

Do debris-covered glaciers demonstrate distinctive hydrological behaviour compared to clean glaciers?

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11 **Abstract**

12 Supraglacial debris is known to strongly influence the distribution of glacier surface melt. Since
13 melt inputs drive the formation and evolution of glacial drainage systems, it should follow that
14 the drainage systems of debris-covered glaciers will differ from those of debris-free glaciers. This
15 would have implications for the proglacial runoff regime, subglacial erosion and glacier
16 dynamics. This paper presents analysis of return curves from 33 successful dye injections into
17 the extensively debris-covered Miage Glacier, Italian Alps. It demonstrates that the spatial
18 distribution of supraglacial debris influences the structure and seasonal evolution of the glacial
19 drainage system. Where the debris cover is continuous, melt is lower and the surface topography
20 is chaotic, with many small supraglacial catchments. These factors result in an inefficient

21 englacial/subglacial drainage network beneath continuous debris, which drains to the conduit
22 system emanating from the upper ablation zone. Melt rates are high in areas of clean and dirty
23 ice above the continuous debris. Runoff from these areas is concentrated by inter-moraine
24 troughs into large supraglacial streams, which encourages the early-season development of an
25 efficient englacial/subglacial conduit system downstream of this area. Drainage efficiency from
26 the debris-covered area increases over the melt season but dye-trace transit velocity remains
27 lower than from moulins on the upper glacier. Future runoff models should account for the
28 influence of supraglacial debris on the hydrological system.

29 Keywords: debris-covered glaciers; dye tracing; glacier-hydrology

30 **1 Introduction**

31 This paper presents a systematic glacier-scale study of the internal hydrology of a debris-covered
32 glacier, based on data collected over two ablation seasons. Debris-covered glaciers are especially
33 prevalent in mid-latitude, high elevation mountain ranges (Scherler et al., 2018), such as the
34 Pamirs, Karakoram and Himalaya (Scherler et al., 2011; Bolch et al., 2012; Minora et al., 2016),
35 Caucasus Mountains, Russia (Stokes et al., 2007), and the Western Alps (Deline et al., 2012).
36 Glacier-runoff is important for downstream water resources in these regions, especially during
37 dry seasons (Xu et al., 2009; Maurya et al., 2011). Due to negative glacier mass balance, the
38 extent and thickness of supraglacial debris appears to be increasing globally, with implications for
39 future glacier mass balance (Bolch et al., 2008; Bhambri et al., 2011; Lambrecht et al., 2011;
40 Kirkbride and Deline, 2013; Scherler et al., 2018). Debris more than a few cm thick attenuates
41 the diurnal melt signal, due to the time taken for energy to be conducted through the debris to
42 the ice surface (Fyffe et al., 2014). The dominant effect of a continuous debris cover is a
43 reduction in the melt rate compared to bare ice, except where debris is thin or patchy where melt
44 is enhanced (Østrem, 1959; Mattson et al., 1993; Kirkbride and Dugmore, 2003; Mihalcea et al.,

45 2006; Nicholson and Benn, 2006; Brock et al., 2010; Lejeune et al., 2013; Fyffe et al., 2014;
46 Minora et al., 2015).

47 Understanding the nature and evolution of the drainage system of debris-covered glaciers is
48 important because it controls how meltwater inputs influence both glacier dynamics and
49 proglacial runoff regimes (e.g. Mair et al., 2002). Considering the strong influence debris has on
50 surface ablation rates, extensive debris cover can be expected to influence the morphology and
51 evolution of a glacier's hydrological system, but the nature and extent of this impact is not
52 currently known. On debris-covered glaciers melt rates are particularly high just above the limit
53 of continuous debris (Fyffe et al., 2014), contrasting with clean glaciers where melt rates increase
54 towards the terminus. It is this change to the patterns of meltwater generation that may alter the
55 structure and evolution of the hydrological system of debris-covered glaciers. On debris-free
56 temperate glaciers, dye-tracing has demonstrated that the seasonal evolution of the
57 englacial/subglacial hydrological system is characterised by increasing efficiency over time. This
58 increase in efficiency is linked to the increase in volume and daily amplitude of surface meltwater
59 inputs, associated with the upglacier retreat of the seasonal snowline (Nienow et al., 1998; Willis
60 et al., 2002; Campbell et al., 2006). Understanding these processes is as important for debris-
61 covered glaciers, which can display significant velocity variations: both seasonally, with faster
62 summer than winter velocities, as found at Baltoro Glacier, Pakistan Karakoram (Quincey et al.,
63 2009; Mayer et al., 2006), Gangotri Glacier, Indian Himalaya (Scherler et al., 2008), Biafo Glacier,
64 central Karakoram (Scherler and Strecker, 2012) and Miage Glacier, Italian Alps (Fyffe, 2012);
65 and in response to daily weather fluctuations, with short term periods of faster flow also
66 measured on Miage Glacier (Fyffe, 2012). Any impact on drainage form and development would
67 also be expected to alter the relationship between melt production and the proglacial runoff
68 regime relative to a debris-free glacier. Studies of debris-covered glaciers are therefore needed to
69 examine whether the influence of debris on melt rates affects the glacier hydrological system.

70 There have been only a few preliminary dye tracing studies on debris-covered glaciers. Hasnain
71 et al. (2001) studied the autumn close-down of the hydrological system on Dokriani Glacier, and
72 Pottakkal et al. (2014) traced the transfer of proglacial stream water beneath the tongue of
73 Gangotri Glacier. Neither study dealt explicitly with the influence of the debris cover (Table 1).
74 Direct investigation of englacial conduit systems within debris-covered glaciers (see Table 1 for
75 details) has not yet revealed the morphology of inaccessible regions of the glacial drainage
76 network, nor gauged the efficiency of the entire system.

77 This dye-tracing study of Miage Glacier has two objectives: (i) to assess the influence of debris
78 cover on supraglacial topography and hydrology, and therefore on the amplitude, magnitude and
79 spatial distribution of meltwater inputs into the glacial drainage system; and (ii) to determine the
80 morphology and seasonal evolution of the coupled englacial-subglacial hydrological system and
81 its relationship to the spatial distribution of supraglacial debris cover. These objectives provide
82 the structural sub-headings used in the following Methods, Results and Discussions sections.

