. Further,
483   the repeated filling and drainage cycle of perched ponds suggests that englacial conduits may play
484   an important role in perched lake life cycles (Benn et al., 2017; Miles et al., 2017). Narama et al.
485   (2017) found that the seasonal drainage cycle of supraglacial ponds on seven glaciers in the Tien
486   Shan was characterised by a connection to an established englacial drainage system later in the
487   summer: 94% of lakes drained and connected to an englacial system on all three years studied.
488   Benn et al. (2012) proposed that the influx of large volumes of monsoon precipitation during the
489   summer months may result in the opening of englacial (and subglacial) conduits, leading to the
490   potential for considerable englacial ablation, subsequently calculated by Miles et al. (2016a) to be
491   ~2600 m^3 for a surface pond of 500 m^2 over a single monsoon season.

492   There have been a number of direct observations of englacial channels using
493   glaciospeleological techniques to explore and map conduits, and subsequently formulate theories

494 of channel development. Glaciolospeleology has been carried out on several DCGs primarily in the
495 Nepal Himalaya, including Khumbu Glacier (Gulley et al., 2009a), Ngozumpa Glacier (Benn et al.,
496 2009, 2017; Gulley and Benn, 2007), Ama Dablam and Lhotse Glaciers (Gulley and Benn, 2007),
497 and several DCGs in the Tien Shan (Narama et al., 2017). These investigations have provided direct
498 confirmation of a linked supraglacial-englacial system, often created by the drainage of
499 supraglacial ponds through englacial conduits (Gulley and Benn, 2007; Narama et al., 2017). On
500 Southern Inylchek Glacier, Tien Shan mountains, Narama et al. (2017) discovered both short
501 englacial channels linking chains of supraglacial ponds and longer channels with steeper gradients
502 extending from surface moulins. The latter may occur at the hydrological base-level of the glacier,
503 and show multiple levels of incision from progressive supraglacial pond drainages over time as the
504 base level has been eroded downwards (Gulley and Benn, 2007). Similar observations have been
505 made on Khumbu Glacier (Gulley et al., 2009a); however, Gulley & Benn (2007) noted that conduits
506 at different elevations may have varying local base-levels (Section 3.3). In such scenarios, this may
507 suggest that a subglacial drainage system either does not exist beneath the glacier, or is not linked
508 to the englacial system, or re-emerges to base level. Such scenarios, however, remain unreported.

509 Three formation mechanisms for englacial channels within DCGs have been proposed,
510 primarily from glaciospeleological investigations (Gulley et al., 2009a):

511 (I)   'Cut-and-closure' type conduits begin as supraglacial streams that incise downwards
512   over time, followed by roof closure through ice creep and supplemented by filling with
513   snow, ice and debris (Jarosch and Gudmundsson, 2012). This process requires a high
514   meltwater discharge such that downward incision is more rapid than glacier surface
515   ablation. Under such conditions, downcutting will continue to the hydrologic base-level
516   of the glacier (Gulley et al., 2009a). Cut-and-closure type conduits have been reported
517   by Gulley et al. (2009) on Khumbu Glacier, and Thompson et al. (2012) on Ngozumpa
518   Glacier. These conduits may be subject to repeated cycles of abandonment and
519   reactivation as water supply varies through the year, with abandoned channels closing
520   by ice creep. However, such channels rarely close completely due to their shallow
521   depth, and may be filled with sediment traces which provide lines of secondary
522   permeability by which the channel can be reactivated (Benn et al., 2009; Gulley and
523   Benn, 2007; Gulley et al., 2009a).
524 (II) Meltwater may aggregate to form englacial channels by exploiting lines or planes of
525   secondary permeability; for example those left by relict cut-and-closure channels, or
526   debris-filled and/or compressed former surface crevasses (Benn et al., 2012; Gulley and
527   Benn, 2007; Gulley et al., 2009b). This may also be one mechanism by which perched
528   supraglacial ponds can drain (Miles et al., 2017). Along these low-permeability zones,
529   discharge through the icy matrix leads to the development of enlarging lines of
530   preferential flow due to viscous dissipation, eventually forming a phreatic conduit
531   (Benn et al., 2012).
532 (III) Englacial channels may also form by hydrofracturing (Benn et al., 2009, 2012; Gulley et
533   al., 2009b). However, this is considered to be uncommon on DCGs as it requires surface
534   runoff to enter an open crevasse and is therefore generally restricted to elevations
535   above the debris-covered areas of DCGs (Benn et al., 2012). Nonetheless, channel
536   formation by hydrofracturing has been reported on Khumbu Glacier, where the

13

537   channels formed within a region of strong transverse extension that resulted in the
538   formation of longitudinal crevasses (Benn et al., 2009, 2012). In such zones, repeated
539   hydrofracturing is encouraged by the combined effect of elevated water pressure in
540   the base of a supraglacial lake with transverse stresses, producing successively lower
541   niches in the walls indicating multiple stages of hydrofracturing followed by channel
542   closure by freeze-on (Benn et al., 2009).

543   Longer-distance water transport has also been observed through perennial sub-marginal
544   channels located along the edge of DCGs, likely formed by cut-and-closure of supraglacial channels
545   (Benn et al., 2017; Thompson et al., 2016). Gulley & Benn (2007) suggested that such marginal
546   features could provide longer-distance and more hydraulically efficient pathways than shallower
547   englacial conduits that occur more centrally within the glacier, due to the frequent presence of
548   infilled crevasse traces that can be exploited by water flowing at the margins. Centrally-located
549   englacial conduits are more likely to be discontinuous in nature, as a result of enhanced surface
550   lowering which can expose part of a conduit and re-route the water back to the surface (Figure 6)
551   (Miles et al., 2017).

552   Englacial conduits within DCGs may increase in efficiency through the melt season, as water
553   transported through the channels provides additional energy for melt and consequently erodes
554   the channel walls (Miles et al., 2016b, 2017; Sakai et al., 2000). For englacial channels located near
555   the surface, rapid expansion can result in conduit collapse as the roof is not sufficiently supported,
556   with the conduit walls forming ice cliffs and contributing to more rapid surface lowering of the
557   glacier surface (Benn et al., 2017; Kraaijenbrink et al., 2016a; Miles et al., 2016a; Sakai et al., 2000;
558   Thompson et al., 2016, 2012). Pond drainage events can further accelerate this process, as well as
559   adding to the total glacier mass loss as the drained water conveys large amounts of energy,
560   contributing to more rapid erosion of the conduit walls (Sakai et al. 2000; Benn et al. 2012; Miles
561   et al. 2016a; Thompson et al. 2016). Rounce *et al.* (2017) observed an outburst flood at Lhotse
562   Glacier, Nepal Himalaya, which they attributed to be at least partly triggered by the release water
563   stored in englacial conduits that became overburdened during the transitional pre-monsoon
564   season, when meltwater production is increasing and the subsurface hydrology is not fully
565   developed.

566   Englacial conduit collapse, or closure in areas of transverse compression, can provide new
567   depressions for supraglacial ponds to form, or facilitate the formation of larger lakes (Benn et al.,
568   2001, 2012; Kirkbride, 1993; Kraaijenbrink et al., 2016a; Sakai et al., 2000; Thompson et al., 2012).
569   Conduit collapse results in new bare ice faces, including ice cliffs, where melt rates will be
570   enhanced and the depression may become flooded by that increased meltwater production
571   supplementing inputs from upglacier (Benn et al., 2012). The enhanced ablation of both the
572   meltwater retained within the depression, and the surrounding newly-formed ice cliffs (the old
573   channel walls), will accelerate the melt and surface subsidence of the glacier (Thompson et al.,
574   2012).

575   5. Subglacial hydrology

576   Almost nothing is known about the subglacial drainage of DCGs due to the difficulties in accessing
577   such systems, resulting in no direct measurements to date. Further, the existence of base-level

578   englacial streams and a perched water table are highly likely to complicate the detection of, and
579   distinction between, englacial and subglacial systems, at least approaching the terminus of DCGs.
580   Furthermore, the majority of reported DCGs terminate in ponds as a result of progressive surface
581   lowering (Section 3.3). This both increases the likelihood of some form of subglacial drainage but
582   at the same time reduces the likelihood of that system being channelised and severely hampers
583   its direct access. An additional complication arises from the high-elevation of some DCG source
584   areas, making it possible that the ice may be too cold throughout for water to penetrate to the
585   bed.

586   Nonetheless, remote sensing-based (Quincey et al., 2009) and field-based GPS
587   (Bartholomaus et al., 2008, 2011) studies of DCG surface velocities have inferred the occurrence
588   of basal sliding, which requires the presence of lubricating meltwater at the ice-bed interface.
589   While cold-based glaciers are frozen to the bed and move primarily by internal ice deformation
590   and creep (Glen, 1955; Nye, 1957; Weertman, 1983), glaciers with warm basal ice conversely have
591   water present at the bed, partly from pressure melting, and can move at greater speeds through
592   an additional basal motion component (Kamb, 1970; Nye, 1969; Weertman, 1957), either at the
593   rock-ice interface or from deformation of soft sediment (Boulton and Hindmarsh, 1987; Walder
594   and Fowler, 1994). Relatively rapid surface velocities, most notably in the central areas of glaciers
595   and in the summer months (when melting and rainfall delivery are greatest) (Figure 7) have been
596   recorded, for example, by Copland et al. (2009) who recorded a maximum velocity of >200 m a^{-1}
597   on the South Skamri Glacier, Pakistan Karakoram. Such velocity increases have been interpreted
598   as indicative of basal motion lubricated by the presence of subglacial meltwater (Copland et al.,
599   2009; Kääb, 2005; Kodama and Mae, 1976; Kraaijenbrink et al., 2016b; Kumar and Dobhal, 1997;
600   Mayer et al., 2006; Quincey et al., 2009). For some high-elevation DCGs, whose ablation areas
601   often include the base of ice falls, heavy crevassing may provide one route for water to access the
602   internal and basal drainage system of the glacier even if the ice is too cold for an englacial system
603   to reach the bed (Kodama and Mae, 1976). Although rare, such pathways can persist and lead to
604   the formation of moulins, which have occasionally been observed in the upper ablation areas of
605   glaciers, for example on Baltoro Glacier in the Pakistan Karakoram (Quincey et al., 2009), and
606   provide a direct connection to the bed. If meltwater can penetrate to the bed, it not only suggests
607   that the basal conditions are above the pressure melting point (at least locally), but that subsurface
608   hydrological systems are possible and even likely. However, to date very few measurements of the
609   internal or basal ice temperature of DCGs have been made (Section 4).

610   Further support for the existence of channelised subglacial drainage, at least near the
611   terminus of DCGs, is provided by the presence of single outlet channels at such glaciers. These also
612   discharge large volumes of heavily debris-laden water implying that the water had been
613   transported along the bed, entraining sediment (Quincey et al., 2009). On Ngozumpa Glacier, Benn
614   et al. (2017) interpreted spatially localised seasonal variations in glacier surface velocity as basal
615   sliding and inferred from this the presence of channelised subglacial drainage in the lower 10 km
616   of the glacier. Whether these fluctuations resulted from basal sliding and/or subglacial till
617   deformation is unknown in the absence of knowledge of subglacial conditions at the glacier (Benn
618   et al., 2017).

619   The existence of active subglacial drainage has additionally been inferred from bulk
620 meltwater analysis. Hasnain & Thayyen (1994) used EC measurements of the proglacial discharge
621 of Dokriani Bamak Glacier to argue for a perennially-active subglacial system that is interconnected
622 with the englacial system. On the same glacier, Hasnain et al. (2001) used dye-tracing studies to
623 investigate the subglacial drainage system, inferring a possible switch between an inefficient and
624 an efficient drainage system, as has been observed on lower-elevation alpine glaciers (Mair et al.,
625 2002; Nienow et al., 1998). The same methods were used on Gangotri Glacier, Garhwal Himalaya,
626 to show that an efficient channelised system exists at atmospheric pressure and develops through
627 the melt season with increasing meltwater inputs (Pottakkal et al., 2014). Wilson et al. (2016)
628 inferred a large amount of subglacial meltwater storage on Lirung Glacier, compared to the debris-
629 free Khimsung Glacier, Nepal Himalaya, due to the smaller magnitude of diurnal discharge
630 variability from the former. Bhatt et al. (2007) measured higher solute (Ca^{2+} and SO_4^{2-})
631 concentrations in the proglacial discharge of Lirung Glacier compared to those in the supraglacial
632 ponds on the glacier, inferring that these chemical species were acquired through contact with
633 reactive debris during subglacial drainage.

634   The particle-size distribution of sediment suspended within the proglacial stream of
635 Gangotri Glacier was interpreted in terms of the waterborne evacuation of subglacially-eroded
636 fines (Haritashya et al., 2010). Both the net flux and size of the suspended particles increased
637 through the melt season, implying that the glacier's drainage system became progressively more
638 competent and interconnected through the melt season.

639   Thus, although several lines of investigation point to the likely existence of subglacial
640 drainage beneath DCGs, evidence to date has been invaluable, but – in the absence of first-hand
641 access – necessarily inferential and ambiguous.

6. Proglacial hydrology

6.1 Proglacial lakes and GLOFs

644 Proglacial lakes predominantly form by a continuation of the processes of glacier thinning and
645 supraglacial pond growth, as described in Section 3.3. Perched supraglacial ponds grow both
646 downwards, eventually cutting to base-level, and laterally, with many lakes eventually coalescing
647 to produce one large lake above and over the terminus (Figure 8) (Kattelmann, 2003; Mertes et
648 al., 2016; Röhl, 2008; Watanabe et al., 2009). Base-level lakes that penetrate the full glacier
649 thickness can form farther upglacier and expand downglacier through the isolated stagnant
650 terminus ice, for example Imja Lake on Imja Glacier (Watanabe et al., 2009), though this is less
651 common and perhaps reflects stagnant ice towards the terminus acting as a flow impediment. The
652 exact location of such a proglacial lake may also be determined by the location of shallow englacial
653 conduits that provide pre-existing lines of weakness as the perched ponds grow (Benn et al., 2017;
654 Thompson et al., 2012). Proglacial lakes will therefore be at the hydrological base-level of the
655 glacier, and are often dammed by the terminal moraine (Thompson et al., 2012). With time, such
656 lakes continue to erode downwards into the ice, eventually reaching bedrock or basal sediment.
657 Hooker Lake, New Zealand Southern Alps (Figure 8), which initially formed in 1994 in front of
658 Hooker Glacier is now approximately 2.5 km long, 500 m wide, and has a maximum water depth
659 of 140 m (Robertson et al., 2012).

660   The formation of moraine-dammed proglacial lakes characterises a further and final stage
661 in the surface lowering and overall mass loss of DCGs. Benn et al. (2012) defined three stages in
662 the development of DCGs: in regime one, all parts of the glacier are dynamically active; in regime
663 two, surface lowering has begun and ice velocities decrease; in regime three, glaciers are
664 completely stagnant and rapid recession may occur. The formation of base-level lakes indicates
665 that a glacier has entered this third regime, and rapid recession may then occur through further
666 expansion of this proglacial lake (Benn et al., 2012). An increasing number of lakes of increasing
667 size have been observed in recent decades around the world (Carrivick and Tweed, 2013), for
668 example in the Caucasus Mountains (Stokes et al., 2007) and across the Himalaya (Gardelle et al.,
669 2011; Thompson et al., 2012). However, the pattern of proglacial lake formation has varied across
670 the Himalaya, with glacial lake coverage in the western Himalaya decreasing 30-50% from 1990-
671 2009 compared to an increased area of 20-65% in the eastern Himalaya, concurrent with the much
672 greater observed glacial mass loss in the former region over this period (Gardelle et al., 2011).

673   Proglacial lakes continue to expand through similar mechanisms to supraglacial ponds until
674 they are limited by subglacial topography, enhancing glacial mass loss and thus meltwater
675 production where the lake is underlain by ice (Carrivick and Tweed, 2013; Röhl, 2008). Initial
676 growth occurs through subaqueous melting and subaerial ice-face melting, causing both
677 deepening and areal growth, but once calving is triggered it becomes the dominant method of lake
678 growth (Röhl, 2008; Thompson et al., 2012). Calving from a proglacial lake progresses from notch-
679 development and roof collapse to large-scale, full-height slab calving that can substantially
680 increase mass loss from a glacier (Kirkbride and Warren, 1997; Thompson et al., 2012). If the lake
681 deepens to the glacier bed, allowing full-height slab calving, the lake may become unstable
682 because the water depth will be sufficient to trigger extending flow in the now-unsupported ice
683 cliff (Kirkbride and Warren, 1999; Thompson et al., 2012). This may weaken the ice by forming
684 crevasses, and allow the ice cliff to calve at a faster rate again; several kilometres of such rapid
685 calving was reported by Kirkbride & Warren (1999) for Tasman Glacier, New Zealand Southern
686 Alps. The process could also result in an upglacier expansion of the lake (Watanabe et al., 2009),
687 which may have implications for the glacier's drainage system, such as by earlier interruption of
688 meltwater routing (Carrivick and Tweed, 2013).

689   Very large proglacial lakes can alter a glacier's microclimate, due to a lake's lower albedo
690 and higher thermal heat capacity relative to surrounding ice and soil surfaces, producing relatively
691 cooler summer air temperatures and warmer autumn temperatures (Carrivick and Tweed, 2013).
692 This can slow summer ice ablation and consequently reduce the amount of meltwater being
693 produced and transported through the glacier, with implications for the development of englacial
694 and subglacial drainage systems. If a moraine-dammed proglacial lake is present then the
695 overwhelming majority of water transported through a DCG will pass through it (Benn et al., 2017).
696 This has implications for water drainage through the glacier, and for the potential occurrence of
697 glacial lake outburst floods (GLOFs) if the lake overflows or the dam is breached.

698   GLOFs can be a major hazard in regions such as the Andes and Himalaya, and can result in
699 fatalities as well as the destruction of land and infrastructure (Richardson and Reynolds, 2000;
700 Rounce et al., 2016). GLOFs can either occur through a breach of the dam or by dam failure. Dam
701 breach can be triggered by the increase of lake water-level and/or the creation of waves through:

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702   the addition of water from a lake higher up on the glacier (Buchroithner et al., 1982); an ice
703   avalanche (Vuichard and Zimmermann, 1987); a rock avalanche or mass movement entering the
704   lake (Harrison et al., 2006; Rounce et al., 2017); glacier calving (Kattelmann, 2003); rainfall events,
705   particularly during the monsoon (Kattelmann, 2003; Osti et al., 2011); or an earthquake-triggered
706   overtopping (Rounce et al. 2016).

707   The second mechanism by which a GLOF can occur is through dam failure, of either an ice-
708   cored or a sediment-cored moraine. Ice-cored moraine dams are inferred to be common features
709   at DCG proglacial lakes, as dead ice can be left beyond the glacier terminus as a result of both
710   glacier retreat and differential mass wasting (Richardson and Reynolds, 2000). Ice-cored moraines
711   degrade progressively by ablation beneath the debris layer and from the warmer lake water; this
712   accelerates once the ice is exposed and subjected to enhanced aerial melt, and the dam may finally
713   fail when water routes through relict glacial drainage features, such as voids, reducing the dam's
714   structural strength (Kattelmann, 2003; Richardson and Reynolds, 2000). Non-ice-cored moraines
715   are entirely composed of glacial sediment, and have been observed to destabilise and fail as a
716   landslide after rainfall or earthquake events (Osti *et al.* 2011). Waves generated by the dam breach
717   mechanism can also initiate rapid erosion of either type of moraine (Hubbard et al., 2005), possibly
718   eventually triggering moraine failure (Kattelmann, 2003).

719   The onset of DCG recession by rapid calving could allow major rock and debris avalanches
720   into a proglacial lake, which could trigger a GLOF and potentially destabilise a mountainside, with
721   the possibility of further hazards such as landslides and rockfalls (Hubbard et al., 2005; Kirkbride
722   and Warren, 1999). Risks from GLOFs can be mitigated, for example, by artificially lowering the
723   proglacial lake water-level (Rana et al., 2000), or monitored using on-site or remotely sensed data
724   (Bajracharya and Mool, 2009; Bolch et al., 2008; Nie et al., 2013; Rounce et al., 2016; Watson et
725   al., 2015). As an increasing number of receding glaciers form a proglacial lake that not only
726   withholds proglacial discharge, but dams it up against a potentially unstable moraine-dam, the
727   possibility of devastating GLOFs could rise (Carrivick and Tweed, 2013; Gardelle et al., 2011; Stokes
728   et al., 2007; Thompson et al., 2012).

6.2 Proglacial streams

730   Proglacial runoff from DCGs can form a significant proportion of the discharge of large rivers
731   downstream, particularly in High Mountain Asia: the Indus, Dudh Koshi, Ganges and Brahmaputra
732   rivers all stem from glacial meltwaters (Pritchard, 2017; Ragettli et al., 2015; Wilson et al., 2016).
733   Proglacial discharge measurements, estimates and models have been made across the Himalaya,
734   for example on individual glaciers in Nepal (Braun et al., 1993; Fujita and Sakai, 2014; Ragettli et
735   al., 2015; Rana et al., 1997; Savéan et al., 2015; Soncini et al., 2016; Tangborn and Rana, 2000),
736   Tibet (Kehrwald et al., 2008), the Tien Shan (Caiping and Yongjian, 2009; Han et al., 2010; Sorg et
737   al., 2012), India (Hasnain, 1996, 1999; Khan et al., 2017; Singh et al., 1995, 2005; Singh and
738   Bengtsson, 2004; Thayyen and Gergan, 2010), and for multiple catchments and entire regions
739   (Winiger et al., 2005). However, few of these measurements have been made for longer than a
740   decade: of the studies listed above, five measure discharge for a year or less; three have 2-3 years
741   of measurements; and only one has 6 years of measurements; the rest use modelling to obtain
742   estimates of proglacial discharge. Although Pritchard (2017) found that the glacial contribution to

743 seven river basins in the Himalaya is proportionally small (0.1-3.0%) it increases upstream and was
744 argued to be vital to support the freshwater needs of millions of people.

745 The presence of surface debris can have a notable effect on the proglacial discharge of a
746 DCG, resulting in a proglacial hydrograph that is different from that of a clean-ice glacier (Figure
747 9). For example, discharge both diurnally and through the ablation season are muted at debris-
748 covered Dome Glacier, Canadian Rockies, compared to neighbouring clean-ice Athabasca Glacier
749 (Figure 9), producing an annual variance in volumetric discharge of 1% compared to 24%
750 respectively (Mattson, 2000). This is due in part to the suppression of surface melt by a debris
751 cover, and in part to the lags that are induced as a result of the debris layer. On a clean-ice glacier,
752 the maximum melt rate occurs close to the time of maximum incoming solar radiation. Conversely,
753 on a DCG, the additional time to conduct heat through a debris layer and the warmer local air
754 temperatures due to the warming debris introduces a delay. Thus, peak melt can occur up to
755 several hours after the maximum radiation receipt at the debris surface (Carenzo et al., 2016;
756 Conway and Rasmussen, 2000; Evatt et al., 2015), and has been measured occurring up to 24 hours
757 later for debris layers >0.85 m thick (Fyffe et al., 2014). This lag in diurnal peak melt is thus reflected
758 in the timing of peak stream flow, producing a later and less pronounced peak in a proglacial
759 stream's diurnal pattern (Fyffe et al., 2014).

760 Lags in the proglacial discharge at DCGs are also caused by the temporary storage of water
761 within the debris layer. This has been observed during rainfall events and has been suggested to
762 influence discharge through both the subglacial and proglacial drainage networks by delaying and
763 buffering water transfer at the surface, potentially affecting basal water pressures and minimising
764 peaks in proglacial discharge (Brock et al., 2010). However, in the Himalaya, the monsoon
765 precipitation is thought to exert only a weak control on the proglacial discharge hydrograph of
766 glaciers unless the intensity is $>\sim$20 mm d^{-1}, which occurred on 20% of rainfall days during four
767 years of monsoon measurements (Thayyen et al., 2005). Early in the melt season, meltwater is
768 additionally stored within the snowpack of DCGs as well as within the debris layer year-round,
769 providing a further delay in the transport of meltwater from the surface into the subsurface
770 drainage system (Singh et al., 2006b). However, in the last two decades the amount of snowfall
771 accumulation has decreased across the Himalaya, and is projected to decrease a further 20-40%
772 by 2100 (Salerno et al., 2015; Viste and Sorteberg, 2015) which is highly likely to reduce this buffer
773 and influence the proglacial hydrograph pattern of DCGs in the future.

774 Groundwater storage within glacial catchments has been inferred to interact with
775 proglacial (and subglacial) stream networks, affecting the discharge patterns of the streams due
776 to additional water storage and subsequent release (Gremaud et al., 2009; Smart, 1988, 1996).
777 Andermann et al. (2012a) observed a lag between precipitation and discharge for 12 Himalayan
778 catchments (both glacierised and non-glacierised), indicating that up to two-thirds of the river
779 discharge is stored for approximately 45 days in a groundwater aquifer system before the
780 monsoon, greatly affecting the annual discharge pattern. This has been recorded in further studies
781 measuring SSC, with much lower concentrations measured post-monsoon once this groundwater
782 begins to be released and reduces the SSC of these rivers (Andermann et al., 2012b; Andermann
783 et al., 2012c). Such a significant effect of groundwater storage and release downriver from DCG
784 catchments would suggest that similar processes may occur beneath the glaciers themselves.

785   Other studies of glacierised limestone karst aquifers have used dye-tracing and modelling
786 to investigate links to the glacial drainage system. At Glacier de la Plaine Morte, Swiss Alps, this
787 showed that a greater proportion of the glacial meltwater was transported through a karst system
788 during the winter; in the summer, the karst capacity was exceeded and the excess water drained
789 through the glacier instead (Finger et al., 2013). A similar system in the Jade Dragon Snow
790 Mountain region of southwest China was studied for stable isotopes and modelled by Zeng *et al.*
791 (2015), showing that 29% of the glacier meltwater was transported into the karst aquifer.
792 Groundwater sinks of subglacial meltwater can therefore comprise a significant portion of the total
793 glacial output, potentially resulting in the glacial ablation being underestimated if this is not taken
794 into account.

795   As DCGs provide a significant source of water for large populations, quantifying future
796 runoff volumes is vital for planning and mitigating water resource issues. Models have been used
797 to predict future runoff from DCGs for a single glacier basin (Ragettli et al., 2015; Singh et al.,
798 2006a, 2008; Zhang et al., 2007), and multiple glacier basins (Immerzeel et al., 2012; Lowe and
799 Collins, 2001) up to a regional scale (Rees and Collins, 2006; Shea and Immerzeel, 2016),
800 investigating various future climatic scenarios. Currently, a large proportion of DCGs worldwide,
801 particularly in the Himalaya, have negative mass balances (Bolch et al., 2011, 2012; Kääb et al.,
802 2012; Scherler et al., 2011). The projected decrease in snowfall will additionally contribute to the
803 decreasing mass of these glaciers, both by reducing accumulation rates but also by exposing the
804 glacier surface to atmospheric melting earlier in the melt season (Salerno et al., 2015). Glacier
805 contributions to catchment discharge in many regions have been predicted to increase over the
806 next few decades, but as the glaciers continue to shrink, this proportion will begin to reduce
807 substantially due to the significantly smaller volume of glaciers remaining (Barnett et al., 2005;
808 Bolch, 2017; Bolch et al., 2012; Huss, 2011; Lutz et al., 2014). Shea & Immerzeel (2016) estimated
809 that most basins will have declining glacier contributions to streamflow by 2100, and water
810 shortage may then be a concern for many populated areas in the Karakoram, while peak flows may
811 represent a greater concern in the eastern Himalaya.

812   A further concern for future water supplies is the water quality provided by glacial
813 discharge, which is commonly assessed through measurements of the EC and SSC of proglacial
814 streams. Although based on simplified mixing models, studies have used proglacial stream SSC to
815 calculate the contribution of glacial systems to overall catchment sediment yields (Collins 1996;
816 Collins 1999; Hasnain & Thayyen 1999a; Singh et al. 2005; Haritashya et al. 2010). For example,
817 Collins (1996) determined from investigations at Batura Glacier, Karakoram Himalaya, that 40% of
818 the sediment yield of the Indus river, and 60% of the Hunza river are glacially-derived. Tectonic
819 uplift also contributes through enhanced weathering to the high sediment flux in these regions
820 (Collins, 1996). However, the glacially-derived proportion of total sediment yield can vary widely
821 with, in general, glaciers with more extensive subglacial systems and higher discharges
822 contributing greater amounts of sediment (Collins, 1999). Proglacial SSC therefore increases with
823 discharge during the ablation season, particularly with monsoon rainfall (Collins 1999; Hasnain &
824 Thayyen 1999a) when supraglacial debris weathering is enhanced and the increased discharge
825 flushes sediment through the system, increasing chemical weathering rates (Hasnain & Thayyen
826 1999b; Hodson et al. 2002) which may have implications on the water quality downstream as
827 discharge increases with glacier mass loss. Although the monsoon rains contribute to enhanced

828 sediment transport, they are not considered to affect weathering within the subglacial systems of
829 such glaciers, where sulphide oxidation and calcium carbonate dissolution dominate (Tranter et
830 al., 2002). On a diurnal scale, Kumar et al. (2009) found that the total ion concentration of
831 proglacial meltwater increased from the afternoon onwards, as the (inferred) englacial and
832 subglacial systems of Gangotri Glacier, became more active.

833   The water quality of proglacial runoff, including carbon export and other nutrient delivery
834 from glacial basins, exerts a critical influence on biogeochemical fluxes, ecosystem services,
835 downstream ecology and aquatic ecosystem biodiversity (Jacobsen et al., 2012). Ecological
836 responses are extremely sensitive to reductions in glacier area, with studies finding that freshwater
837 biodiversity in glacier-fed streams will decrease rapidly with the reduction (and ultimate
838 disappearance) of glacier area (Cauvy-Fraunié et al., 2016; Jacobsen et al., 2012; Milner et al.,
839 2009; Wilhelm et al., 2013). The potential loss of species is a key issue for future conservation and
840 the evolution of glaciers, particularly DCG, will have a large influence on any loss of species
841 (Jacobsen et al., 2012), an area of study largely beyond the remit of this review, but which deserves
842 further investigation.

7. Summary and future research priorities

844 The hydrology of DCGs is sufficiently distinctive to warrant bespoke treatment, separate from that
845 of clean-ice valley glaciers. This distinctiveness stems principally from the extensive and thick
846 debris cover on DCGs as well as, in many cases, their high elevation and local climate (such as the
847 South Asian monsoon) affecting the mass balance. These factors combine to produce a reverse
848 ablation gradient, where the point of maximum melt is located several kilometres up-glacier from
849 the terminus. In times of recession, a low angle, or even reversed, longitudinal surface profile
850 develops that is hummocky, promoting the surface storage of water in steep-sided supraglacial
851 ponds. These ponds serve to attenuate flows and regulate the outlet hydrograph. Additionally, in
852 contrast to their clean-ice counterparts, DCGs convey at least some of their surface water at the
853 ice-debris interface, likely as a thin film, and the debris layer itself can provide temporary water
854 storage that delays peak flow at the terminus.

855   Englacially, channels are likely formed through downcutting and/or the exploitation of
856 structural weaknesses, and the surface debris layer probably plays an important role in
857 determining the thermal characteristics of the upper part of the ice column. Subglacially, little is
858 known, but inferences point to the likely presence of water through the observation of seasonal
859 velocity speed-ups and bulk meltwater analysis. At the terminus, recent recession has resulted in
860 the development of moraine-impounded lakes, which are increasing in both number and size in
861 many areas of the world (Gardelle et al., 2011). Downstream, many millions of people rely on
862 glacially-sourced water for irrigation, power and sanitation, but a key gap remains in determining
863 the importance (in terms of quantity and quality) of meltwater as opposed to groundwater and
864 precipitation with increasing distance from the glacier terminus.

865   Despite the importance of glacially-sourced meltwater for many populations around the
866 world, knowledge of the hydrology of DCGs lags behind that of their clean-ice counterparts. In
867 particular, the subsurface hydrology of DCGs remains largely un-investigated and poorly
868 understood. Similarly, key parameters governing the formation and structure of these systems,

869 particularly thermal regime and basal conditions, are also largely unknown at DCGs. On the basis
870 of the above review, we summarise the current status of our understanding of the hydrology of
871 DCGs as a schematic illustration in Figure 10.

872 Inspection of Figure 10 reveals eight candidate areas for future hydrological research,
873 considered below:

874 1. **Water flow through and beneath the supraglacial debris layer**. Currently, there has been
875 minimal research into debris layer hydrology, whether it be a focus on water movement,
876 water storage, water chemistry, links to other parts of the glacier hydrological system, or
877 the removal of meltwater from the system through evaporation from the debris layer. Not
878 only is this important for considering potential delays within the drainage system due to
879 water storage and the impact upon thermal properties of the debris layer due to the role
880 moisture holds in dictating thermal conductivity, but it could also have an effect on water
881 quality downstream. It was noted in Section 5 that water flowing through debris or
882 sediment, particularly at the bed, entrains greater concentrations of solutes and SSC. With
883 flowpaths through debris at the surface increasing in extent as both the debris cover
884 continues to increase upglacier (Kirkbride and Warren, 1999; Stokes et al., 2007) and
885 meltwater production increases with warming temperatures, these solute levels could be
886 expected to be raised further, affecting proglacial water quality. Furthermore, the
887 hydrology of the sub-debris layer ice surface is likely to exert an important influence on
888 ablation, and thus the production of meltwater.