83 **2 Study Site**

84 Miage Glacier is situated in the Western Italian Alps (Fig. 1). It originates from four main icefall
85 tributaries: the Mont Blanc, Dome, Bionassay and Tête Carrée Glaciers. As the main glacier
86 tongue enters Val Veny it bends eastwards before splitting into the large northern and southern
87 lobes and smaller central lobe. The glacier area is 10.5 km² over an elevation range of 1740 to
88 4640 m a.s.l.. The lower 5 km of the glacier is completely covered by debris which averages 0.25
89 m in thickness (Foster et al., 2012), except for isolated debris-free ice cliffs (Reid and Brock,
90 2014). Debris thickness increases down-glacier with sub-debris ice melt suppressed over most of
91 the lower tongue. At higher elevations (above c. 2500 m a.s.l.) debris is confined to medial and
92 lateral moraines with the intervening ice having a patchy covering of dust to boulder-sized
93 sediment (hereafter 'dirty ice'). The debris originates predominantly from rockfalls and mixed

94 snow and rock avalanches from the steep valley sides (Deline, 2009). Debris cover has been
95 shown to influence the distribution of glacier thinning across the ablation zone (Smiraglia et al.,
96 2000; Thomson et al., 2000; Diolaiuti et al., 2009). Annual horizontal glacier velocity measured
97 between June 2010 and June 2011 was c. 32 m a⁻¹ for most of the main tongue, decreasing
98 downstream to c. 13 m a⁻¹ at 1.5 km above the southern terminus (Fyffe, 2012). Summer
99 velocities exceed winter velocities by $\geq 26\%$ on the main tongue, although the difference
100 decreases to 20% above the divergence of the three lobes (Fyffe, 2012). Ice thickness along the
101 glacier centre line is a maximum of 380 m at around 2350 m a.s.l., decreasing down the main
102 tongue to 250 m, before shallowing past the bend to c. 120 m (Supplementary Material A details
103 the derivation of the ice thickness data). A distributed surface energy-balance melt model (Fyffe
104 et al., 2014) was used to quantify spatial and temporal variation in surface melting over the 2010
105 and 2011 summers.

106 [Fig. 1 here].

107 **3 Methods**

108 **3.1 Background data**

109 Field data were collected over two ablation seasons, from 5 June 2010 to 13 September 2010,
110 and from 4 June 2011 to 16 September 2011. Three meteorological stations were located on the
111 glacier. The lower and upper weather stations (LWS and UWS hereafter) were full energy-
112 balance stations situated on continuous debris cover, with the ice weather station (IWS)
113 measuring only air temperature on an area of dirty ice (Fig. 1). Details of the instruments
114 installed on LWS, UWS and IWS are given in Brock et al. (2010) and Fyffe et al. (2014).

115 The main outflow stream from the glacier exits the northern lobe, while very little drainage exits
116 the southern lobe. Discharge was monitored at a gauging station directly downstream of the
117 northern portal (Fig. 1). Stage was measured using a pressure transducer mounted in a well

118 attached to a large, stable boulder (see Table 2 for details). The Onset HOB0 pressure data were
119 compensated using air pressure data from Mont de la Saxe, 7.6 km from the gauging station. A
120 high flow event in June 2011 caused damage to the well, resulting in data loss between 18 June
121 2011 and 3 August 2011, and the repositioning of the well. Other data voids are 27 to 28 August
122 2010 and 4 to 8 September 2010. All recorded stages were adjusted to the datum of the June
123 2010 stilling well so that a single stage-discharge rating could be applied to the entire record. The
124 stage-discharge rating was derived from discharges calculated from dye dilution gauging using
125 rhodamine WT. In total 16 dye dilution gaugings performed in both 2010 and 2011 provided a
126 two-part rating curve which has a standard error of the estimate of $0.76 \text{ m}^3 \text{ s}^{-1}$, which gave a
127 percentage error of 14.6% using the average daily discharge in 2010 of $5.37 \text{ m}^3 \text{ s}^{-1}$. The use of a
128 single rating curve for the whole period was justified by the correspondence of gaugings from
129 different field visits.

130 **3.2 The influence of debris on the supraglacial topography and hydrology**

131 *3.2.1 Delimiting supraglacial catchments and routing*

132 Supraglacial streams and their catchments were defined by applying Arnold's (2010) lake and
133 catchment identification algorithm (LCIA) to a digital elevation model (DEM). This supraglacial
134 algorithm is favoured because it does not rely on the artificial filling of sinks before calculating
135 the flow routing. Arnold (2010) provides detailed model methods. The DEM was derived from
136 airborne LiDAR surveys in 2008 (provided by Regione Autonoma Valle d'Aosta, VDA DEM
137 hereafter) and has a spatial resolution of 2 m and a vertical accuracy of $< 0.5 \text{ m}$. The VDA DEM
138 was resampled to a 4 m cell size and was clipped to the glacier area. Supraglacial catchments were
139 categorised by surface cover type (debris-covered ice, clean ice or dirty ice, as shown in Fig. 1) by
140 determining the surface cover within which their centroid was located.

141 3.2.2 *Supraglacial stream measurements*

142 Prior to conducting a dye trace, the discharge and velocity of the chosen supraglacial stream (Q_s
143 and u_s , respectively) were measured in 2011 only. Either the velocity-area method or salt dilution
144 gauging was used to measure supraglacial stream discharge. Dilution gauging was preferred but
145 was not always possible. The cross-sectional area was calculated by multiplying stream width by
146 depth, measured on average at 9 points across the channel. Surface velocity was measured by
147 timing the passage of floats, which probably overestimate mean depth-averaged velocity
148 (Dingman, 2002). Floats usually followed the stream thalweg and so travelled faster than the
149 width and depth-averaged flow. Salt dilution gauging was performed using a portable
150 conductivity probe (Table 2), where the dilution gauging velocity was the distance between
151 injection and detection points divided by the time between injection and peak of the
152 concentration curve. This gives a better measure of velocity than is provided by the float
153 method. Therefore, discharges measured using the velocity-area method were adjusted using the
154 ratio of dilution velocity to float velocity found from simultaneous measurements.

155 **3.3 The influence of debris on the englacial and subglacial hydrological system**

156 In total 48 dye injections were conducted into 16 surface streams, with 33 breakthrough curves
157 successfully detected. All dye injections were carried out using 21% rhodamine WT liquid dye.
158 Between 40 and 280 ml of dye was used per injection. To allow comparison of breakthrough
159 curves from the same streams, repeat injections were conducted at similar times of day,
160 particularly for upglacier streams. The injection times for streams traced on multiple occasions
161 (resulting in successful traces) are 14:27-16:50 (S3), 13:00-17:10 (S5/S5b), 16:31-19:02 (S7),
162 15:15-16:22 (S12/S12b), 12:08-15:12 (S14/S14b), and 13:18-15:29 (S15). Dye traces were
163 detected at the gauging station using a fluorometer (see Table 2) and a Campbell data-logger
164 (CR500, until 14 June 2011 when it was replaced with a CR10X) at either 5 or 1 minute intervals.
165 Each fluorometer was calibrated in the field with each batch of dye, with the dye concentration

166 calculated after subtraction of the background fluorescence (which was either a constant value or
167 occasionally a gradient over time). However, high frequency background variation remained.
168 Genuine dye breakthrough curves were therefore distinguished from background variability on
169 the basis of a) the dye concentration being greater than 2 x the standard deviation of the
170 background concentration and b) a period of continuous above background dye concentration
171 surrounding the peak value lasting at least 10 minutes. The latter condition is necessary to
172 distinguish from short term noise which can, on occasion, exceed the background concentration
173 threshold for 1-2 minutes. The standard deviation of the background variability was calculated
174 using data from the first 30 minutes after dye injection. This time period was chosen to be
175 sufficient to capture the background variation (which had a period in the order of several
176 minutes) while not including data influenced by dye emergence. The minimum period of
177 continuous above background dye concentration for interpreted traces was 15 minutes (mean 2
178 hours 59 minutes). This supports the initial qualitative assessment that all interpreted
179 breakthrough curves are due to dye.

180 Dye was always injected into flowing streams. On the lower glacier, streams could be obscured
181 by debris (hiding moulins from view) or contain clasts, but the streams used for injections did
182 not flow through the debris matrix itself. On the lower glacier, moulins were not located for the
183 S1, S2 and S6 streams; the S3 and S4 streams did apparently sink into a moulin a few metres
184 from the injection point but this was hidden by large boulders; S7 became englacial a short
185 distance from the injection site through the 'cut and closure' mechanism rather than via a
186 moulin; and the S8 injection was directly into an englacial conduit. The injection point into S5
187 was into a stream 446 m upstream of the moulin and so the trace transit velocity (u) was adjusted
188 to account for the time spent in the supraglacial stream, using the measured supraglacial stream
189 velocity (u_s) at the time of the test (2011 only). Henceforth, only adjusted u is given. On the

190 upper glacier all successfully traced streams (S10, S12, S13, S14 and S15), flowed directly into a
191 moulin, except S11 which likely flowed into the S10 moulin.

192 During 2011, repeat injections at individual points were prioritized. Five injection points were
193 chosen, two on the lower glacier (S5 and S7), and three on the upper glacier (S12, S14 and S15)
194 (see Fig. 1). The three upper points were intended to be spread equally along the glacier, but all
195 the moulins present were clustered in a relatively small area. The parameters calculated for each
196 dye-breakthrough curve are given in Table 3.

197 **4 Results**

198 **4.1 The influence of debris on the supraglacial topography and hydrology**

199 The tributary ice falls have small surface catchments due to the crevassing creating a chaotic
200 surface (Fig. 2 and Table 4). Drainage capture will keep inputs small and widely-distributed,
201 although large subglacial streams were exposed at the base of the tributary glaciers, indicating
202 that subglacial drainage does become channelized at this elevation.

203 In the mid-glacier area the catchments are relatively large, since areas of dirty ice are laterally
204 enclosed by debris-covered moraine crests (Figs. 2 and 3a). Table 4 shows that the dirty ice
205 catchments (which include the main medial and lateral moraines) correspond to a larger mean
206 catchment area. Streams injected in this area include S12 (the main stream draining the eastern
207 side of the upper glacier, Supplementary Fig. 1a) and S14 (the main stream draining the western
208 side of the upper glacier, Supplementary Fig. 1b). These streams had the highest Q_s and u_s of
209 those measured (Table 6), with the Q_s range 0.378-0.888 $\text{m}^3 \text{s}^{-1}$ for S14 and 0.025-0.341 $\text{m}^3 \text{s}^{-1}$ for
210 S12 and the u_s range 0.92-2.16 m s^{-1} for S14 and 0.43-0.50 m s^{-1} for S12. In contrast, the
211 crevassed, debris-covered lateral moraines had smaller supraglacial catchments due to the en-
212 echelon crevasses intersecting surface runoff.

213 On the heavily debris-covered lower tongue the debris cover resulted in hummocky topography
214 and consequently consistently small supraglacial catchments (the maximum catchment area and
215 mean area of the largest 10 catchments are much lower than for clean and dirty ice) (Figs. 2, 3b
216 and Table 4). Supraglacial streams were difficult to find in this region of the glacier and there was
217 a lack of well-defined moulins. Streams injected in the continuously debris-covered zone
218 included S5 (the largest stream observed on the lower glacier) and S7, both of which had
219 relatively low Q_s and u_s . The Q_s range was 0.027-0.032 m³ s⁻¹ for S5 and 0.006-0.032 m³ s⁻¹ for S7
220 and the u_s range was 0.13-0.25 m s⁻¹ for S5 and 0.17-0.28 m s⁻¹ for S7. Crevasses are scarce here,
221 confirming the importance of the debris cover in determining the supraglacial catchment
222 boundaries (Fig. 2).

223 [Fig. 2 and 3 here].

224 **4.2 The influence of debris on the englacial and subglacial hydrological system**

225 For context, meteorological and proglacial runoff fluctuations are given in Fig. 4. Dye trace
226 parameters for all 2010 and 2011 injections are reported in Tables 5 and 6, with dye return
227 curves shown in Figs. 5, 6, 7 and 8. Injections into S10 and above are termed upper glacier
228 injections (draining patchy debris and bare ice), while those into S8 and below are termed lower
229 glacier injections (continuously debris-covered ice). No successful traces were obtained from S9.

230 *4.2.1 Spatial Patterns*

231 Generally, water entering the glacier via the main moulins around the upper limit of continuous
232 debris cover travelled quickly to the proglacial stream, with mean u of the S10 to S15 dye traces
233 being 0.56 m s⁻¹. These traces mostly had single-peaked return curves (Figs. 6a and 7bdf) and
234 relatively high percentage dye returns (P_r), confirming that the majority of the water was routed
235 efficiently. Most streams from the lower glacier had low u (the average for all lower glacier
236 injection points was 0.26 m s⁻¹), with the exception of S6 and S8 (Fig. 6a) which had a higher u of

237 0.58 m s⁻¹ and 0.43 m s⁻¹, respectively. Return curves from lower glacier injections were generally
238 broader and several displayed multiple peaks (Figs. 5, 6b and 7ace).

239 A striking result is that average u increases with distance upglacier and is significantly positively
240 correlated with the distance from the gauging station (in June of both years and September 2011,
241 with all time periods giving positive correlations, see Fig. 9a). P_r was also significantly positively
242 correlated with distance from the gauging station in June of both years (excluding P_r values
243 greater than 100%, see Fig. 9b).

244 [Figs. 4, 5, 6, 7, 8 and 9 here or at least in section 4].