889 2. **Supraglacial pond hydrology**. Although substantial recent effort has been directed to the
890 study of supraglacial ponds and lakes at DCGs, the flow of water within these features, and
891 between them and other parts of the hydrological system, remains poorly understood, as
892 does their biogeochemistry (Bhatt et al., 2007; Takeuchi et al., 2012). Meltwater is stored
893 within supraglacial ponds and lakes - and as more, larger lakes form with greater future
894 meltwater production - this could delay outflow regimes both diurnally and seasonally.
895 How water is transported out of a lake is also poorly understood: are there supraglacial or
896 englacial links between ponds; if they are englacial, is all of this water transported to the
897 next pond or is some routed deeper into the glacier? As a result, this could influence the
898 development of englacial and subglacial drainage networks by altering the amount of water
899 that is, or can be, transported within the glacier.

900 3. **DCG thermal regime**. An almost complete lack of knowledge of the thermal regime of high-
901 elevation DCGs has resulted in a critically poor understanding of the existence and
902 character of englacial and subglacial drainage systems. If water cannot drain into such
903 glaciers it is unlikely that an englacial system can exist at all. Yet, it is unknown whether
904 englacial systems are entirely limited by the thermal regime of the ice.

905 4. **DCG englacial drainage.** Despite detailed glaciospeleological investigations, access has
906 limited these to large, open channels in accessible areas. Hence, little is known about active
907 englacial hydrology or smaller englacial hydrological pathways deeper within DCGs, or how
908 meltwater is transported from the supraglacial system into the glacier. The small scale
909 (microporous) movement of water between ice crystals has also received very little
910 attention and may form an important meltwater flowpath, for example through rotted
911 surface ice. At the larger scale, englacial drainage appears to be governed by base-levels,

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but controls over such levels and flow pathway configurations are poorly understood (they could be local or dictated by proglacial lake level), while englacial drainage below such levels (i.e. within the phreatic zone) remains un-investigated.

5. **DCG subglacial drainage.** Perhaps the greatest hydrological unknown of DCGs is that of the existence and character of subglacial drainage, which is critical to governing both ice motion and meltwater quality. Several indirect studies have suggested the existence of such effects, but no definitive evidence has yet been reported. If the presence of subglacial drainage is reported at high-elevation DCGs, then exploring the character and spatio-temporal variability of such drainage represents a key research priority.

6. **Groundwater flows.** While water loss from DCGs has been inferred, no study has yet reported on the mechanisms and rates of water transfer between a DCG's drainage system and that of the underlying substrate. It is therefore important to understand the proportion of river discharge that is provided by glacier meltwater and runoff, and how much is being stored within or immediately beyond the glacial drainage network at different time scales.

7. **Long-term water delivery from DCGs**. Long term records of proglacial discharge from DCGs are scarce, being limited to less than a decade of measurements for a small number of Himalayan DCGs. As DCGs are predicted to ablate more rapidly with the formation and growth of more supraglacial ponds and ice cliffs, discharge has been projected to increase in the short-term but decrease in the long-term, creating concerns for future water availability in many regions. A greater understanding of how DCGs are and will respond to the current and future warming climate would constrain future proglacial discharge volumes and thus help to mitigate water resource issues and other hazards such as potentially unstable moraine-dammed proglacial lakes.

8. **Local climate influence on DCG hydrology**. Regionally, the local climate is highly likely to have a substantial influence on the hydrological systems of DCGs, for example, monsoon-related weather. However, largely due to the inclement weather associated with monsoon precipitation at high elevations, the hydrological influence of the monsoon has not yet been addressed. Research to understand the role of monsoon conditions, and its relationship to non-monsoon conditions, is therefore required.

8. Author contribution

KM and BH planned the manuscript. KM led the manuscript writing and illustration with all co-authors contributing to specific sections.

9. Competing interests

The authors declare that they have no conflict of interest.

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11. References

Academica Sinica: Basic features of the glaciers of Mt. Jolmo Lungma Region, Southern part of the Tibet Autonomous Region, China, Sci. Sin., 18(1), 106–130, 1975.

Adhikary, S., Nakawo, M., Seko, K. and Shakya, B.: Dust influence on the melting process of glacier ice: experimental results from Lirung Glacier, Nepal Himalayas, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 43–52, International Association of Hydrological Sciences, Oxford. [online] Available from: http://books.google.com/books?hl=en&lr=&id=ghn8LV2CZ20C&oi=fnd&pg=PA43&dq=Dust+influence +on+the+melting+process+of+glacier+ice&ots=RssBAePMH3&sig=EPkFJdTtWOJ3F8fWog5LKpCJ3OY%5 Cnpapers2://publication/uuid/14FFD620-FE05-4094-92DE-FCD92BF61E6E, 2000.

Ageta, Y., Iwata, S., Yabuki, H., Naito, N., Sakai, A., Narama, C. and Karma: Expansion of glacier lakes in recent decades in the Bhutan Himalayas, in Debris-Covered Glaciers, vol. 264, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 165–175, International Association of Hydrological Sciences, Oxford., 2000.

Andermann, C., Crave, A., Gloaguen, R., Davy, P. and Bonnet, S.: Connecting source and transport: Suspended sediments in the Nepal Himalayas, Earth Planet. Sci. Lett., 351–352, 158–170, doi:10.1016/j.epsl.2012.06.059, 2012a.

Andermann, C., Longuevergne, L., Bonnet, S., Crave, A., Davy, P. and Gloaguen, R.: Impact of transient groundwater storage on the discharge of Himalayan rivers, Nat. Geosci., 5(2), 127–132, doi:10.1038/ngeo1356, 2012b.

Andermann, C., Bonnet, S., Crave, A., Davy, P., Longuevergne, L. and Gloaguen, R.: Sediment transfer and the hydrological cycle of Himalayan rivers in Nepal, Comptes Rendus Geosci., 344(11–12), 627–635, doi:10.1016/j.crte.2012.10.009, 2012c.

Anderson, L. S. and Anderson, R. S.: Modeling debris-covered glaciers: Response to steady debris deposition, Cryosph., 10(3), 1105–1124, doi:10.5194/tc-10-1105-2016, 2016.

Anderson, R.: A model of ablation-dominated medial moraines and the generation of debris-mantled glacier snouts, J. Glaciol., 46(154), 459–469, 2000.

Bajracharya, S. R. and Mool, P.: Glaciers, glacial lakes and glacial lake outburst floods in the Mount Everest region, Nepal, Ann. Glaciol., 50(53), 81–86, doi:10.3189/172756410790595895, 2009.

Baker, V. R.: Water and the martian landscape, Nature, 412(6843), 228–236, doi:10.1038/35084172, 2001.

Barnett, T. P., Adam, J. C. and Lettenmaier, D. P.: Potential impacts of a warming climate on water availability in snow-dominated regions, Nature, 438(7066), 303–309, doi:10.1038/nature04141, 2005.

Bartholomaus, T. C., Anderson, R. S. and Anderson, S. P.: Response of glacier basal motion to transient water storage, Nat. Geosci., 1(1), 33–37, doi:10.1038/ngeo.2007.52, 2008.

Bartholomaus, T. C., Anderson, R. S. and Anderson, S. P.: Growth and collapse of the distributed subglacial hydrologic system of Kennicott Glacier, Alaska, USA, and its effects on basal motion, J. Glaciol., 57(206), 985–1002, doi:10.3189/002214311798843269, 2011.

Benn, D., Gulley, J., Luckman, A., Adamek, A. and Glowacki, P. S.: Englacial drainage systems formed by

992 hydrologically driven crevasse propagation, J. Glaciol., 55(191), 513–523,
993 doi:10.3189/002214309788816669, 2009.

994 Benn, D. I., Wiseman, S. and Warren, C. R.: Rapid growth of a supraglacial lake, Ngozumpa Glacier, Khumbu
995 Himal, Nepal, IAHS Publ., 264(2), 177–185, 2000.

996 Benn, D. I., Wiseman, S. and Hands, K. A.: Growth and drainage of supraglacial lakes on debris-mantled
997 Ngozumpa Glacier, Khumbu Himal, Nepal, J. Glaciol., 47(159), 626–638,
998 doi:10.3189/172756501781831729, 2001.

999 Benn, D. I., Bolch, T., Hands, K., Gulley, J., Luckman, A., Nicholson, L. I., Quincey, D., Thompson, S., Toumi,
1000 R. and Wiseman, S.: Response of debris-covered glaciers in the Mount Everest region to recent warming,
1001 and implications for outburst flood hazards, Earth-Science Rev., 114(1–2), 156–174,
1002 doi:10.1016/j.earscirev.2012.03.008, 2012.

1003 Benn, D. I., Thompson, S., Gulley, J., Mertes, J., Luckman, A. and Nicholson, L.: Structure and evolution of
1004 the drainage system of a Himalayan debris-covered glacier, and its relationship with patterns of mass
1005 loss, Cryosph. Discuss., (March), 1–43, doi:10.5194/tc-2017-29, 2017.

1006 Bhatt, M. P., Masuzawa, T., Yamamoto, M. and Takeuchi, N.: Chemical characteristics of pond waters within
1007 the debris area of Lirung Glacier in Nepal Himalaya, J. Limnol., 66(2), 71–80, 2007.

1008 Blatter, H. and Hutter, K.: Polythermal conditions in Arctic glaciers, J. Glaciol., 37(126), 261–269,
1009 doi:http://adsabs.harvard.edu/abs/1990JGlac..37..261B, 1991.

1010 Bocchiola, D., Senese, A., Mihalcea, C., Mosconi, B., D'Agata, C., Smiraglia, C. and Diolaiuti, G.: An ablation
1011 model for debris-covered ice: The case study of Venerocolo glacier (Italian Alps), Geogr. Fis. e Din. Quat.,
1012 38(2), 113–128, doi:10.4461/GFDQ.2015.38.11, 2015.

1013 Bolch, T.: Hydrology: Asian glaciers are a reliable water source, Nature, 545(7653), 161–162,
1014 doi:10.1038/545161a, 2017.

1015 Bolch, T., Buchroithner, M. F., Peters, J., Baessler, M. and Bajracharya, S.: Identification of glacier motion
1016 and potentially dangerous glacial lakes in the Mt. Everest region/Nepal using spaceborne imagery, Nat.
1017 Hazards Earth Syst. Sci., 8(6), 1329–1340, doi:10.5194/nhess-8-1329-2008, 2008.

1018 Bolch, T., Pieczonka, T. and Benn, D. I.: Multi-decadal mass loss of glaciers in the Everest area (Nepal
1019 Himalaya) derived from stereo imagery, Cryosph., 5(2), 349–358, doi:10.5194/tc-5-349-2011, 2011.

1020 Bolch, T., Kulkarni, a., Kaab, a., Huggel, C., Paul, F., Cogley, J. G., Frey, H., Kargel, J. S., Fujita, K., Scheel, M.,
1021 Bajracharya, S. and Stoffel, M.: The State and Fate of Himalayan Glaciers, Science (80-.)., 336(6079),
1022 310–314, doi:10.1126/science.1215828, 2012.

1023 Boulton, G. S. and Hindmarsh, R. C. A.: Sediment deformation beneath glaciers: rheology and geological
1024 consequences, J. Geophys. Res., 92(B9), 9059--9082, 1987.

1025 Braun, L. N., Grabs, W. and Rana, B.: Application of a Conceptual Precipitation- Runoff Model in the
1026 Langtang Khola Basin , Nepal Himalaya, Snow Glacier Hydrol. (Proceedings Kathmandu Symp. Novemb.
1027 1992). IAHS Publ. no. 218,1993., 218(218), 221–237, 1993.

1028 Brock, B. W., Mihalcea, C., Kirkbride, M. P., Diolaiuti, G., Cutler, M. E. J. and Smiraglia, C.: Meteorology and
1029 surface energy fluxes in the 2005-2007 ablation seasons at the Miage debris-covered glacier, Mont Blanc
1030 Massif, Italian Alps, J. Geophys. Res. Atmos., 115(9), 1–16, doi:10.1029/2009JD013224, 2010.

1031 Brun, F., Buri, P., Miles, E. S., Wagnon, P., Steiner, J., Berthier, E., Ragettli, S., Kraaijenbrink, P., Immerzeel,
1032 W. W. and Pellicciotti, F.: Quantifying volume loss from ice cliffs on debris-covered glaciers using high-
1033 resolution terrestrial and aerial photogrammetry, J. Glaciol., 62(234), 684–695,
1034 doi:10.1017/jog.2016.54, 2016.

Buchroithner, M. F., Jentsch, G. and Wanivenhaus, B.: Monitoring of Recent Geological Events in the Khumbu Area (Himalaya, Nepal) by Digital Processing of Landsat MSS Data, Rock Mech., 15, 181–197, 1982.

Buri, P., Pellicciotti, F., Steiner, J. F., Miles, E. S. and Immerzeel, W. W.: A grid-based model of backwasting of supraglacial ice cliffs on debris-covered glaciers, Ann. Glaciol., 57(71), 199–211, doi:10.3189/2016AoG71A059, 2016a.

Buri, P., Miles, E. S., Steiner, J. F., Immerzeel, W. W., Wagnon, P. and Pellicciotti, F.: A physically-based 3-D model of ice cliff evolution on a debris-covered glacier, J. Geophys. Res. Earth Surf., 121, 2471–2493, doi:10.1002/2016JF004039, 2016b.

Caiping, C. and Yongjian, D.: The application of artificial neural networks to simulate meltwater runoff of Keqikaer Glacier, south slope of Mt. Tuomuer, western China, Environ. Geol., 57(8), 1839–1845, doi:10.1007/s00254-008-1471-1, 2009.

Carenzo, M., Pellicciotti, F., Mabillard, J., Reid, T. and Brock, B. W.: An enhanced temperature index model for debris-covered glaciers accounting for thickness effect, Adv. Water Resour., 94, 457–469, doi:10.1016/j.advwatres.2016.05.001, 2016.

Carrivick, J. L. and Tweed, F. S.: Proglacial Lakes: Character, behaviour and geological importance, Quat. Sci. Rev., 78, 34–52, doi:10.1016/j.quascirev.2013.07.028, 2013.

Casey, K. A., Kääb, A. and Benn, D. I.: Geochemical characterization of supraglacial debris via in situ and optical remote sensing methods: a case study in Khumbu Himalaya, Nepal, Cryosph., 6(1), 85–100, doi:10.5194/tc-6-85-2012, 2012.

Cauvy-Frauné, S., Andino, P., Espinosa, R., Calvez, R., Jacobsen, D. and Dangles, O.: Ecological responses to experimental glacier-runoff reduction in alpine rivers, Nat. Commun., 7(May), 12025, doi:10.1038/ncomms12025, 2016.

Chaturvedi, R. K., Kulkarni, A., Karyakarte, Y., Joshi, J. and Bala, G.: Glacial mass balance changes in the Karakoram and Himalaya based on CMIP5 multi-model climate projections, Clim. Change, 123(2), 315–328, doi:10.1007/s10584-013-1052-5, 2014.

Chikita, K., Jha, J. and Yamada, T.: The basin expansion mechanism of a supraglacial lake in the Nepal Himalaya, J. Hokkaido Univ. Fac. Sci. Ser. VII Geophys., 11(2), 501–521, 1998.

Chikita, K., Jha, J. and Yamada, T.: Sedimentary effects on the expansion of a Himalayan supraglacial lake, Glob. Planet. Change, 28(1–2), 23–34, doi:10.1016/S0921-8181(00)00062-X, 2001.

Chinn, T. J. H. and Dillon, A.: Observation on a debris-covered polar glacier "Whisky Glacier", James Ross Island, Antarctic Peninsula, Antarctica, J. Glaciol., 33(115), 300–310 [online] Available from: https://www.researchgate.net/publication/258964567_Observation_on_a_debris-covered_polar_glacier_'Whisky_Glacier'_James_Ross_Island_Antarctic_Peninsula_Antarctica, 1987.

Chu, V. W.: Greenland ice sheet hydrology: A review, Prog. Phys. Geogr., 38(1), 19–54, doi:10.1177/0309133313507075, 2014.

Clark, D., Clark, M. and Gillespie, A.: Debris-covered glaciers in the Sierra Nevada, California, and their implications for snowline reconstructions, Quat. Res., 41, 139–153, 1994.

Collier, E., Nicholson, L. I., Brock, B. W., Maussion, F., Essery, R. and Bush, A. B. G.: Representing moisture fluxes and phase changes in glacier debris cover using a reservoir approach, Cryosph., 8, 1429–1444, doi:10.5194/tc-8-1429-2014, 2014.

Collier, E., Maussion, F., Nicholson, L., Mölg, T., Immerzeel, W. and Bush, A.: Impact of debris cover on glacier ablation and atmosphere-glacier feedbacks in the Karakoram, Cryosph., 9, 1617–1632,

The Cryosphere
Discussions

1078   doi:10.5194/tc-9-1617-2015, 2015.

1079   Collins, D. N.: Sediment transport from glacierized basins in the Karakoram Mountains, IAHS Publ. Proc.
1080   Exet. Symp. July 1996, (236), 85–96 [online] Available from:
1081   http://books.google.com/books?hl=en&lr=&id=bZ-
1082   ufVQV5yAC&oi=fnd&pg=PA85&dq=Sediment+transport+from+glacierized+basins+in+th
1083   e+Karakoram+mountains&ots=u_RhK-Cx1T&sig=mPSIDAxp6cTfdAH3zojJEXj576Y, 1996.