245 4.2.2 *Seasonal evolution*

246 4.2.2.1 Lower glacier

247 Lower glacier traces in early June generally showed low u (e.g. injections into S1, S3, S5 and S7
248 had $u < 0.2$ m s⁻¹) and often displayed multiple peaks (e.g. S5_06Jun11 and S7_05Jun11, Fig. 5).

249 An increase in the efficiency of the drainage network during June was evidenced by a decrease in
250 the number of peaks in the S5 (from seven to three more prominent peaks) and S7 (from multi-
251 to single-peaked) dye breakthrough curves (Fig. 5). Further evolution of the drainage network
252 between June and July was shown at S5 by an increase in u and P_r , and a decrease in the
253 dispersion coefficient (D) and dispersivity (b), alongside the increased prominence of the first
254 peak in the breakthrough curve (Fig. 7c). The S3_29Jul10 injection produced a single peak,
255 much clearer than its June counterpart, with a higher u and much larger P_r (Fig. 7a). A slight
256 reduction in drainage efficiency between July and September was shown at S3 in 2010 by a
257 reduced u and D (Fig. 7a and Table 5) and at S5 in 2011 by a reduced u and increase in the
258 number of peaks in the breakthrough curve (Fig.7c).

259 4.2.2.2 Upper glacier

260 Most upper glacier injections in June (into S10, S12, S13, S14) had $u > 0.4 \text{ m s}^{-1}$, with low D and
261 b , despite the early season stage and extensive snow cover on the upper glacier. Traces tended to
262 have discrete, narrow peaks, although the secondary peak on the S13_11Jun10 trace suggests
263 temporary water storage in the moulin or a secondary channel (Fig. 8). The shoulder of the
264 S15_13Jun11 trace (Fig. 8) indicates that water was being released gradually, most likely past an
265 englacial channel constriction.

266 Comparing June and July traces, the S15_28Jul11 u was much higher than in June (Fig. 7f) and
267 no longer had a flat top to the trace, causing a reduction in D and b . Conversely, the late July
268 return curves S12_30Jul11 and S14_29Jul11 were slower and more dispersed than in June,
269 although they still had single peaks (Figs. 7b and 7d, respectively), indicating the efficiency of the
270 channel system had decreased. S12, S14 and S15 were all injected again 3 or 4 days later at the
271 start of August. All three traces showed a strong increase in u (Fig. 10a), a decrease in D and b ,
272 and an increase in A_m (Fig. 10b), compared to the return curves prior to 31 July 2011 – indicating
273 an increase in channel efficiency between late July and early August. The September traces from
274 S12, S14 and S15 had higher u than the June and end of July traces, but similar to, or in the case
275 of S12, slightly lower, than their early August traces (Fig. 7bdf and Table 6).

276 [Fig. 10 here].

277 5 Discussion

278 5.1 The influence of debris on the supraglacial topography and hydrology

279 In the region of the glacier between c. 2300 and 2500 m a.s.l., surface topography is strongly
280 controlled by contrasting ablation rates between thick moraine-crest debris ($\sim 0.02 \text{ m w.e. d}^{-1}$;
281 Fyffe et al., 2014), and more sparsely debris-covered ice in the intervening troughs ($\sim 0.05 \text{ m w.e.}$
282 d^{-1} ; Fyffe et al., 2014), generating predominantly large valley-shaped catchments. Medial moraines

283 grow downglacier to 30-40 m vertical amplitude (Fig. 3a). Thus, relatively high meltwater
284 discharges are focused into the inter-moraine troughs, resulting in a small number of high
285 discharge streams, feeding the cluster of moulins at S12-S15. This explains the relatively large Q_s
286 and u_s measured at S12 and S14 (Sect. 4.1).

287 Surface relief decreases downglacier below 2300 m a.s.l. due to the gravitational redistribution of
288 debris down moraine flanks into the troughs. This inverts relief development by reversing the
289 ablation gradient down the moraine flanks, resulting in the hummocky and chaotic topography
290 of the lower tongue (Fig. 3b). Consequently, there is less potential for the formation of an
291 integrated channel network on the continuously debris-covered zone, resulting in a local stream
292 network with a large number of small catchments and hollows which may form ponds (Fig. 2
293 and Table 4). Melt rates beneath continuous debris on the lower glacier averaged $0.019 \text{ m w.e. d}^{-1}$
294 in 2010, (Fyffe et al., 2014), hence, much less meltwater per unit area is produced on the lower
295 glacier, despite the low elevation and relatively high air temperatures. This explains the small Q_s
296 and low u_s of the streams on the lower tongue.

297 The diurnal amplitude of Q_s on the debris-covered part of the glacier is also likely reduced
298 because the melt signal is attenuated beneath thick debris due to the time taken for the energy
299 receipt at the surface to be conducted through the debris to the ice/debris interface (see Figs. 10
300 and 11 of Fyffe et al. (2014)). Furthermore, meltwater may be further delayed while flowing
301 through the debris matrix prior to reaching a supraglacial stream, although this delay has not yet
302 been quantified. The cumulative effect of the surface debris cover is to reduce and attenuate the
303 inputs into the rest of the hydrological system.

304 **5.2 The influence of debris on the englacial and subglacial hydrological system**

305 *5.2.1 Early formation and evolution of a channelized system draining the upper glacier*

306 Fast, peaked and low dispersion dye breakthrough curves from the upper glacier indicate that a
307 channelized system connects surface streams originating on clean and dirty ice, above the
308 continuously debris-covered zone, to the proglacial stream. This was the case even in early June
309 2010 when the glacier was snow-covered well below the elevation of the upper moulins.

310 The seasonal evolution of a temperate glacier's hydrological system is caused by an increase in
311 the magnitude and amplitude of inputs into the system, usually initiated by the switch from snow
312 to ice melt, which causes pressure fluctuations large enough to destabilise the hydraulically
313 inefficient distributed system into a more efficient discrete channel system (e.g. Nienow et al.,
314 1998; Willis et al., 2002; Campbell et al., 2006). The establishment of a channelized network
315 draining the upper glacier moulins prior to the depletion of the winter snow cover cannot be due
316 to the year-round survival of subglacial conduits, because this is precluded by our closure-rate
317 calculations (Supplementary Material A).

318 It is therefore argued that early season snowmelt inputs were able to initiate channelization. This
319 could be a consequence of the large catchment areas upstream of the moulins in the central
320 ablation area (Sect. 5.1). The S12 and S14 catchments are at a relatively low elevation (2400-2500
321 m a.s.l.) for the region, lower than the terminus elevation of most debris-free glaciers in the
322 western European Alps. Melt-induced evolution of a persistent snow pack may allow the
323 development of an efficient supraglacial drainage system beneath the early-summer snowpack to
324 give input hydrographs of sufficient amplitude to channelize the system (Mair et al., 2002). The
325 evidence from Miage Glacier therefore adds to Mair et al.'s (2002) argument that the retreat of
326 the snowline is not always necessary for channelization. Both flow concentration between
327 longitudinal moraines and efficient snow pack drainage are not exclusive to, but may be more
328 prevalent on, debris-covered glaciers.

329 The drainage network emanating from upglacier remained channelized from June to September,
330 indicated by relatively fast and single-peaked return curves in both years. However, between June
331 and July 2011 there was a relative reduction in channel efficiency (section 4.2.2.2), likely due to
332 cold weather in July resulting in partial closure of the main subglacial conduit. When the weather
333 warmed after 28 July 2011 the conduit was unable to evacuate the increased discharges
334 efficiently, resulting in hydraulic damming caused by the conduit geometry being small relative to
335 the flow through the conduit (Supplementary Material B discusses this interpretation). By the
336 end of July and beginning of August, the dye breakthrough curves suggested an increase in
337 channel efficiency, likely a result of rapid conduit growth in response to the increased input
338 discharges.

339 *5.2.2 Englacial and subglacial drainage beneath continuous debris: co-existence of inefficient distributed and*
340 *efficient channelized drainage*

341 Several characteristics of dye return curves at Miage Glacier indicate that the hydrological system
342 draining the continuously debris-covered zone was far less efficient than that draining the upper
343 debris-free area. These include low flow velocities, multiple peaks, high dispersion, and low
344 percentage dye returns. Since dye breakthrough curves integrate the effects of the whole drainage
345 path, the source of this inefficiency could be the supraglacial, englacial and/or subglacial part of
346 the network.

347 Supraglacial streams on the lower glacier all had low discharges and velocities so any supraglacial
348 dye transport would be slower and more dispersed (Sect. 4.1). Similarly, flow within the englacial
349 network would likely be slow if inputs are relatively small. The englacial path may also be longer
350 if the stream becomes englacial by the cut-and-closure mechanism rather than a vertical moulin
351 (e.g. S7). A less efficient englacial network in the debris-covered zone is therefore a result of the
352 debris reducing the supraglacial stream size and velocity, as explained in Sect. 5.1. Single peaked

353 return curves with a similar n to supraglacial stream velocity could be the result of inefficient
354 (slow) transport through a channelized englacial and subglacial system.

355 However, the multi-peaked nature of the June dye breakthrough curves from S5 and S7 indicates
356 the existence of a distributed subglacial system emanating from these streams. This interpretation
357 is based on the lack of other convincing mechanisms that could result in multi-peaked traces.

358 Englacial networks may alter the trace n and D but observations of their structure (Gulley and
359 Benn, 2007; Gulley et al. 2009a, 2009b; Benn et al., 2009, 2017) suggests flow remains
360 channelized, meaning multi-peaked traces would be unlikely. Variations in supraglacial and main
361 channel discharge (Nienow et al., 1996; Schuler et al., 2004; Werder et al., 2010) or an increase in
362 roughness (Gulley et al., 2012) have also resulted in breakthrough curves with low n and large D
363 and b values. However, curves from these studies still exhibited one main peak, although some
364 displayed a shoulder or small secondary peak. Where multi-peaked breakthrough curves have
365 been detected they have been interpreted as resulting from flow in a distributed system: e.g. at
366 Midtdalsbreen (resulting from flow in a linked cavity system (Willis et al., 1990)); Storglaciaren
367 (due to flow in an anabranching braided system (Seaberg et al., 1988); the debris-covered
368 Dokriani Glacier (suggesting a distributed system (Hasnain et al., 2001)) or from boreholes (e.g.
369 Hooke and Pohjola (1994) and Iken and Bindschadler (1986)). The S5 and S7 streams therefore
370 likely drained into an inefficient distributed system early in the melt season.

371 The role of debris in reducing meltwater inputs (Sect. 5.1) below the critical discharge at which
372 channels develop, appears crucial in inhibiting channelization (Hewitt and Fowler, 2008). These
373 low magnitude and low amplitude inputs are not able to create water pressures great enough to
374 reach the ice overburden pressure, so there is no initiation of the unstable cavity growth needed
375 to create an efficient channelized system downstream of where englacial water reaches the bed
376 (Walder, 1986). The sediment at the bed of the lower glacier (Pavan et al. 1999, cited in Deline,

377 2002) may also inhibit channelization, since the transfer of water within a ‘soft’ bed acts to
378 reduce subglacial water pressures and therefore prevent conduit formation (Flowers, 2008).

379 These results imply the coexistence of an inefficient distributed drainage system with an efficient
380 channelized system beneath the continuously debris-covered zone. Distributed and channelized
381 systems are known to co-exist on other glaciers (e.g. beneath Haut Glacier d’Arolla (Nienow et
382 al., 1996), Midtalsbreen (Willis et al., 1990) and South Cascade Glacier (Fountain, 1993)).

383 Beneath Miage Glacier channelized drainage from the upper glacier travels beneath the lower
384 glacier before exiting at the northern lobe proglacial stream. Meanwhile, sub-debris melt from
385 the lower glacier is routed from supraglacial streams into the inefficient englacial/subglacial
386 system (which subglacially may consist of a distributed or inefficient channelized system) before
387 joining the main efficient subglacial channel system with the rest of the water from upglacier.

388 Those lower glacier streams with a relatively fast n (S6 and S8) may have a more direct link to the
389 main subglacial channel system. The overall drainage system structure implied from dye tracing is
390 shown schematically in Fig. 11.

391 Given the apparent association between surface topography and the geometry of the englacial
392 and subglacial systems, it is possible that the topographic evolution of debris-covered glaciers
393 during deglaciation could influence their hydrological system. As the topography changes from
394 ‘youth’ (debris cover restricted to moraines) to ‘maturity’ (continuous thick debris, uneven
395 topography) (Clayton, 1963), supraglacial drainage would become constrained to small
396 catchments draining to supraglacial lakes (e.g. on the Ngozumpa (Benn et al., 2001, 2017) and
397 Tasman Glaciers (Kirkbride, 1995; Röhl, 2008)). Subglacial drainage would be mainly inefficient,
398 except for conduit systems emanating from upglacier moulins or from sporadic lake drainage.
399 Clearly, further investigations of debris-covered glaciers with a more mature topographic
400 development are required.

401 [Fig. 11 here].