1084   Collins, D. N.: Solute flux in meltwaters draining from a glacierized basin in the Karakoram mountains,
1085   Hydrol. Process., 13(18), 3001–3015, doi:10.1002/(SICI)1099-1085(19991230)13:18<3001::AID-
1086   HYP15>3.0.CO;2-N, 1999.

1087   Conway, H. and Rasmussen, L. A.: Summer temperature profiles within supraglacial debris on Khumbu
1088   Glacier, Nepal, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp.
1089   89–97, International Association of Hydrological Sciences, Oxford., 2000.

1090   Copland, L., Pope, S., Bishop, M. P., Shroder, J. F., Clendon, P., Bush, a. B. G., Kamp, U., Seong, Y. B. and
1091   Owen, L. a.: Glacier velocities across the Karakoram Himalaya, Ann. Glaciol., 50(52), 1–18,
1092   doi:10.3189/172756409789624229, 2009.

1093   Dunning, S. A., Rosser, N. J., Mccoll, S. T. and Reznichenko, N. V: Rapid sequestration of rock avalanche
1094   deposits within glaciers, Nat. Commun., 6(7964), 1–7, doi:10.1038/ncomms8964, 2015.

1095   Emmer, A., Loarte, E. C., Klimeš, J. and Vilímek, V.: Recent evolution and degradation of the bent Jatunraju
1096   glacier (Cordillera Blanca, Peru), Geomorphology, 228, 345–355, doi:10.1016/j.geomorph.2014.09.018,
1097   2015.

1098   Eriksson, M., Jianchu, X., Shrestha, A., Vaidya, R., Nepal, S. and Sandström, K.: The Changing Himalayas:
1099   Impact of climate change on water resources and livelihoods in the greater Himalayas, Int. Cent. Integr.
1100   Mt. Dev., 114, 1–28, doi:10.1144/SP312.3, 2009.

1101   Etzelmüller, E., Ødegård, R., Vatne, G., Mysterud, R., Tonning, T. and Sollid, J.: Glacier characteristics and
1102   sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen, Nor. Geogr. Tidsskr. -
1103   Nor. J. Geogr., 54(4), 157–168, 2001.

1104   Evatt, G. W., Abrahams, I. D., Heil, M., Mayer, C., Kingslake, J., Mitchell, S. L., Fowler, A. C. and Clark, C. D.:
1105   Glacial melt under a porous debris layer, J. Glaciol., 61(229), 825–836, doi:10.3189/2015JoG14J235,
1106   2015.

1107   Finger, D., Hugentobler, A., Huss, M., Voinesco, A., Wernli, H., Fischer, D., Weber, E., Jeannin, P. Y., Kauzlaric,
1108   M., Wirz, A., Vennemann, T., Hüsler, F., Schädler, B. and Weingartner, R.: Identification of glacial
1109   meltwater runoff in a karstic environment and its implication for present and future water availability,
1110   Hydrol. Earth Syst. Sci., 17(8), 3261–3277, doi:10.5194/hess-17-3261-2013, 2013.

1111   Foster, L. A., Brock, B. W., Cutler, M. E. J. and Diotri, F.: A physically based method for estimating supraglacial
1112   debris thickness from thermal band remote-sensing data, J. Glaciol., 58(210), 677–691,
1113   doi:10.3189/2012JoG11J194, 2012.

1114   Fountain, A. G. and Walder, J. S.: Water flow through temperate glaciers, Rev. Geophys., 36(97), 299,
1115   doi:10.1029/97RG03579, 1998.

1116   Fujita, K. and Sakai, A.: Modelling runoff from a Himalayan debris-covered glacier, Hydrol. Earth Syst. Sci.,
1117   18(7), 2679–2694, doi:10.5194/hess-18-2679-2014, 2014.

1118   Fyffe, C. L., Reid, T. D., Brock, B. W., Kirkbride, M. P., Diolaiuti, G., Smiraglia, C. and Diotri, F.: A distributed
1119   energy-balance melt model of an alpine debris-covered glacier, J. Glaciol., 60(221), 587–602,
1120   doi:10.3189/2014JoG13J148, 2014.

1121 Gades, A., Conway, H., Nereson, N., Naito, N. and Kadota, T.: Radio echo-sounding through supraglacial
1122   debris on Lirung and Khumbu Glaciers, Nepal Himalayas, in Debris-Covered Glaciers, vol. 264, edited by
1123   M. Nakawo, C. F. Raymond, and A. Fountain, pp. 13–22, International Association of Hydrological
1124   Sciences, Oxford., 2000.

1125 Gardelle, J., Arnaud, Y. and Berthier, E.: Contrasted evolution of glacial lakes along the Hindu Kush Himalaya
1126   mountain range between 1990 and 2009, Glob. Planet. Change, 75(1–2), 47–55,
1127   doi:10.1016/j.gloplacha.2010.10.003, 2011.

1128 Gibson, M., Glasser, N., Quincey, D., Rowan, A. and Irvine-Fynn, T.: Changes in glacier surface cover on
1129   Baltoro glacier, Karakoram, north Pakistan, 2001-2012, J. Maps, 13(2), 100–108,
1130   doi:10.1080/17445647.2016.1264319, 2017a.

1131 Gibson, M. J., Glasser, N. F., Quincey, D. J., Mayer, C., Rowan, A. V. and Irvine-Fynn, T. D. L.: Temporal
1132   variations in supraglacial debris distribution on Baltoro Glacier, Karakoram between 2001 and 2012,
1133   Geomorphology, 295, 572–585, doi:10.1016/j.geomorph.2017.08.012, 2017b.

1134 Glen, J.: The Creep of Polycrystalline Ice, Proc. R. Soc. A Math. Phys. Eng. Sci., 228(1175), 519–538, 1955.

1135 Greenwood, S. L., Clason, C. C., Helanow, C. and Margold, M.: Theoretical, contemporary observational and
1136   palaeo-perspectives on ice sheet hydrology: Processes and products, Earth-Science Rev., 155, 1–27,
1137   doi:10.1016/j.earscirev.2016.01.010, 2016.

1138 Gremaud, V., Goldscheider, N., Savoy, L., Favre, G. and Masson, H.: Geological structure, recharge processes
1139   and underground drainage of a glacierised karst aquifer system, Tsanfleuron-Sanetsch, Swiss Alps,
1140   Hydrogeol. J., 17(8), 1833–1848, doi:10.1007/s10040-009-0485-4, 2009.

1141 Gruber, S. and Haeberli, W.: Permafrost in steep bedrock slopes and its temperatures-related
1142   destabilization following climate change, J. Geophys. Res. Earth Surf., 112(2), 1–10,
1143   doi:10.1029/2006JF000547, 2007.

1144 Gruber, S., Hoelzle, M. and Haeberli, W.: Permafrost thaw and destabilization of Alpine rock walls in the hot
1145   summer of 2003, Geophys. Res. Lett., 31(13), 1–4, doi:10.1029/2004GL020051, 2004.

1146 Gulley, J. and Benn, D. I.: Structural control of englacial drainage systems in Himalayan debris-covered
1147   glaciers, J. Glaciol., 53(182), 399–412, doi:10.3189/002214307783258378, 2007.

1148 Gulley, J. D., Benn, D. I., Müller, D. and Luckman, A.: A cut-and-closure origin for englacial conduits in
1149   uncrevassed regions of polythermal glaciers, J. Glaciol., 55(189), 66–80,
1150   doi:10.3189/002214309788608930, 2009a.

1151 Gulley, J. D., Benn, D. I., Screaton, E. and Martin, J.: Mechanisms of englacial conduit formation and their
1152   implications for subglacial recharge, Quat. Sci. Rev., 28(19–20), 1984–1999,
1153   doi:10.1016/j.quascirev.2009.04.002, 2009b.

1154 Hambrey, M. J., Quincey, D. J., Glasser, N. F., Reynolds, J. M., Richardson, S. J. and Clemmens, S.:
1155   Sedimentological, geomorphological and dynamic context of debris-mantled glaciers, Mount Everest
1156   (Sagarmatha) region, Nepal, Quat. Sci. Rev., 28(11–12), 1084, doi:10.1016/j.quascirev.2009.04.009,
1157   2008.

1158 Han, H., Wang, J., Wei, J. and Liu, S.: Backwasting rate on debris-covered Koxkar glacier, Tuomuer mountain,
1159   China, J. Glaciol., 56(196), 287–296, doi:10.3189/002214310791968430, 2010.

1160 Hannah, D. M., Kansakar, S. R., Gerrard, A. J. and Rees, G.: Flow regimes of Himalayan rivers of Nepal: Nature
1161   and spatial patterns, J. Hydrol., 308(1–4), 18–32, doi:10.1016/j.jhydrol.2004.10.018, 2005.

1162 Haritashya, U. K., Kumar, A. and Singh, P.: Particle size characteristics of suspended sediment transported
1163   in meltwater from the Gangotri Glacier, central Himalaya - An indicator of subglacial sediment

1164   evacuation, Geomorphology, 122(1–2), 140–152, doi:10.1016/j.geomorph.2010.06.006, 2010.

1165   Harrison, S., Glasser, N., Winchester, V., Haresign, E., Warren, C. and Jansson, K.: A glacial lake outburst
1166   flood associated with recent mountain glacier retreat, Patagonian Andes, The Holocene, 16(4), 611–620,
1167   doi:10.1191/0959683606hl957rr, 2006.

1168   Hasnain, S. I.: Factors controlling suspended sediment transport in Himalayan glacier meltwaters, J. Hydrol.,
1169   181(1–4), 49–62, doi:10.1016/0022-1694(95)02917-6, 1996.

1170   Hasnain, S. I.: Runoff characteristics of a glacierized catchment, Garhwal Himalaya, India, Hydrol. Sci. J.,
1171   44(6), 847–854, doi:10.1080/02626669909492284, 1999.

1172   Hasnain, S. I. and Thayyen, R. J.: Hydrograph separation of bulk meltwaters of Dokriani Bamak glacier basin,
1173   based on electrical conductivity, Curr. Sci., 67(3), 189–193, 1994.

1174   Hasnain, S. I. and Thayyen, R. J.: Controls on the major-ion chemistry of the Dokriani glacier meltwaters,
1175   Ganga basin, Garhwal Himalaya, India, J. Glaciol., 45(149), 87–92, 1999a.

1176   Hasnain, S. I. and Thayyen, R. J.: Discharge and suspended-sediment concentration of meltwaters, draining
1177   from the Dokriani glacier, Garhwal Himalaya, India, J. Hydrol., 218(3–4), 191–198, doi:10.1016/S0022-
1178   1694(99)00033-5, 1999b.

1179   Hasnain, S. I., Jose, P. G., Ahmad, S. and Negi, D. C.: Character of the subglacial drainage system in the
1180   ablation area of Dokriani glacier, India, as revealed by dye-tracer studies, J. Hydrol., 248(1–4), 216–223,
1181   doi:10.1016/S0022-1694(01)00404-8, 2001.

1182   Head, J. W. and Marchant, D. R.: Cold-based mountain glaciers on Mars: Western Arsia Mons, Geology,
1183   31(7), 641–644, doi:10.1130/0091-7613(2003)031<0641:CMGOMW>2.0.CO;2, 2003.

1184   Hewitt, K., Wake, C. P., Young, G. J. and David, C.: Hydrological investigations at Biafo Glacier, Karakoram
1185   range, Himalaya: an important source of water for the Indus River, Ann. Glaciol., 13, 103–108, 1989.

1186   Hewitt, K., Clague, J. J. and Orwin, J. F.: Legacies of catastrophic rock slope failures in mountain landscapes,
1187   Earth-Science Rev., 87(1–2), 1–38, doi:10.1016/j.earscirev.2007.10.002, 2008.

1188   Hodson, A., Porter, P., Lowe, A. and Mumford, P.: Chemical denudation and silicate weathering in
1189   Himalayan glacier basins: Batura Glacier, Pakistan, J. Hydrol., 262(1–4), 193–208, doi:10.1016/S0022-
1190   1694(02)00036-7, 2002.

1191   Hubbard, B. and Nienow, P.: Alpine subglacial hydrology, Quat. Sci. Rev., 16(9), 939–955,
1192   doi:10.1016/S0277-3791(97)00031-0, 1997.

1193   Hubbard, B., Heald, A., Reynolds, J. M., Quincey, D., Richardson, S. D., Luyo, M. Z., Portilla, N. S. and
1194   Hambrey, M. J.: Impact of a rock avalanche on a moraine-dammed proglacial lake: Laguna Safuna Alta,
1195   Cordillera Blanca, Peru, Earth Surf. Process. Landforms, 30(10), 1251–1264, doi:10.1002/esp.1198, 2005.

1196   Hubbard, B., Milliken, R. E., Kargel, J. S., Limaye, A. and Souness, C.: Geomorphological characterisation and
1197   interpretation of a mid-latitude glacier-like form: Hellas Planitia, Mars, Icarus, 211(1), 330–346,
1198   doi:10.1016/j.icarus.2010.10.021, 2011.

1199   Hubbard, B., Souness, C. and Brough, S.: Glacier-like forms on Mars, Cryosph., 8, 2047–2061,
1200   doi:10.5194/tc-8-2047-2014, 2014.

1201   Huggel, C., Clague, J. J. and Korup, O.: Is climate change responsible for changing landslide activity in high
1202   mountains?, Earth Surf. Process. Landforms, 37(1), 77–91, doi:10.1002/esp.2223, 2012.

1203   Huss, M.: Present and future contribution of glacier storage change to runoff from macroscale drainage
1204   basins in Europe, Water Resour. Res., 47(7), 1–14, doi:10.1029/2010WR010299, 2011.

1205 Immerzeel, W. W., van Beek, L. P. H. and Bierkens, M. F. P.: Climate change will affect the Asian water
1206 towers, Science (80-.)., 328, 1382–5, doi:10.1126/science.1183188, 2010.

1207 Immerzeel, W. W., van Beek, L. P. H., Konz, M., Shrestha, A. B. and Bierkens, M. F. P.: Hydrological response
1208 to climate change in a glacierized catchment in the Himalayas, Clim. Change, 110(3–4), 721–736,
1209 doi:10.1007/s10584-011-0143-4, 2012.

1210 Immerzeel, W. W., Kraaijenbrink, P. D. A., Shea, J. M., Shrestha, A. B., Pellicciotti, F., Bierkens, M. F. P. and
1211 De Jong, S. M.: High-resolution monitoring of Himalayan glacier dynamics using unmanned aerial
1212 vehicles, Remote Sens. Environ., 150, 93–103, doi:10.1016/j.rse.2014.04.025, 2014.

1213 Inoue, J. and Yoshida, M.: Ablation and Heat Exchange over the Khumbu Glacier, J. Japanese Soc. Snow Ice,
1214 41(Glaciological Expedition of Nepal, Contribution No. 65), 26–33, doi:10.5331/seppyo.41.Special_26,
1215 1980.

1216 Irvine-Fynn, T. D. L., Hodson, A. J., Moorman, B. J., Vatne, G. and Hubbard, A. L.: Polythermal Glacier
1217 Hydrology: A Review, Rev. Geophys., 49(2010), 1–37, doi:10.1029/2010RG000350.1.INTRODUCTION,
1218 2011.

1219 Iwata, S., Watanabe, O. and Fushimi, H.: Surface morphology in the ablation area of the Khumbu Glacier, J.
1220 Japanese Soc. Snow Ice, 41(Special), 9–17, 1980.

1221 Iwata, S., Aoki, T., Kadota, T., Seko, K. and Yamaguchi, S.: Morphological evolution of the debris cover on
1222 Khumbu Glacier, Nepal, between 1978 and 1995, in Debris-Covered Glaciers, vol. 264, edited by M.
1223 Nakawo, C. F. Raymond, and A. Fountain, pp. 3–11, International Association of Hydrological Sciences,
1224 Oxford., 2000.

1225 Jacobsen, D., Milner, A. M., Brown, L. E. and Dangles, O.: Biodiversity under threat in glacier-fed
1226 river systems, Nat. Clim. Chang., 2(5), 361–364, doi:10.1038/nclimate1435, 2012.

1227 Janke, J. R., Bellisario, A. C. and Ferrando, F. A.: Classification of debris-covered glaciers and rock glaciers in
1228 the Andes of central Chile, Geomorphology, 241, 98–121, doi:10.1016/j.geomorph.2015.03.034, 2015.

1229 Jansson, P., Naslund, J.-O., Pettersson, R., Richardson-Naslund, C. and Holmlund, P.: Debris entrainment
1230 and polythermal structure in the terminus of Storglaciaren, in Debris Covered Glaciers, edited by M.
1231 Nakawo, C. F. Raymond, and A. Fountain, pp. 143–151, International Association of Hydrological
1232 Sciences, Oxford. [online] Available from:
1233 http://books.google.com/books?hl=en&lr=&id=ghn8LV2CZ20C&oi=fnd&pg=PA143&dq=Debris+entrain
1234 ment+and+polythermal+structure+in+the+terminus+of+Storglaciaren&ots=RtmyzeOLF-
1235 &sig=wsOei42_Aiq1m9x99RLHGZYFH80, 2000.

1236 Jansson, P., Hock, R. and Schneider, T.: The concept of glacier storage: a review, J. Hydrol., 282, 116–129,
1237 doi:10.1016/S0022-1694(03)00258-0, 2003.

1238 Jarosch, A. H. and Gudmundsson, M. T.: A numerical model for meltwater channel evolution in glaciers,
1239 Cryosph., 6(2), 493–503, doi:10.5194/tc-6-493-2012, 2012.

1240 Juen, M., Mayer, C., Lambrecht, A., Han, H. and Liu, S.: Impact of varying debris cover thickness on ablation:
1241 A case study for Koxkar Glacier in the Tien Shan, Cryosphere, 8(2), 377–386, doi:10.5194/tc-8-377-2014,
1242 2014.

1243 Kääb, A.: Combination of SRTM3 and repeat ASTER data for deriving alpine glacier flow velocities in the
1244 Bhutan Himalaya, Remote Sens. Environ., 94(4), 463–474, doi:10.1016/j.rse.2004.11.003, 2005.

1245 Kääb, A., Berthier, E., Nuth, C., Gardelle, J. and Arnaud, Y.: Contrasting patterns of early twenty-first-century
1246 glacier mass change in the Himalayas, Nature, 488(7412), 495–498, doi:10.1038/nature11324, 2012.

1247 Kamb, B.: Sliding motion of glaciers: Theory and observation, Rev. Geophys., 8(4), 673–728,

1248 doi:10.1029/RG008i004p00673, 1970.

1249 Karimi, N., Farokhnia, A., Karimi, L., Eftekhari, M. and Ghalkhani, H.: Combining optical and thermal remote
1250 sensing data for mapping debris-covered glaciers (Alamkouh Glaciers, Iran), Cold Reg. Sci. Technol., 71,
1251 73–83, doi:10.1016/j.coldregions.2011.10.004, 2012.

1252 Kattelmann, R.: Glacial lake outburst floods in the Nepal Himalaya: A manageable hazard?, Nat. Hazards,
1253 28(1), 145–154, doi:10.1023/A:1021130101283, 2003.

1254 Kayastha, R. B., Takeuchi, Y., Nakawo, M. and Ageta, Y.: Practical prediction of ice melting beneath various
1255 thickness of debris cover on Khumbu Glacier, Nepal, using a positive degree-day factor, in Debris-
1256 Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 71–81, International
1257 Association of Hydrological Sciences, Oxford., 2000.

1258 Kehrwald, N. M., Thompson, L. G., Tandong, Y., Mosley-Thompson, E., Schotterer, U., Alfimov, V., Beer, J.,
1259 Eikenberg, J. and Davis, M. E.: Mass loss on Himalayan glacier endangers water resources, Geophys. Res.
1260 Lett., 35(22), 2–7, doi:10.1029/2008GL035556, 2008.

1261 Khan, A. A., Pant, N. C., Sarkar, A., Tandon, S. K., Thamban, M. and Mahalinganathan, K.: The Himalayan
1262 cryosphere: A critical assessment and evaluation of glacial melt fraction in the Bhagirathi basin, Geosci.
1263 Front., 8(1), 107–115, doi:10.1016/j.gsf.2015.12.009, 2017.

1264 Kienholz, C., Herreid, S., Rich, J. L., Arendt, A. A., Hock, R. and Burgess, E. W.: Derivation and analysis of a
1265 complete modern-date glacier inventory for Alaska and northwest Canada, J. Glaciol., 61(227), 403–420,
1266 doi:10.3189/2015JoG14J230, 2015.

1267 King, O., Quincey, D. J., Carrivick, J. L. and Rowan, A. V.: Spatial variability in mass loss of glaciers in the
1268 Everest region, central Himalaya, between 2000 and 2015, Cryosph., 11, 407–426, doi:10.5194/tc-2016-
1269 99, 2017.

1270 Kirkbride, M.: Debris-Covered Glaciers, in Encyclopedia of Snow, Ice and Glaciers, edited by V. Singh, P.
1271 Singh, and U. Haritashya, pp. 180–182, Springer Netherlands., 2011.

1272 Kirkbride, M. P.: The temporal significance of transitions from melting to calving termini in the glaciers of
1273 the central Southern Alps of New Zealand, The Holocene, 3(3), 232–240, 1993.

1274 Kirkbride, M. P. and Deline, P.: The formation of supraglacial debris covers by primary dispersal from
1275 transverse englacial debris bands, Earth Surf. Process. Landforms, 38(15), 1779–1792,
1276 doi:10.1002/esp.3416, 2013.

1277 Kirkbride, M. P. and Warren, C. R.: Calving processes at a grounded ice cliff, Ann. Glaciol., 24, 116–121,
1278 1997.

1279 Kirkbride, M. P. and Warren, C. R.: Tasman Glacier, New Zealand: 20th-century thinning and predicted
1280 calving retreat, Glob. Planet. Change, 22(1–4), 11–28, doi:10.1016/S0921-8181(99)00021-1, 1999.

1281 Kodama, H. and Mae, S.: The Flow of Glaciers in the Khumbu Region, J. Japanese Soc. Snow Ice,
1282 38(Glaciological Expedition to Nepal, Contribution), 31–36, 1976.

1283 Komori, J.: Recent expansions of glacial lakes in the Bhutan Himalayas, Quat. Int., 184(1), 177–186,
1284 doi:10.1016/j.quaint.2007.09.012, 2008.

1285 Kraaijenbrink, P., Shea, J., Pellicciotti, F., de Jong, S. and Immerzeel, W.: Object-based analysis of unmanned
1286 aerial vehicle imagery to map and characterise surface features on a debris-covered glacier, Remote
1287 Sens. Environ., 186, 581–595, doi:10.1016/j.rse.2016.09.013, 2016a.

1288 Kraaijenbrink, P., Meijer, S. W., Shea, J. M., Pellicciotti, F., De Jong, S. M. and Immerzeel, W. W.: Seasonal
1289 surface velocities of a Himalayan glacier derived by automated correlation of unmanned aerial vehicle

1290   imagery, Ann. Glaciol., 57(71), 103–113, doi:10.3189/2016AoG71A072, 2016b.

1291   Kraaijenbrink, P. D. A., Bierkens, M. F. P., Lutz, A. F. and Immerzeel, W. W.: Impact of a global temperature
1292   rise of 1.5 degrees Celsius on Asia's glaciers, Nature, 549(7671), 257–260, doi:10.1038/nature23878,
1293   2017.

1294   Kumar, K., Miral, M. S., Joshi, S., Pant, N., Joshi, V. and Joshi, L. M.: Solute dynamics of meltwater of Gangotri
1295   glacier, Garhwal Himalaya, India, Environ. Geol., 58(6), 1151–1159, doi:10.1007/s00254-008-1592-6,
1296   2009.

1297   Kumar, S. and Dobhal, D. P.: Climatic effects and bedrock control on rapid fluctuations of Chhota Shigri
1298   glacier, northwest Himalaya, India, J. Glaciol., 43(145), 467–472, 1997.

1299   Lambrecht, A., Mayer, C., Hagg, W., Popovnin, V., Rezepkin, A., Lomidze, N. and Svanadze, D.: A comparison
1300   of glacier melt on debris-covered glaciers in the northern and southern Caucasus, Cryosph., 5, 525–538,
1301   doi:10.5194/tc-5-525-2011, 2011.

1302   Langston, G., Bentley, L. R., Hayashi, M., Mcclymont, A. and Pidlisecky, A.: Internal structure and
1303   hydrological functions of an alpine proglacial moraine, Hydrol. Process., 25(19), 2967–2982,
1304   doi:10.1002/hyp.8144, 2011.

1305   Lejeune, Y., Bertrand, J., Wagnon, P. and Morin, S.: A physically based model of the year-round surface
1306   energy   and   mass   balance   of   debris-covered   glaciers,   J.   Glaciol.,   59(214),   327–344,
1307   doi:10.3189/2013JoG12J149, 2013.

1308   Levy, J. S., Marchant, D. R. and Head, J. W.: Distribution and origin of patterned ground on Mullins Valley
1309   debris-covered glacier, Antarctica: The roles of ice flow and sublimation, Antarct. Sci., 18(3), 385–397,
1310   doi:10.1002/dat.20032, 2006.

1311   Liu, F., Williams, M. W. and Caine, N.: Source waters and flow paths in an alpine catchment, Colorado Front
1312   Range, United States, Water Resour. Res., 40(9), 1–16, doi:10.1029/2004WR003076, 2004.

1313   Lowe, A. T. and Collins, D. N.: Modelling runoff from large glacierized basins in the Karakoram Himalaya
1314   using remote sensing of the transient snowline, Remote Sens. Hydrol., (267), 99–104, 2001.

1315   Lukas, S., Nicholson, L., Ross, F. and Humlum, O.: Formation, Meltout Processes and Landscape Alteration
1316   of High-Arctic Ice-Cored Moraines - Examples from Nordenskiold Land, Central Spitsbergen, Polar
1317   Geogr., 29(3), 157–187, 2005.

1318   Lutz, A. F., Immerzeel, W. W., Shrestha, A. B. and Bierkens, M. F. P.: Consistent increase in High Asia's runoff
1319   due   to   increasing   glacier   melt   and   precipitation,   Nat.   Clim.   Chang.,   4(7),   587–592,
1320   doi:10.1038/nclimate2237, 2014.

1321   Mackay, S., Marchant, D., Lamp, J. and Head, J.: Cold-based debris-covered glaciers: Evaluating their
1322   potential as climate archives through studies of ground-penetrating radar and surface morphology, J.
1323   Geophys. Res. Earth Surf., 119, 2505–2540, doi:10.1002/2014JF003178.Received, 2014.

1324   Mae, S.: Ice Temperature of Khumbu Glacier, J. Japanese Soc. Snow Ice, 38(Special), 37–38, 1976.

1325   Mae, S., Wushiki, H., Ageta, Y. and Higuchi, K.: Thermal Drilling and Temperature Measurements in Khumbu
1326   Glacier, Nepal Measurements Himalayas, J. Japanese Soc. Snow Ice, 37(4), 161–169, 1975.

1327   Mair, D., Nienow, P., Sharp, M., Wohlleben, T. and Willis, I.: Influence of subglacial drainage system
1328   evolution on glacier surface motion: Haut Glacier d'Arolla, Switzerland, J. Geophys. Res., 107(B8, 2175),
1329   1–13, doi:10.1029/2001JB000514, 2002.

1330   Marston, R.: Supraglacial Stream Dynamics on the Juneau Icefield, Ann. Assoc. Am. Geogr., 73(4), 597–608,
1331   1983.

1332 Mattson, L. E.: The influence of a debris cover on the midsummer discharge of Dome Glacier, Canadian
1333   Rocky Mountains, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp.
1334   25–33, International Association of Hydrological Sciences, Oxford. [online] Available from:
1335   http://books.google.co.uk/books?hl=en&lr=&id=ghn8LV2CZ20C&oi=fnd&pg=PA25&dq=energy+excha
1336   nges+and+ablation+rates+on+the+debris+covered&ots=RtmzubOMH6&sig=riB-
1337   kZ2Yl9IMB5Ak5ceYpjyFhy8, 2000.

1338 Mayer, C., Lambrecht, A., Belò, M., Smiraglia, C. and Diolaiuti, G.: Glaciological characteristics of the ablation
1339   zone of Baltoro glacier, Karakoram, Pakistan, Ann. Glaciol., 43, 123–131,
1340   doi:10.3189/172756406781812087, 2006.

1341 McCarthy, M., Pritchard, H., Willis, I. A. N. and King, E.: Ground-penetrating radar measurements of debris
1342   thickness on Lirung Glacier, Nepal, J. Glaciol., 63(239), 543–555, doi:10.1017/jog.2017.18, 2017.

1343 Mertes, J. R., Thompson, S. S., Booth, A. D., Gulley, J. D. and Benn, D. I.: A conceptual model of supraglacial
1344   lake formation on debris-covered glaciers based on GPR facies analysis., Earth Surf. Process. Landforms,
1345   doi:10.1002/esp.4068, 2016.

1346 Mihalcea, C., Mayer, C., Diolaiuti, G., Lambrecht, A., Smiraglia, C. and Tartari, G.: Ice ablation and
1347   meteorological conditions on the debris-covered area of Baltoro glacier, Karakoram, Pakistan, Ann.
1348   Glaciol., 43(1894), 292–300, doi:10.3189/172756406781812104, 2006.

1349 Mihalcea, C., Brock, B. W., Diolaiuti, G., D'Agata, C., Citterio, M., Kirkbride, M. P., Cutler, M. E. J. and
1350   Smiraglia, C.: Using ASTER satellite and ground-based surface temperature measurements to derive
1351   supraglacial debris cover and thickness patterns on Miage Glacier (Mont Blanc Massif, Italy), Cold Reg.
1352   Sci. Technol., 52(3), 341–354, doi:10.1016/j.coldregions.2007.03.004, 2008.

1353 Miles, E. S., Pellicciotti, F., Willis, I. C., Steiner, J. F., Buri, P. and Arnold, N. S.: Refined energy-balance
1354   modelling of a supraglacial pond, Langtang Khola, Nepal, Ann. Glaciol., 57(71), 29–40,
1355   doi:10.3189/2016AoG71A421, 2016a.

1356 Miles, E. S., Willis, I. C., Arnold, N. S., Steiner, J. F. and Pellicciotti, F.: Spatial, seasonal, and interannual
1357   variability of supraglacial ponds in the Langtang Valley of Nepal, 1999 to 2013, J. Glaciol., 1–18,
1358   doi:10.1017/jog.2016.120, 2016b.

1359 Miles, E. S., Steiner, J., Willis, I. C., Buri, P., Immerzeel, W. W., Chesnokova, A. and Pellicciotti, F.: Pond
1360   dynamics and supraglacial-englacial connectivity on debris-covered Lirung Glacier, Front. Earth Sci.,
1361   doi:10.3389/feart.2017.00069, 2017.

1362 Milner, A., Brown, L. and Hannah, D.: Hydroecological response of river systems to shrinking glaciers,
1363   Hydrol. Process., 23, 62–77, doi:10.1002/hyp, 2009.

1364 Nagai, H., Fujita, K., Nuimura, T. and Sakai, A.: Southwest-facing slopes control the formation of debris-
1365   covered glaciers in the Bhutan Himalaya, Cryosph., 7(4), 1303–1314, doi:10.5194/tc-7-1303-2013, 2013.

1366 Narama, C., Daiyrov, M., Tadono, T., Yamamoto, M., Kääb, A., Morita, R. and Jinro, U.: Seasonal drainage of
1367   supraglacial lakes on debris-covered glaciers in the Tien Shan Mountains, Central Asia, Geomorphology,
1368   doi:10.1016/j.geomorph.2017.03.002, 2017.

1369 Nicholson, L. and Benn, D. I.: Calculating ice melt beneath a debris layer using meteorological data, J.
1370   Glaciol., 52(178), 463–470, doi:10.3189/172756506781828584, 2006.

1371 Nicholson, L. and Benn, D. I.: Properties of natural supraglacial debris in relation to modelling sub-debris ice
1372   ablation, Earth Surf. Process. Landforms, 38(5), 490–501, doi:10.1002/esp.3299, 2013.

1373 Nie, Y., Liu, Q. and Liu, S.: Glacial lake expansion in the Central Himalayas by landsat images, 1990-2010,
1374   PLoS One, 8(12), 1–8, doi:10.1371/journal.pone.0083973, 2013.

1375 Nienow, P., Sharp, M. and Willis, I.: Seasonal changes in the morphology of the subglacial drainage system,
1376   Haut Glacier d'Arolla, Switzerland, Earth Surf. Process. Landforms, 23(9), 825–843,
1377   doi:10.1002/(SICI)1096-9837(199809)23:9<825::AID-ESP893>3.0.CO;2-2, 1998.

1378 Nuimura, T., Fujita, K., Yamaguchi, S. and Sharma, R. R.: Elevation changes of glaciers revealed by
1379   multitemporal digital elevation models calibrated by GPS survey in the Khumbu region, Nepal Himalaya,
1380   1992-2008, J. Glaciol., 58(210), 648–656, doi:10.3189/2012JoG11J061, 2012.

1381 Nye, J.: The Distribution of Stress and Velocity in Glaciers and Ice-Sheets, Proc. R. Soc. A Math. Phys. Eng.
1382   Sci., 239(1216), 113–133, 1957.

1383 Nye, J. F.: A Calculation on the Sliding of Ice Over a Wavy Surface Using a Newtonian Viscous Approximation,
1384   Proc. R. Soc. A Math. Phys. Eng. Sci., 311(1506), 445–467, doi:10.1098/rspa.1969.0127, 1969.

1385 Osti, R., Bhattarai, T. N. and Miyake, K.: Causes of catastrophic failure of Tam Pokhari moraine dam in the
1386   Mt. Everest region, Nat. Hazards, 58(3), 1209–1223, doi:10.1007/s11069-011-9723-x, 2011.

1387 Østrem, G.: Ice Melting under a Thin Layer of Moraine, and the Existence of Ice Cores in Moraine Ridges,
1388   Geogr. Ann., 41(4), 228–230, 1959.