402 **6 Conclusion**

403 Through an extensive dye tracing investigation of the hydrological system of a debris-covered
404 glacier, this paper demonstrates that the structure and seasonal evolution of the hydrological
405 system of debris-covered glaciers is distinct from that of debris-free glaciers. This is significant
406 because it influences the timing and magnitude of proglacial runoff, with the slower transport of
407 sub-debris melt through an inefficient system resulting in a more attenuated proglacial
408 hydrograph. It also influences glacier dynamics, with faster and more variable glacier velocities
409 observed in the mid-glacier corresponding with the locations of the largest moulines. On the
410 lower tongue glacier velocities are slower even though the subglacial drainage is inefficient, likely
411 because of the smaller meltwater inputs from the continuously debris-covered area (Fyffe, 2012).
412 There are also implications for future runoff models which should consider the influence of
413 supraglacial debris on the hydrological system, since it cannot be assumed that the runoff models
414 determined for clean glaciers will hold for debris-covered glaciers.

415 The key findings are that:

- 416 1. On Miage Glacier: i) the formation of an efficient channelized network develops
417 downglacier of areas of clean ice/snow and discontinuous debris on the upper ablation
418 zone, but drainage is inefficient beneath the continuously debris-covered lower ablation
419 zone, and ii) transit velocity through the hydrological system increases linearly with
420 distance upglacier, in contrast to debris-free glaciers.
- 421 2. Runoff from the upper ablation zone is connected to the main proglacial stream via an
422 efficient channelized system, which becomes established early in the melt season when
423 snow-cover is still extensive and is maintained throughout the ablation season. The
424 establishment and maintenance of this efficient system is promoted both by very high
425 ablation rates on dirty-ice areas, and by the topographic concentration of flow into large

426 channels within the inter-moraine troughs. These troughs are themselves a result of
427 differential ablation between the debris-covered moraines and mainly debris-free valleys.

428 3. The majority of meltwater from the lower, continuously debris-covered area is drained
429 via an inefficient englacial/subglacial network which may feed gradually into the main
430 channelized network, although on occasion there is a more direct link with the main
431 conduit system. Hence, both efficient and inefficient drainage systems co-exist beneath
432 the continuous debris zone. The inefficient network is a consequence of the dispersed,
433 low magnitude melt inputs which result in slower transport through the system and may
434 prevent water pressure fluctuations becoming great enough to destabilise a distributed
435 subglacial system. The small discharges in this area are themselves a consequence of both
436 low and attenuated melt peaks beneath thick debris, possible delays as water is
437 transferred through the debris layer and the hummocky topography which restricts
438 catchment and stream size.

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643

644 Supplementary Material

645 **A Conduit Closure**

646 Conduit closure rates were calculated by integrating equation 7 in Hooke (1984, cited in Nienow
647 et al., 1998). The time, t (s) for a conduit to close to a given radius, r_r (m) is given by:

648

$$649 \quad t = \frac{\ln(r_r) - \ln(r_i)}{\left(\frac{\rho_i g h}{n A_G}\right)^3}, \quad (3)$$

650

651 where ρ_i is the ice density (kg m^{-3}), g is gravitational acceleration (9.81 ms^{-2}), $n=3$ and $A_G = 5.8$
652 $\times 10^7 \text{ Pa s}^{-1/3}$, both constants in Glen's flow law (Nienow et al., 1998). The ice thickness (h , m)
653 was calculated as the difference between the surface and bed elevation. The VDA DEM was
654 used to give the surface topography. A map of the bed topography in Deline (2002) (based on
655 seismic reflection surveys by Carabelli (1961) and Casati (1998), and an indirect method using
656 surface velocities by Lesca (1974)), was scanned, georeferenced, digitised and interpolated into a
657 raster with a 25 m cell size. Unfortunately, the resolution of the map contours was low and the
658 fit of the map to the glacier outline was poor due to a lack of clear control points. The resulting
659 conduit closure rates should therefore be treated with caution. The value of h used in equation 3
660 was derived by extracting a profile of ice thickness measurements (at approximately 25 m
661 intervals) from the proglacial stream portal, up the northern lobe and along the glacier centre
662 line. It was assumed that a single conduit links the upper moulines and proglacial stream, with the
663 initial conduit radius (r_i , m) derived by linearly interpolating the measured input (see below) and
664 proglacial stream discharge and dividing this by n to give the channel cross sectional area along
665 the entire stream length. The assumption of a single subglacial conduit allows the closure
666 calculations to be applied to the likely maximum conduit cross-sectional area and should not be

667 taken to imply that this is the most likely drainage structure. The conduit was assumed to be
668 semi-circular and to have effectively closed when it had a radius of 0.01 m.

669 To understand the sensitivity of the calculations to r_p , the time taken for the conduit to close was
670 calculated using either the S12, S14 or the sum of the S12 and S14 September 2011 supraglacial
671 discharges. The proglacial discharge was taken as the mean of the proglacial discharge at the
672 injection and peak of the return curve for the respective injection, or the mean for the combined
673 S12 and S14 test. The ice density was also varied from 830 kg m^{-3} (lowest density of glacial ice
674 (Paterson, 1994)) to 920 kg m^{-3} (pure ice at 0°C , Oke (1978)).

675 In all simulations, the largest distance from the gauging station at which the conduits would take
676 4 months to close was between 1820 m and 1844 m, around 3 km downglacier of S12 and S14.
677 Furthermore, if a subglacial conduit was broad and low rather than semi-circular (which is
678 suggested by the form of the proglacial stream outlet), closure rates would be even faster than
679 estimated (Hooke et al., 1990). It is therefore concluded that the conduits emanating from the
680 S12 and S14 moulins would have closed over the winter.

681 **B Hydraulic damming of the channelized system in July 2011**

682 Dye tracing of the upper glacier moulins in 2011 showed a pattern of slower and more dispersed
683 traces in late July compared to June, with a change back to faster and less dispersed traces in
684 early August. The decreased channel efficiency in late July may have reflected increased conduit
685 roughness, caused by a smaller discharge allowing boulders and cobbles on the conduit floor to
686 decrease flow velocity (Gulley et al., 2012). However, June and July proglacial discharges were
687 similar and the dispersion was lower in June than in July. Rapid changes in transit velocity can
688 also result from inflow modulation and/or changes in the channel geometry (Nienow et al.,
689 1996; Schuler and Fischer, 2009). Similar patterns were observed from three different moulins at
690 similar times on different days (Table 6), so it is unlikely that an increase in the diurnal
691 supraglacial input discharges, resulting in inflow modulation of the tracer transit velocity, was the

692 cause of the differences between the July and August traces. More plausible is that a period of
693 cold weather (air temperatures $< 10^{\circ}\text{C}$ at LWS) reduced meltwater inputs between 17 and 27 July
694 2011 (Fig. 4b), resulting in partial closure of the main subglacial conduit (Röthlisberger, 1972).
695 The increase in air temperatures after 28 July 2011 would have resulted in increased inputs which
696 could not be fully accommodated within the conduit, resulting in hydraulic damming and a
697 decrease in channel efficiency. This would have resulted in the lower n and greater D observed in
698 July. Hydraulic damming of the main subglacial conduit was caused by changes in the channel
699 geometry rather than diurnal variations of supraglacial or proglacial discharge. The similarity of
700 response from the three streams (S12, S14 and S15) suggests that they drain to the same conduit
701 and that any damming occurred downstream of their confluences with this conduit. This
702 interpretation is corroborated by a measured increase in glacier surface velocity over the same
703 period, which implies that the effective normal stress was reduced by elevated basal water
704 pressure as increasing water discharges were forced across large areas of the bed (Fyffe, 2012).
705 Conduit diameters likely grew rapidly so that by August the network could accommodate the
706 increased discharges.

707 Table 1 Summary of relevant papers addressing the hydrology of debris-covered glaciers,
 708 specifically relating to the form and evolution of the supraglacial, englacial and/or subglacial
 709 system. The current paper is added for completeness. Studies concentrating exclusively on the
 710 location and evolution of ponds and lakes have been excluded for brevity.

Study	Location (only debris-covered glaciers)	Investigation method	Key results
Benn et al. (2009)	Khumbu Glacier, Nepal	Direct investigation using speleological techniques	Englacial drainage in conduits. Conduit formation initiated by hydrofracturing.
Benn et al. (2017)	Ngozumpa Glacier, Nepal	Direct investigation using speleological techniques and analysis of satellite imagery and DEMs.	The drainage system was composed of supraglacial channels and seasonal subglacial drainage beneath the upper ablation zone, submarginal channels, perched ponds which occasionally link to the englacial system, cut-and-closure conduits and a moraine-dammed base-level lake.
Gulley and Benn (2007)	Ngozumpa Glacier, Ama Dablam Glacier and Lhotse Glacier, Nepal	Direct investigation using speleological techniques	Englacial drainage in conduits. Passages develop along debris-filled crevasse traces. Conduits evolve via headward nick point migration and vertical incision.
Gulley et al. (2009a)	Khumbu Glacier, Nepal	Direct investigation using speleological techniques	Englacial drainage in conduits. Conduits can form on uncrevassed areas of debris-covered glaciers via a 'cut-and-closure' mechanism which can occur where debris reduces surface melt. Channels can reach the glacier bed even where the basal ice is cold.
Gulley et al. (2009b)	Khumbu Glacier, Kangri Glacier and Ngozumpa Glacier, Nepal	Direct investigation using speleological techniques and published data.	Shreve-type englacial drainage does not exist. Englacial conduits are formed via hydrofracturing given sufficient water supply or via 'cut-and-closure' if channel incision is faster than surface melt. Conduits which exploit debris bands or debris-filled crevasse traces are found on stagnant, low gradient tongues of debris-covered glaciers.
Hasnain et al. (2001)	Dokriani Glacier, India	Dye tracing in July, August and September	Drainage was via efficient trunk channels in July, but via an inefficient distributed system in August and September.
Kirkbride and Spedding (1996)	Meuller Glacier and Tazman Glacier, New Zealand	Mapping of supraglacial debris characteristics	Existence of past englacial conduits inferred from rounded, water-worked debris found on the surface.
Miles et al. (2017)	Lirung Glacier, Nepal	Measurements of pond water level, DEM-based analysis of catchments and field-based observations of supraglacial hydrology	Ponds represent an area of reduced drainage efficiency in a coupled supraglacial and englacial drainage system. Pond drainage via inefficient conduits. Supraglacial drainage system configuration follows relict englacial conduit systems.
Pottakkal et al. (2014)	Gangotri Glacier, India	Dye tracing of channels from Chaturangi and Raktavarn tributary glaciers as they flow through the Gangotri	The tributary streams are transported within an efficient channelized system beneath the Gangotri Glacier. The pathway from the supraglacial channel was less efficient than the tributary stream channels.

		tongue. Plus one injection into a supraglacial stream.	
Röhl (2008)	Tasman Glacier, New Zealand	Limnological and glaciological measurements of supraglacial ponds.	As well as low surface slope ($<2^\circ$) and velocity, pond development is determined by the pond's connection to the englacial drainage system.
Fyffe et al., (2018)	Miage Glacier, Italy	Dye tracing over two ablation seasons. Topographical analysis and supraglacial stream measurements.	An efficient conduit system drains the upper ablation zone whereas the lower continuously debris-covered area is drained by an inefficient network which drains to the channelized network. The intermoraine troughs in the upper ablation area concentrate drainage whereas the hummocky topography of the continuously debris-covered region results in smaller catchments.

712 Table 2 Details of supraglacial and proglacial stream instruments. *The manufacturer's shade cap
 713 was used on the Turner fluorometer (sensor shading is integrated into the design of the Seapoint
 714 fluorometer).

Quantity	Location	Time period	Manufacturer	Type	Accuracy
Stage	Proglacial	2010 and June 2011	GE Sensing	Druck PTX1830 (vented)	$\pm 0.1\%$ full scale (or $\pm 0.06\%$ full scale)
	Proglacial	Aug and Sep 2011	Onset	HOBO U20 - 001-04 (non- vented)	$\pm 0.075\%$ full scale, ± 0.3 cm
Fluorescence	Proglacial	2010 and June 2011	Seapoint	Rhodamine fluorometer	Not stated but minimum detection 0.02 ppb
	Proglacial	July, Aug, Sep 2011	Turner	Cyclops-7 Rhodamine*	Not stated but minimum detection 0.01 ppb
Conductivity	Supraglacial	2010 and 2011	Hanna	HI9033 with HI 76302W probe	$\pm 1\%$ full scale (excluding probe)

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717 Table 3 Parameters calculated for each dye breakthrough curve.