1389 Paul, F., Huggel, C. and Kääb, A.: Combining satellite multispectral image data and a digital elevation model
1390   for mapping debris-covered glaciers, Remote Sens. Environ., 89(4), 510–518,
1391   doi:10.1016/j.rse.2003.11.007, 2004.

1392 Pellicciotti, F., Stephan, C., Miles, E., Herreid, S., Immerzeel, W. W. and Bolch, T.: Mass-balance changes of
1393   the debris-covered glaciers in the Langtang Himal, Nepal, from 1974 to 1999, J. Glaciol., 61(226), 373–
1394   386, doi:10.3189/2015JoG13J237, 2015.

1395 Pelto, M. S.: Mass balance of adjacent debris-covered and clean glacier ice in the North Cascades,
1396   Washington, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 35–
1397   42, International Association of Hydrological Sciences, Oxford., 2000.

1398 Pieczonka, T., Bolch, T., Junfeng, W. and Shiyin, L.: Heterogeneous mass loss of glaciers in the Aksu-Tarim
1399   Catchment (Central Tien Shan) revealed by 1976 KH-9 Hexagon and 2009 SPOT-5 stereo imagery,
1400   Remote Sens. Environ., 130, 233–244, doi:10.1016/j.rse.2012.11.020, 2013.

1401 Planet Team: Planet Application Program Interface: In Space for Life on Earth, [online] Available from:
1402   https://api.planet.com, 2017.

1403 Pottakkal, J. G., Ramanathan, A., Singh, V. B., Sharma, P., Azam, M. F. and Linda, A.: Characterization of
1404   subglacial pathways draining two tributary meltwater streams through the lower ablation zone of
1405   Gangotri glacier system, Garhwal Himalaya, India, Curr. Sci., 107(4), 613–621, 2014.

1406 Pritchard, H. D.: Asia's glaciers are a regionally important buffer against drought, Nature, 545(7653), 169–
1407   174, doi:10.1038/nature22062, 2017.

1408 Qiao, L., Mayer, C. and Liu, S.: Distribution and interannual variability of supraglacial lakes on debris-covered
1409   glaciers in the Khan Tengri-Tumor Mountains, Central Asia, Environ. Res. Lett., 10(1), 1–10,
1410   doi:10.1088/1748-9326/10/1/014014, 2015.

1411 Quincey, D. J., Richardson, S. D., Luckman, A., Lucas, R. M., Reynolds, J. M., Hambrey, M. J. and Glasser, N.
1412   F.: Early recognition of glacial lake hazards in the Himalaya using remote sensing datasets, Glob. Planet.
1413   Change, 56(1–2), 137–152, doi:10.1016/j.gloplacha.2006.07.013, 2007.

1414 Quincey, D. J., Copland, L., Mayer, C., Bishop, M., Luckman, A. and Belò, M.: Ice velocity and climate
1415   variations for Baltoro Glacier, Pakistan, J. Glaciol., 55(194), 1061–1071,
1416   doi:10.3189/002214309790794913, 2009.

1417 Racoviteanu, A. E., Arnaud, Y., Williams, M. W. and Ordoñez, J.: Decadal changes in glacier parameters in
1418   the Cordillera Blanca, Peru, derived from remote sensing, J. Glaciol., 54(186), 499–510, 2008.

1419 Radić, V., Bliss, A., Beedlow, A. C., Hock, R., Miles, E. and Cogley, J. G.: Regional and global projections of
1420   twenty-first century glacier mass changes in response to climate scenarios from global climate models,
1421   Clim. Dyn., 42(1–2), 37–58, doi:10.1007/s00382-013-1719-7, 2014.

1422 Ragettli, S., Pellicciotti, F., Immerzeel, W. W., Miles, E. S., Petersen, L., Heynen, M., Shea, J. M., Stumm, D.,
1423   Joshi, S. and Shrestha, A.: Unraveling the hydrology of a Himalayan catchment through integration of
1424   high resolution in situ data and remote sensing with an advanced simulation model, Adv. Water Resour.,
1425   78, 94–111, doi:10.1016/j.advwatres.2015.01.013, 2015.

1426 Ragettli, S., Immerzeel, W. W. and Pellicciotti, F.: Contrasting climate change impact on river flows from
1427   high-altitude catchments in the Himalayan and Andes Mountains, Proc. Natl. Acad. Sci. U. S. A., 113(33),
1428   9222–9227, doi:10.1073/pnas.1606526113, 2016a.

1429 Ragettli, S., Bolch, T. and Pellicciotti, F.: Heterogeneous glacier thinning patterns over the last 40 years in
1430   Langtang Himal, Nepal, Cryosph., 10(5), 2075–2097, doi:10.5194/tc-10-2075-2016, 2016b.

1431 Rana, B., Nakawo, M., Fukushima, Y. and Ageta, Y.: Application of a conceptual precipitation-runoff model
1432   (HYCYMODEL) in a debris-covered glacierized basin in the Langtang Valley, Nepal Himalaya, Ann. Glaciol.,
1433   25(199 0), 226–231, 1997.

1434 Rana, B., Shrestha, A. B., Reynolds, J. M., Aryal, R., Pokhrel, A. P. and Budhathoki, K. P.: Hazard assessment
1435   of the Tsho Rolpa Glacier Lake and ongoing, J. Nepal Geol. Soc., 22, 563–570, 2000.

1436 Raymond, C. F. C. and Nolan, M.: Drainage of a glacial lake through an ice spillway, in Symposium: Debris-
1437   Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 199–210, International
1438   Association   of   Hydrological   Sciences,   Oxford.   [online]   Available   from:
1439   http://books.google.com/books?hl=en&lr=&id=ghn8LV2CZ20C&oi=fnd&pg=PA199&dq=Drainage+of+a
1440   +glacial+lake+through+an+ice+spillway&ots=RsuGBdMSH6&sig=UxOhTWO5tWlE4-kW4eg3PjGNxsg,
1441   2000.

1442 Rees, G. and Collins, D.: Regional differences in response of flow in glacier-fed Himalayan rivers to climatic
1443   warming, Hydrol. Process., 20, 2157–2169, doi:10.1002/hyp, 2006.

1444 Reid, T. D. and Brock, B. W.: An energy-balance model for debris-covered glaciers including heat conduction
1445   through the debris layer, J. Glaciol., 56(199), 903–916, 2010.

1446 Reid, T. D. and Brock, B. W.: Assessing ice-cliff backwasting and its contribution to total ablation of debris-
1447   covered Miage glacier, Mont Blanc massif, Italy, J. Glaciol., 60(219), 3–13, doi:10.3189/2014JoG13J045,
1448   2014.

1449 Reynolds, J. M.: On the formation of supraglacial lakes on debris-covered glaciers, Debris-covered glaciers,
1450   264(264), 153–161, 2000.

1451 Richardson, S. D. and Reynolds, J. M.: Degradation of ice-cored moraine dams: implications for hazard
1452   development, Debris-covered glaciers Proc. an Int. Work. held Univ. Washingt. Seattle, Washington,
1453   USA, (264), 187–197, 2000.

1454 Rist, A. and Phillips, M.: First results of investigations on hydrothermal processes within the active layer
1455   above alpine permafrost in steep terrain, Nor. Geogr. Tidsskr. - Nor. J. Geogr., 59(2), 177–183,
1456   doi:10.1080/00291950510020574, 2005.

1457 Rivera, A., Benham, T., Casassa, G., Bamber, J. and Dowdeswell, J. A.: Ice elevation and areal changes of
1458   glaciers from the Northern Patagonia Icefield, Chile, Glob. Planet. Change, 59(1–4), 126–137,
1459   doi:10.1016/j.gloplacha.2006.11.037, 2007.

1460 Robertson, C. M., Benn, D. I., Brook, M. S., Fuller, I. C. and Holt, K. A.: Subaqueous calving margin
1461   morphology at Mueller, Hooker and Tasman glaciers in Aoraki/Mount Cook National Park, New Zealand,
1462   J. Glaciol., 58(212), 1037–1046, doi:10.3189/2012JoG12J048, 2012.

1463 Röhl, K.: Thermo-erosional notch development at fresh-water-calving Tasman Glacier, New Zealand, J.
1464   Glaciol., 52(177), 203–213, doi:10.3189/172756506781828773, 2006.

1465 Röhl, K.: Characteristics and evolution of supraglacial ponds on debris-covered Tasman Glacier, New
1466   Zealand, J. Glaciol., 54(188), 867–880, doi:10.3189/002214308787779861, 2008.

1467 Rounce, D., Byers, A. C., Byers, E. A. and McKinney, D. C.: Brief Communications: Observations of a glacier
1468   outburst flood from Lhotse Glacier, Everest area, Nepal, Cryosph., 11, 443–449, doi:10.5194/tc-2016-
1469   239, 2017.

1470 Rounce, D. R., Quincey, D. J. and McKinney, D. C.: Debris-covered glacier energy balance model for Imja-
1471   Lhotse Shar Glacier in the Everest region of Nepal, Cryosph., 9(6), 2295–2310, doi:10.5194/tc-9-2295-
1472   2015, 2015.

1473 Rounce, D. R., McKinney, D. C., Lala, J. M., Byers, A. C. and Watson, C. S.: A New Remote Hazard and Risk
1474   Assessment Framework for Glacial Lakes in the Nepal Himalaya, Hydrol. Earth Syst. Sci., (20), 3455–3475,
1475   doi:10.5194/hess-2016-161, 2016.

1476 Rowan, A. V., Egholm, D. L., Quincey, D. J. and Glasser, N. F.: Modelling the feedbacks between mass
1477   balance, ice flow and debris transport to predict the response to climate change of debris-covered
1478   glaciers in the Himalaya, Earth Planet. Sci. Lett., 430, 427–438, doi:10.1016/j.epsl.2015.09.004, 2015.

1479 Roy, J. W. and Hayashi, M.: Multiple, distinct groundwater flow systems of a single moraine-talus feature in
1480   an alpine watershed, J. Hydrol., 373(1–2), 139–150, doi:10.1016/j.jhydrol.2009.04.018, 2009.

1481 Sakai, A.: Glacial Lakes in the Himalayas : A Review on Formation and Expansion Processes, Glob. Environ.
1482   Res., 16(2011), 23–30, 2012.

1483 Sakai, A. and Fujita, K.: Formation conditions of supraglacial lakes on debris-covered glaciers in the
1484   Himalaya, J. Glaciol., 56(195), 9249–9251, 2010.

1485 Sakai, A., Nakawo, M. and Fujita, K.: Melt rate of ice cliffs on the Lirung Glacier, Nepal Himalayas, 1996, Bull.
1486   Glacier Res., 16(October), 57–66, 1998.

1487 Sakai, A., Takeuchi, N., Fujita, K. and Nakawo, M.: Role of supraglacial ponds in the ablation process of a
1488   debris-covered glacier in the Nepal Himalayas, in Debris-Covered Glaciers, edited by M. Nakawo, C. F.
1489   Raymond, and A. Fountain, pp. 119–130, International Association of Hydrological Sciences, Oxford.,
1490   2000.

1491 Sakai, A., Nakawo, M. and Fujita, K.: Distribution Characteristics and Energy Balance of Ice Cliffs on Debris-
1492   Covered Glaciers, Nepal Himalaya, Arctic, Antarct. Alp. Res., 34(1), 12–19, 2002.

1493 Sakai, A., Nishimura, K., Kadota, T. and Takeuchi, N.: Onset of calving at supraglacial lakes on debris-covered
1494   glaciers of the Nepal Himalaya, J. Glaciol., 55(193), 909–917, doi:10.3189/002214309790152555, 2009.

1495 Salerno, F., Thakuri, S., D'Agata, C., Smiraglia, C., Manfredi, E. C., Viviano, G. and Tartari, G.: Glacial lake
1496   distribution in the Mount Everest region: Uncertainty of measurement and conditions of formation,
1497   Glob. Planet. Change, 92–93, 30–39, doi:10.1016/j.gloplacha.2012.04.001, 2012.

1498 Salerno, F., Guyennon, N., Thakuri, S., Viviano, G., Romano, E., Vuillermoz, E., Cristofanelli, P., Stocchi, P.,
1499   Agrillo, G., Ma, Y. and Tartari, G.: Weak precipitation, warm winters and springs impact glaciers of south
1500   slopes of Mt. Everest (central Himalaya) in the last 2 decades (1994-2013), Cryosph., 9(3), 1229–1247,
1501   doi:10.5194/tc-9-1229-2015, 2015.

1502   Sasaki, O., Noguchi, O., Zhang, Y., Hirabayashi, Y. and Kanae, S.: A global high-resolution map of debris on
1503   glaciers derived from multi-temporal ASTER images, Cryosph. Discuss., (November), 1–24,
1504   doi:10.5194/tc-2016-222, 2016.

1505   Savéan, M., Delclaux, F., Chevallier, P., Wagnon, P., Gonga-Saholiariliva, N., Sharma, R., Neppel, L. and
1506   Arnaud, Y.: Water budget on the Dudh Koshi River (Nepal): Uncertainties on precipitation, J. Hydrol.,
1507   531, 850–862, doi:10.1016/j.jhydrol.2015.10.040, 2015.

1508   Scherler, D., Bookhagen, B. and Strecker, M. R.: Spatially variable response of Himalayan glaciers to climate
1509   change affected by debris cover, Nat. Geosci., 4(3), 156–159, doi:10.1038/ngeo1068, 2011.

1510   Schmid, M. O., Baral, P., Gruber, S., Shahi, S., Shrestha, T., Stumm, D. and Wester, P.: Assessment of
1511   permafrost distribution maps in the Hindu Kush Himalayan region using rock glaciers mapped in Google
1512   Earth, Cryosph., 9(6), 2089–2099, doi:10.5194/tc-9-2089-2015, 2015.

1513   Sharp, M., Brown, G. H., Tranter, M., Willis, I. C. and Hubbard, B.: Comments on the use of chemically based
1514   mixing models in glacier hydrology, J. Glaciol., 41(138), 241–246, doi:10.1207/s15327752jpa8502, 1995.

1515   Shea, J. M. and Immerzeel, W. W.: An assessment of basin-scale glaciological and hydrological sensitivities
1516   in the Hindu Kush-Himalaya, Ann. Glaciol., 57(71), 308–318, doi:10.3189/2016AoG71A073, 2016.

1517   Shea, J. M., Immerzeel, W. W., Wagnon, P., Vincent, C. and Bajracharya, S.: Modelling glacier change in the
1518   Everest region, Nepal Himalaya, Cryosph., 9(3), 1105–1128, doi:10.5194/tc-9-1105-2015, 2015.

1519   Shreve, R. L.: Movement of water in glaciers, J. Glaciol., 11, 205-214, doi: 10.3189/S002214300002219X,
1520   1972.

1521   Singh, P. and Bengtsson, L.: Hydrological sensitivity of a large Himalayan basin to climate change, Hydrol.
1522   Process., 18(13), 2363–2385, doi:10.1002/hyp.1468, 2004.

1523   Singh, P., Ramasastri, K. S., Singh, U. K., Gergan, J. T. and Dobhal, D. P.: Hydrological characteristics of the
1524   Dokriani Glacier in the Garhwal Himalayas, Hydrol. Sci. J., 40(2), 243–257,
1525   doi:10.1080/02626669509491407, 1995.

1526   Singh, P., Kumar, N., Ramasastri, K. S. and Singh, Y.: Influence of a fine debris layer on the melting of snow
1527   and ice on a Himalayan glacier, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A.
1528   Fountain, pp. 63–69, International Association of Hydrological Sciences, Oxford., 2000.

1529   Singh, P., Haritashya, U. K., Ramasastri, K. S. and Kumar, N.: Diurnal variations in discharge and suspended
1530   sediment concentration, including runoff-delaying characteristics, of the Gangotri Glacier in the Garhwal
1531   Himalayas, Hydrol. Process., 19(7), 1445–1457, doi:10.1002/hyp.5583, 2005.

1532   Singh, P., Arora, M. and Goel, N. K.: Effect of climate change on runoff of a glacierized Himalayan basin,
1533   Hydrol. Process., 20(9), 1979–1992, doi:10.1002/hyp.5991, 2006a.

1534   Singh, P., Haritashya, U. K., Kumar, N. and Singh, Y.: Hydrological characteristics of the Gangotri Glacier,
1535   central Himalayas, India, J. Hydrol., 327(1–2), 55–67, doi:10.1016/j.jhydrol.2005.11.060, 2006b.

1536   Singh, P., Haritashya, U. and Kumar, N.: Modelling and estimation of different components of streamflow
1537   for Gangotri Glacier basin, Himalayas, Hydrol. Sci. J., 53(2), 309–322, doi:10.1623/hysj.53.2.309, 2008.

1538   Smart, C. C.: Artificial Tracer Techniques for the Determination of the Structure of Conduit Aquifers,
1539   Groundwater, 26(4), 445–453, doi:10.1111/j.1745-6584.1988.tb00411.x, 1988.

1540   Smart, C. C.: Statistical evaluation of glacier boreholes as indicators of basal drainage systems, Hydrol.
1541   Process., 10(4), 599–613, doi:10.1002/(sici)1099-1085(199604)10:4<599::aid-hyp394>3.0.co;2-8, 1996.

1542   Soncini, A., Bocchiola, D., Confortola, G., Minora, U., Vuillermoz, E., Salerno, F., Viviano, G., Shrestha, D.,

1543   Senese, A., Smiraglia, C. and Diolaiuti, G.: Future hydrological regimes and glacier cover in the Everest
1544   region: The case study of the upper Dudh Koshi basin, Sci. Total Environ., 1084–1101,
1545   doi:10.1016/j.scitotenv.2016.05.138, 2016.

1546   Sorg, A., Bolch, T., Stoffel, M., Solomina, O. and Beniston, M.: Climate change impacts on glaciers and runoff
1547   in Tien Shan (Central Asia), Nat. Clim. Chang., 2(10), 725–731, doi:10.1038/nclimate1592, 2012.

1548   Souness, C., Hubbard, B., Milliken, R. E. and Quincey, D.: An inventory and population-scale analysis of
1549   martian glacier-like forms, Icarus, 217(1), 243–255, doi:10.1016/j.icarus.2011.10.020, 2012.

1550   Spedding, N.: Hydrological controls on sediment transport pathways: implications for debris-covered
1551   glaciers, in Debris-covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 133–142,
1552   International Association of Hydrological Sciences, Oxford. [online] Available from:
1553   https://books.google.co.uk/books?hl=en&lr=&id=ghn8LV2CZ20C&oi=fnd&pg=PA133&dq=hydrological
1554   +controls+spedding+sediment+transport&ots=RtpBt5MQK1&sig=i0g_gHXs7o-_ZDS3qPvTRzjzk5g,
1555   2000.

1556   Stokes, C. R., Popovnin, V., Aleynikov, A., Gurney, S. D. and Shahgedanova, M.: Recent glacier retreat in the
1557   Caucasus Mountains, Russia, and associated increase in supraglacial debris cover and supra-/proglacial
1558   lake development, Ann. Glaciol., 46(5642 m), 195–203, doi:10.3189/172756407782871468, 2007.

1559   Sundal, A. V., Shepherd, A., Nienow, P., Hanna, E., Palmer, S. and Huybrechts, P.: Evolution of supra-glacial
1560   lakes across the Greenland Ice Sheet, Remote Sens. Environ., 113(10), 2164–2171,
1561   doi:10.1016/j.rse.2009.05.018, 2009.

1562   Takeuchi, N., Sakai, A., Kohshima, S., Fujita, K., Nakawo, M., Shiro, K., Fujita, K. and Masayoshi, N.: Variation
1563   in Suspended Sediment Concentration of Supraglacial Lakes on Debris-covered Area of the Lirung Glacier
1564   in the Nepal Himalayas, Glob. Environ. Res., 16(July), 95–104 [online] Available from: http://www-
1565   es.s.chiba-u.ac.jp/%7B~%7Dtakeuchi/pdf/12ger%7B_%7DLirung%7B_%7Dlake.pdf, 2012.

1566   Takeuchi, Y., Kayastha, R. B. and Nakawo, M.: Characteristics of ablation and heat balance in debris-free
1567   and debris-covered areas on Khumbu Glacier, Nepal Himalayas, in the pre-monsoon season, in Debris-
1568   Covered Glaciers, vol. 264, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 53–61,
1569   International Association of Hydrological Sciences, Oxford., 2000.

1570   Tangborn, W. and Rana, B.: Mass balance and runoff of the partially debris- covered Langtang Glacier,
1571   Nepal, in Debris-Covered Glaciers, edited by M. Nakawo, C. F. Raymond, and A. Fountain, pp. 99–108,
1572   International Association of Hydrological Sciences, Oxford., 2000.

1573   Thakuri, S., Salerno, F., Smiraglia, C., Bolch, T., D'Agata, C., Viviano, G. and Tartari, G.: Tracing glacier changes
1574   since the 1960s on the south slope of Mt. Everest (central Southern Himalaya) using optical satellite
1575   imagery, Cryosph., 8(4), 1297–1315, doi:10.5194/tc-8-1297-2014, 2014.

1576   Thakuri, S., Salerno, F., Bolch, T., Guyennon, N. and Tartari, G.: Factors controlling the accelerated expansion
1577   of Imja Lake, Mount Everest region, Nepal, Ann. Glaciol., 57(71), 245–257,
1578   doi:10.3189/2016AoG71A063, 2016.

1579   Thayyen, R. J. and Gergan, J. T.: Role of glaciers in watershed hydrology : a preliminary study of a "Himalayan
1580   catchment," Cryosph., 4, 115–128, doi:10.5194/tcd-3-443-2009, 2010.

1581   Thayyen, R. J., Gergan, J. T. and Dobhal, D. P.: Monsoonal control on glacier discharge and hydrograph
1582   characteristics, a case study of Dokriani Glacier, Garhwal Himalaya, India, J. Hydrol., 306(1–4), 37–49,
1583   doi:10.1016/j.jhydrol.2004.08.034, 2005.

1584   Thompson, S., Benn, D. I., Mertes, J. and Luckman, A.: Stagnation and mass loss on a Himalayan debris-
1585   covered glacier: Processes, patterns and rates, J. Glaciol., 62(233), 467–485, doi:10.1017/jog.2016.37,
1586   2016.

1587   Thompson, S. S., Benn, D. I., Dennis, K. and Luckman, A.: A rapidly growing moraine-dammed glacial lake on
1588   Ngozumpa Glacier, Nepal, Geomorphology, 145–146, 1–11, doi:10.1016/j.geomorph.2011.08.015,
1589   2012.

1590   Tranter, M., Brown, G., Raiswell, R., Sharp, M. and Gurnell, A.: A conceptual model of solute acquisition by
1591   Alpine glacial meltwaters, J. Glaciol., 39(133), 573–581, doi:10.3198/1993JoG39-133-573-581, 1993.

1592   Tranter, M., Sharp, M. J., Lamb, H. R., Brown, G. H., Hubbard, B. P. and Willis, I. C.: Geochemical weathering
1593   at the bed of Haut glacier d'Arolla, Switzerland - a new model, Hydrol. Process., 16(5), 959–993,
1594   doi:10.1002/hyp.309, 2002.

1595   Vaughan, D. G., Comiso, J. C., Allison, I., Carrasco, J., Kaser, G., Kwok, R., Mote, P., Murray, T., Paul, F., Ren,
1596   J., Rignot, E., Solomina, O., Steffen, K. and Zhang, T.: Observations: Cryosphere, in Climate Change 2013:
1597   The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the
1598   Intergovernmental Panel on Climate Change, edited by T. F. Stocker, G.-K. D. Qin, M. Plattner, S. Tignor,
1599   K. Allen, J. Boschung, A.Nauels, Y. Xia, V. Bex, and P. M. Midgley, pp. 317–382, Cambridge University
1600   Press, Cambridge, United Kingdom and New York, NY, USA., 2013.

1601   Viste, E. and Sorteberg, A.: Snowfall in the Himalayas: an uncertain future from a little-known past,
1602   Cryosph., 9(3), 1147–1167, doi:10.5194/tc-9-1147-2015, 2015.

1603   Vuichard, D. and Zimmermann, M.: The 1985 Catastrophic Drainage of a Moraine-Dammed Lake, Khumbu
1604   Himal, Nepal : Cause and Consequences, Mt. Res. Dev., 7(2), 91–110, 1987.

1605   Walder, J. and Fowler, A.: Channelized subglacial drainage over a deformable bed, J. Glaciol., 40(134), 3–
1606   15, 1994.

1607   Wang, X., Ding, Y., Liu, S., Jiang, L., Wu, K., Jiang, Z. and Guo, W.: Changes of glacial lakes and implications
1608   in Tian Shan, central Asia, based on remote sensing data from 1990 to 2010, Environ. Res. Lett., 8(4),
1609   44052, doi:10.1088/1748-9326/8/4/044052, 2013.

1610   Watanabe, T., Lamsal, D. and Ives, J. D.: Evaluating the growth characteristics of a glacial lake and its degree
1611   of danger of outburst flooding: Imja Glacier, Khumbu Himal, Nepal, Nor. Geogr. Tidsskr. - Nor. J. Geogr.,
1612   63(4), 255–267, doi:10.1080/00291950903368367, 2009.

1613   Watson, C. S., Carrivick, J. and Quincey, D.: An improved method to represent DEM uncertainty in glacial
1614   lake outburst flood propagation using stochastic simulations, J. Hydrol., 529, 1373–1389,
1615   doi:10.1016/j.jhydrol.2015.08.046, 2015.

1616   Watson, C. S., Quincey, D. J., Carrivick, J. L. and Smith, M. W.: The dynamics of supraglacial ponds in the
1617   Everest   region,   central   Himalaya,   Glob.   Planet.   Change,   142,   14–27,
1618   doi:10.1016/j.gloplacha.2016.04.008, 2016.

1619   Watson, C. S., Quincey, D. J., Carrivick, J. L. and Smith, M. W.: Ice cliff dynamics in the Everest region of the
1620   Central Himalaya, Geomorphology, 278, 238–251, doi:10.1016/j.gloplacha.2016.04.008, 2017.

1621   Weertman, J.: On the sliding of glaciers, J. Glaciol., 3, 33–38, doi:10.1007/978-94-015-8705-1_19, 1957.

1622   Weertman, J.: Creep Deformation of Ice, Annu. Rev. Earth Planet. Sci., 11, 215–240, 1983.

1623   Wessels, R. L., Kargel, J. S. and Kieffer, H. H.: ASTER measurement of supraglacial lakes in the Mount Everest
1624   region of the Himalaya, Ann. Glaciol., 34, 399–408, doi:10.3189/172756402781817545, 2002.

1625   Wilhelm, L., Singer, G. A., Fasching, C., Battin, T. J. and Besemer, K.: Microbial biodiversity in glacier-fed
1626   streams, ISME J., 7(8), 1651–1660, doi:10.1038/ismej.2013.44, 2013.

1627   Wilson, A. M., Williams, M. W., Kayastha, R. B. and Racoviteanu, A.: Use of a hydrologic mixing model to
1628   examine the roles of meltwater, precipitation and groundwater in the Langtang River basin, Nepal, Ann.

1629   Glaciol., 57(71), 155–168, doi:10.3189/2016AoG71A067, 2016.

1630   Winiger, M., Gumpert, M. and Yamout, H.: Karakorum-Hindukush-western Himalaya: Assessing high-
1631   altitude water resources, Hydrol. Process., 19(12), 2329–2338, doi:10.1002/hyp.5887, 2005.

1632   Winter, T. C.: The concept of hydrologic landscapes, J. Am. Water Resour. Assoc., 37(2), 335–349,
1633   doi:10.1111/j.1752-1688.2001.tb00973.x, 2001.

1634   Woo, M.: Permafrost Hydrology, Springer, Berlin., 2012.

1635   Woo, M. -K and Xia, Z.: Suprapermafrost groundwater seepage in gravelly terrain, Resolute, NWT, Canada,
1636   Permafr. Periglac. Process., 6(1), 57–72, doi:10.1002/ppp.3430060107, 1995.

1637   Xin, W., Shiyin, L., Haidong, H., Jian, W. and Qiao, L.: Thermal regime of a supraglacial lake on the debris-
1638   covered Koxkar Glacier, southwest Tianshan, China, Environ. Earth Sci., 67(1), 175–183,
1639   doi:10.1007/s12665-011-1490-1, 2012.

1640   Zeng, C., Liu, Z., Yang, J. and Yang, R.: A groundwater conceptual model and karst-related carbon sink for a
1641   glacierized alpine karst aquifer, Southwestern China, J. Hydrol., 529(P1), 120–133,
1642   doi:10.1016/j.jhydrol.2015.07.027, 2015.

1643   Zhang, Y., Liu, S. and Ding, Y.: Glacier meltwater and runoff modelling, Keqicar Baqi glacier, southwestern
1644   Tien Shan, China, J. Glaciol., 53(180), 91–98, doi:10.3189/172756507781833956, 2007.

1645   Zhao, L., Ding, R. and Moore, J. C.: Glacier volume and area change by 2050 in high mountain Asia, Glob.
1646   Planet. Change, 122, 197–207, doi:10.1016/j.gloplacha.2014.08.006, 2014.

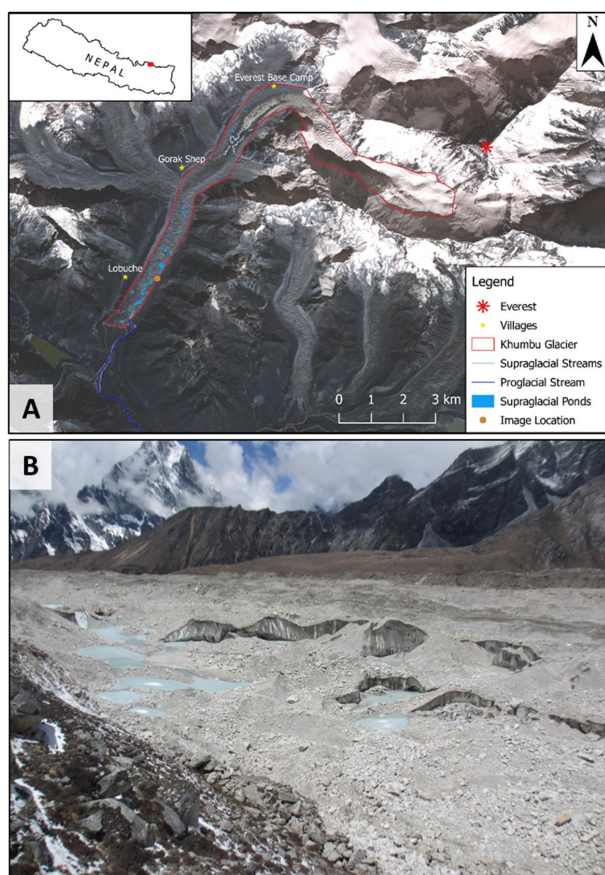1647

12. Figures

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Figure 1 – An example of a typical and particularly well-studied DCG, Khumbu Glacier, Nepal Himalaya. (A) shows a RapidEye image of the glacier acquired on 17.11.2016 (Planet Team, 2017). The major supraglacial hydrological features (larger supraglacial ponds, supraglacial lakes and any supraglacial streams), the proglacial stream and the location from which the image in (B) was acquired are labelled. (B) shows an oblique photograph looking across the glacier surface (image acquisition location shown in (A), taken in the direction of the glacier terminus), also showing some of the supraglacial ponds as well as ice cliffs and variable surface topography. Image credit: KM

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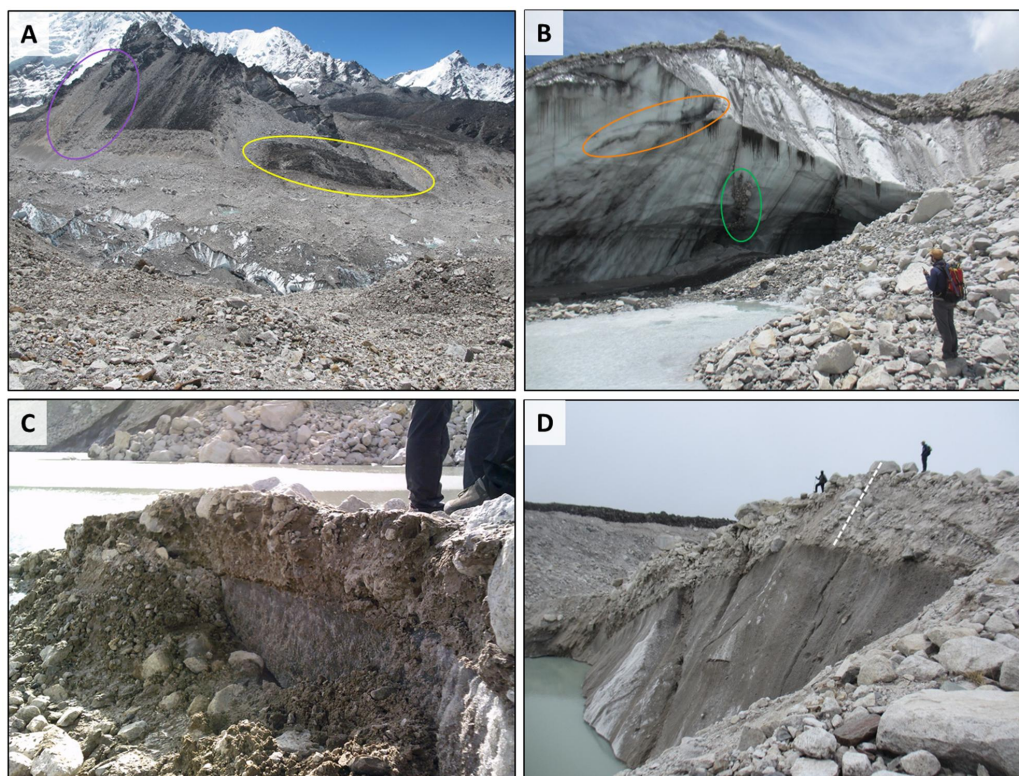
Figure 2 – Images illustrating variations in debris thickness over Khumbu Glacier, Nepal Himalaya: (A) a landslide scar (yellow circle) and unstable rock faces (purple circle) providing debris to the glacier surface; image is taken looking east across the surface of Khumbu Glacier, and the debris layer above ice cliffs can also be seen. (B) shows an ice cliff with entrained debris (green circle), debris-filled crevasse traces (orange circle), and a moderately-thick debris layer above (~1-2 m); (C) a thin debris layer (~20 cm) above ice adjacent to a supraglacial pond; and (D) a thick debris cover (>5 metres, indicated by the white dashed line) above an ice cliff. Image credits: (A) DQ and (B-D) KM

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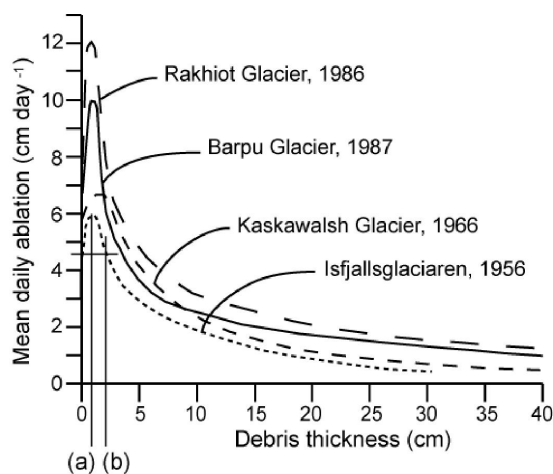
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Figure 3 - Østrem curve examples from Nicholson & Benn (2006, and citations therein), showing the variations in the relationship between debris thickness and ice ablation on different glaciers

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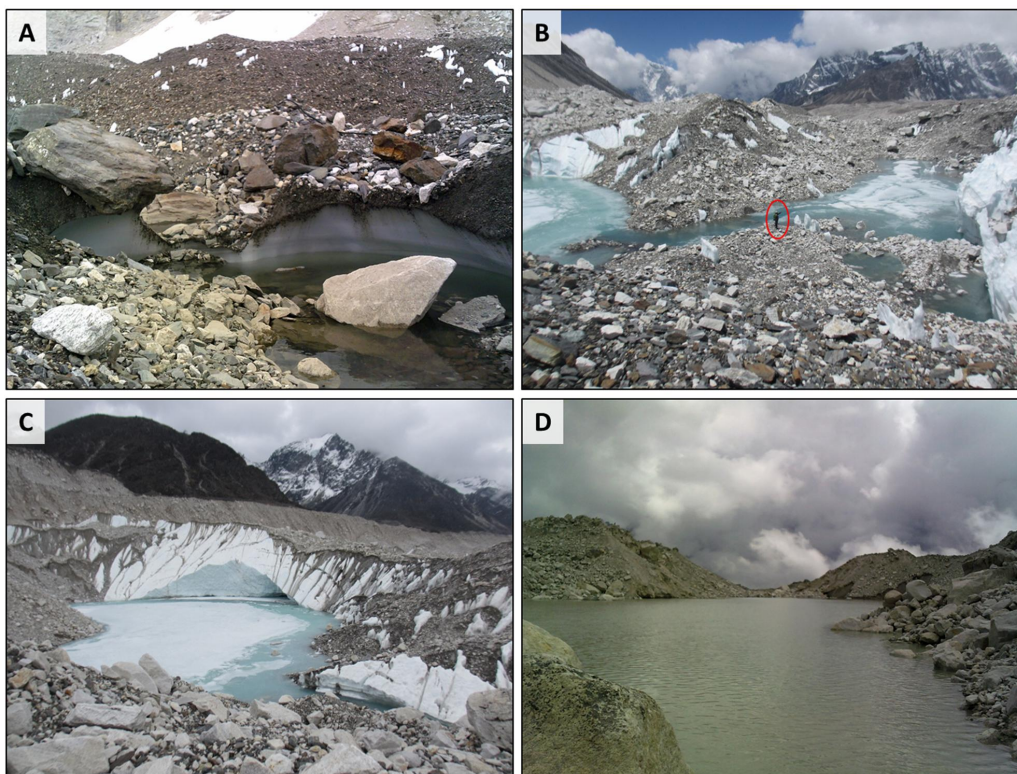
Figure 4 – Examples of supraglacial ponds on Khumbu Glacier, Nepal Himalaya, ranging in diameter from several metres (A), to tens of metres (B, C) and hundreds of metres (D). (A) and (C) also feature a notably large adjacent ice cliff system, relative to each pond/lake size; while (B) has a cliff system on the far of each of the lake sides (a person is circled for scale). Image credits: KM

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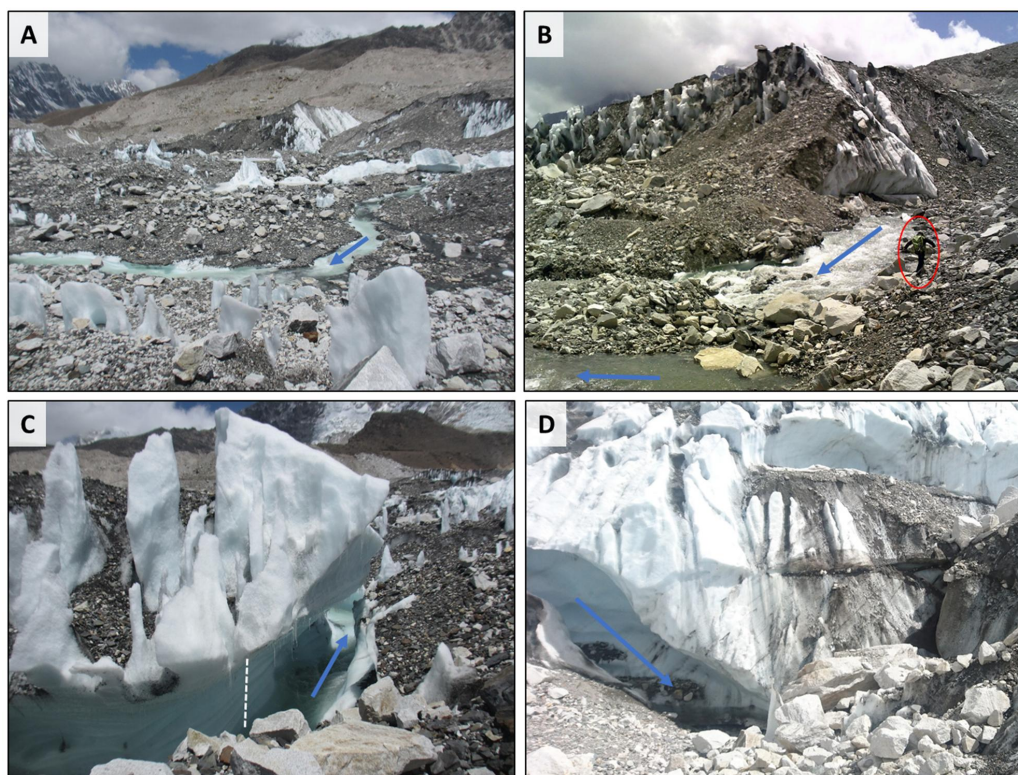
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Figure 5 - Examples of a large supraglacial stream on Khumbu Glacier, Nepal Himalaya; blue arrows indicate water flow direction. (A) shows the stream in the upper ablation area; (B) shows the stream again, approximately 2 km downstream of (A) in the central ablation area and nearly twice the volume, just above a confluence with another large stream (bottom left of image; person shown for scale); (C) is an example of multiple levels of downcutting of the stream (grooves indicated by white dashed line, ~1 m in height), slightly upglacier in location from (A); and (D) shows where this stream eventually disappears below the surface to become englacial, after several hundred metres of progressive downcutting, visible from the multiple relict levels. The drop in the channel is ~10 m, with the stream dropping another few metres beyond the boulders to the right of the image. Image credits: (A-C) KM; (D) EM
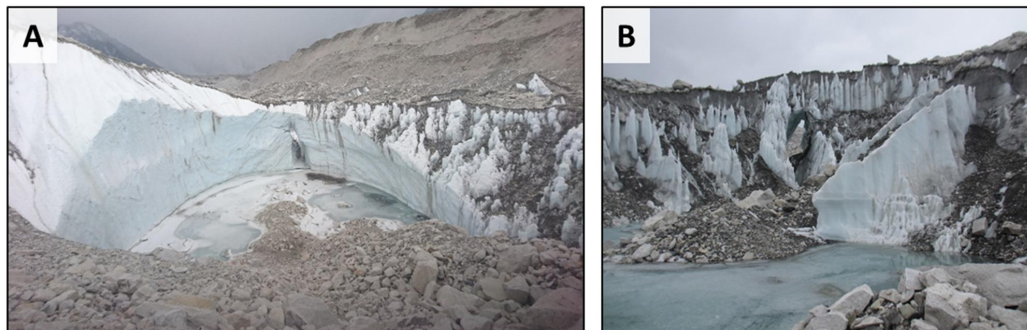
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Figure 6 – A relict englacial feature in the centre of an ice cliff on Khumbu Glacier, Nepal Himalaya, viewed (A) from upglacier, and (B) from downglacier. The associated supraglacial pond is hypothesised to have drained through this feature in the past. Following the drainage event, the pond water-level would have dropped, exposing the ice cliffs around its edge and resulting in the pond water-level being too low to sustain a water flow through the channel. On the downglacier side (B) a vast amount of surface lowering has occurred and the previously englacial channel is now visible from the surface. The relict channel could be seen to continue to meander and downcut for around 200 m further downglacier until joining a pond. The englacial feature is approximately 10 m in height. Image credits: (A) EM; (B) KM
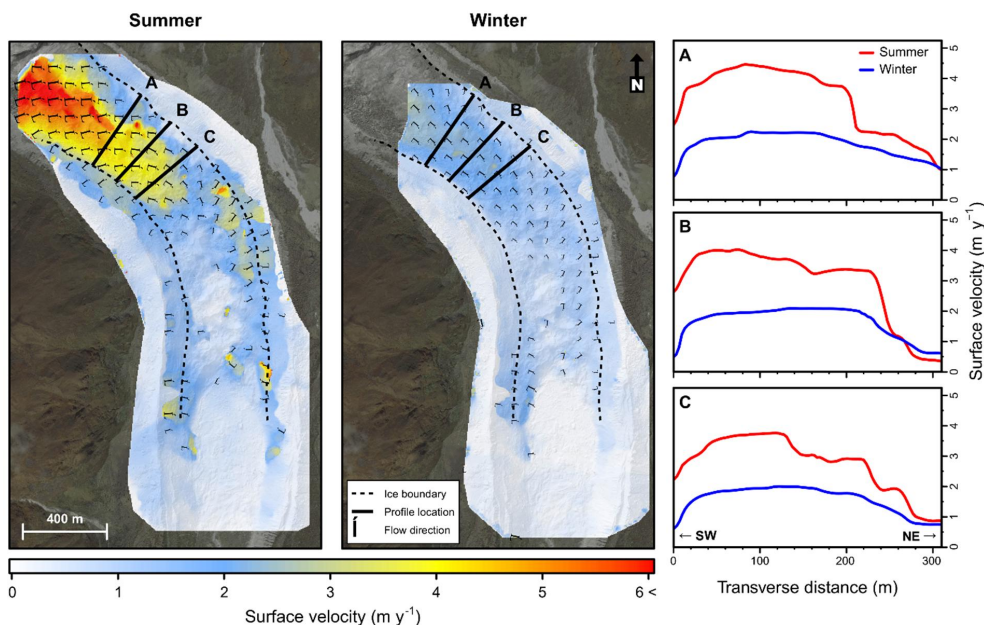
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Figure 7 – Surface velocity maps of Lirung Glacier, Nepal Himalaya, from Kraaijenbrink et al. (2016b) during summer (left) and winter (right), with three transverse velocity profiles (A-C) at the locations marked. Available under a Creative Commons Attribution 4.0 License
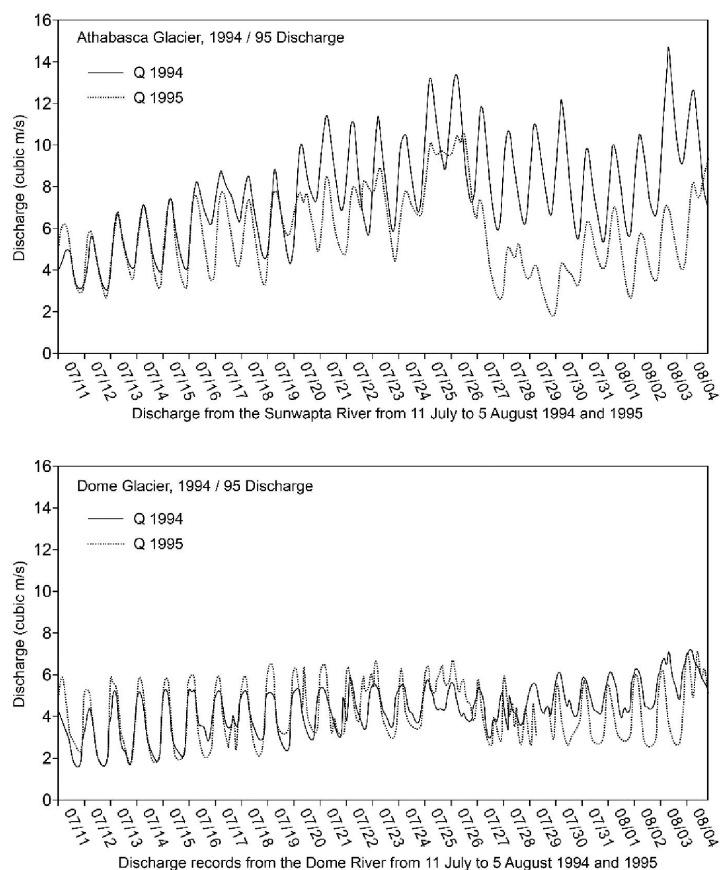
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Figure 8 – Image of Hooker Lake, a proglacial lake in front of the debris-covered Hooker Glacier, New Zealand Southern Alps, taken in 2013. For scale, the ice cliff at the terminus of the glacier is ~30 m in height. Image credit: TIF

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Figure 9 – Hydrographs of proglacial discharge of the clean-ice Athabasca Glacier and the adjacent debris-covered Dome Glacier, Canadian Rockies, over the ablation months of July and August 1994 and 1995. Figure redrawn from Mattson (2000)
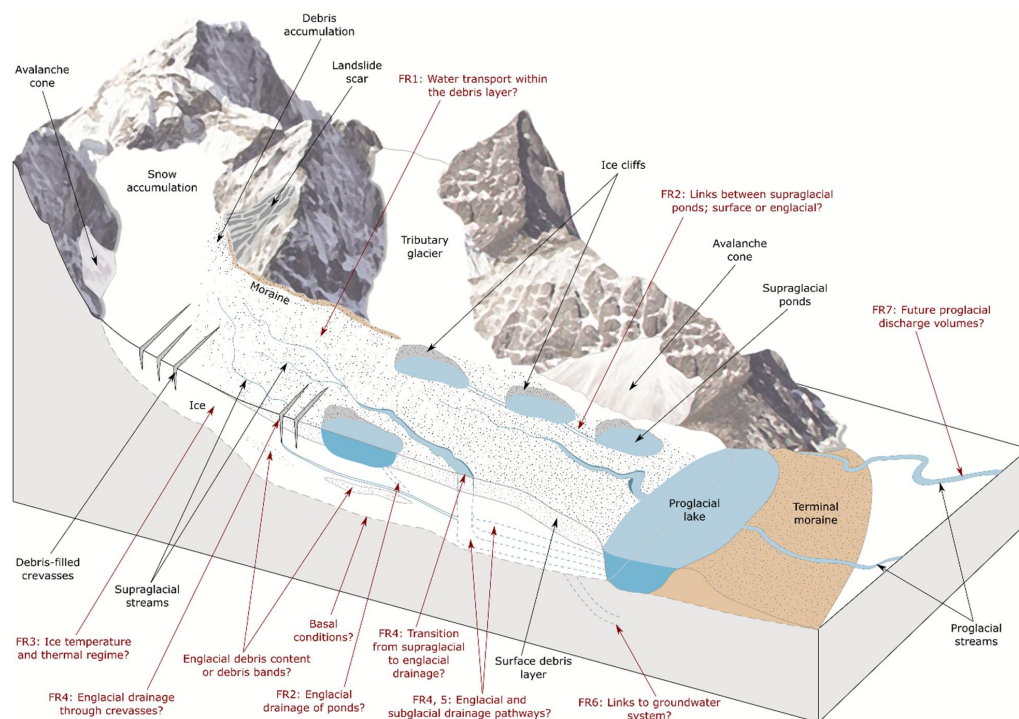
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Figure 10 – A conceptual illustration of the hydrological system of a DCG, including all known (black text), poorly understood and completely unknown potential hydrological features, highlighted in red text and linked to the Future Research (FR) areas for future hydrological research

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