Symbol	Unit	Definition
u	m s^{-1}	The minimum estimate of the average transit velocity of the tracer through the hydrological system (d/t).
d	m	The straight line distance from the gauging station to the injection site. Due to the bend in the glacier above S4, for all injections above this point the distance between the injection point and S4 was used and added to the distance between S4 and the gauging station to give the total distance.
t	s	The time between the injection and peak of the return curve.
D	$\text{m}^2 \text{s}^{-1}$	The dispersion coefficient, which is a measure of the spread of the dye as it travels through the glacier. It is calculated from: $D = \frac{d^2(t-t_i)^2}{4t^2t_i \ln \left[2 \left(\frac{t}{t_i} \right)^{\frac{1}{2}} \right]}$ <p>(Seaberg et al., 1988, p222). Two variants of the equation are calculated: one with t_i the time from injection to half of the dye concentration peak on the rising limb, and the other with t_i the time from injection to half of the dye concentration peak on the falling limb. In this equation t is not measured but found iteratively by determining the value which minimises the difference between the two variants of the equation. The calculated value of t is then used to compute D with either value of t_i.</p>
b	m	The dispersivity, calculated as D/u (Seaberg, 1988, p224).
A_m	m^2	The apparent mean cross-sectional area, calculated as Q_m/u .
A_c	ppb minute	The area under the dye breakthrough curve, calculated by summing all of the dye concentration values composing the breakthrough curve and multiplying this by the logging interval between measurements.
Q_m	$\text{m}^3 \text{s}^{-1}$	The mean discharge between the injection and detection point, calculated as the average of the supraglacial (assumed constant) and proglacial (average of the discharge at the injection and peak of the return curve) discharge.
Q_p	$\text{m}^3 \text{s}^{-1}$	The average proglacial discharge from the time of injection until the time of the peak of the dye return curve.
P_r	%	The percentage dye return ($(V_r/V_i)*100$).
V_r	ml	The volume of dye recovered, calculated from the equation used to calculate discharge from dilution gauging given by Kilpatrick and Cobb (1985, p6): $V_r = \frac{S^{-1} \left(\frac{1}{1.649 \times 10^{-8}} (Q_p A_c) \right)}{c_{di}},$ <p>where S is the specific gravity of the dye used (1.15 for rhodamine WT).</p>
V_i	ml	The volume of dye injected.
c_{di}	ppb	The concentration of dye prior to injection.

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Table 4 Catchment statistics for each surface cover type.

Surface Type	Number of catchments	Maximum	Mean	Mean of largest 10 Catchment size (m²)	Standard deviation
Debris-covered Ice	2625	56093	1588	39172	3632
Clean Ice	4075	211421	1295	85055	5920
Dirty Ice	207	143704	4900	72899	17812

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Table 5 Dye trace parameters for all injection points in 2010, for definitions see Table 3. *Only part of the rising limb of the trace was returned. Mean P_r does not include values $>100\%$, which can be caused by error in Q_p or variations in the background fluorescence which alters A_c .

Name	Date	Time	V_i (ml)	Trace?	u (m s ⁻¹)	D (m ² s ⁻¹)	b (m)	Q_p (m ³ s ⁻¹)	A_c (ppb minute)	P_r (%)
S1	5 June 2010	17:51:00	~4	N	Too little dye injected.					
S2	8 June 2010	16:00:00	40	N	No trace detected.				N	
S6	9 June 2010	17:46:05	40	Y	0.583	0.884	1.52	2.88	20.8	37.7
S8	10 June 2010	12:12:00	120	Y	0.434	1.180	2.72	2.90	55.9	34.0
S13	11 June 2010	12:43:00	200	Y	0.830	1.800	2.17	3.36	129.4	54.6
S1	12 June 2010	12:05:00	40	Y	0.024	0.004	0.15	5.97	34.4	129.0
S10	13 June 2010	15:07:00	160	Y	0.602	2.300	3.82	5.70	40.0	35.8
S3	14 June 2010	16:50:00	80	Y	0.192	0.230	1.20	2.84	3.7	3.3
S9	18 June 2010	17:45:00	120	N	Fluorometer not working.					
S3	19 June 2010	14:25:00	80	N	Missing data.					
S5	20 June 2010	13:21:30	80	N	Missing data.					
S3	29 July 2010	17:52:00	80	Y	0.345	0.860	2.49	10.71	50.6	170.0
S5	30 July 2010	16:15:00	120	Y	0.226	9.490	42.01	5.63	47.4	55.9
S9	31 July 2010	12:11:00	120	N	Fluorometer not working.					
S11	1 Aug 2010	11:32:00	120	Y	0.442	3.550	8.03	7.80	56.3	91.8
S13	3 Aug 2010	12:21:30	160	N	Missing data.					
S16	4 Aug 2010	12:01:00	200	N	Missing data.					
S5b	6 Aug 2010	16:10:00	80	Y*				2.98		
S13	5 Sep 2010	12:15:10	160	N	Missing data.					
S14b	6 Sep 2010	14:30:30	200	Y	0.613	1.770	2.89		181.9	
S3	9 Sep 2010	14:27:00	80	Y	0.265	1.870	7.05	1.65	100.2	51.9
S4	10 Sep 2010	15:56:00	80	N	No trace detected.					
S12b	11 Sep 2010	15:47:00	100	Y	0.318	7.800	24.55	1.93	141.5	68.6

Mean (all)	0.406	2.645	8.21	4.63	71.8	48.2
Mean (upper)	0.561	3.444	8.29	4.74	109.8	62.7
Mean (lower)	0.296	2.074	8.16	4.57	44.7	36.6

Table 6 Dye trace parameters for all 2011 dye injections. The Q_s and u_s type is either ‘D’, dilution gauging, ‘V’, the velocity area method (timing of floats), or ‘AdD’, adjusted to dilution gauging (see Sect. 3.2.2 for details). *Indicates traces with multiple peaks for which the D and b parameters are less reliable. **Only the first part of the trace was returned. ***A trace was returned but was poor quality so has not been interpreted. †The Q_s values are an estimate because the stream cross-sectional area could not be measured, in these cases the mean cross-sectional area was multiplied by the velocity. Means are for detected traces only and mean P_r does not include values >100%. Since the P_r for S5_12Sep11 exceeds 100% this may indicate that the spikes on the tail of the main peak (Fig. 7b) are erroneous.

Name	Date	Time	V_i	Trace?	u	D	b	Q_p	A_c	P_r	Q_s	Q_s	u_s	u_s	A_m	
			(ml)		(m s ⁻¹)	(m ² s ⁻¹)	(m)	(m ³ s ⁻¹)	(ppb minute)	(%)	(m ³ s ⁻¹)	type	(m s ⁻¹)	type	(m ²)	
S7	5 June 2011	19:02:00	160	Y	0.073	2.907*	11.51*	2.14	70.1	23.5						
S5	6 June 2011	15:43:30	120	Y	0.070	14.70*	178.58*	2.08	83.9	36.6	0.027	D	0.24	D	14.68	
S15	8 June 2011	17:28:30	280	N	Missing data.							0.027	D	0.44	D	
S14	9 June 2011	15:57:00	280	N	Missing data.							0.535	V	1.14	V	
S12	10 June 2011	16:22:00	280	Y	0.510	0.700	0.02	2.09	466.8	87.4	0.025	AdD	0.44	AdD	2.06	
S7	11 June 2011	16:31:00	240	Y	0.124	2.070	3.88	2.01	124.0	26.1	0.011	AdD	0.17	AdD	8.14	
S5	12 June 2011	15:35:00	200	Y	0.070	9.380*	113.82*	2.21	109.8	30.5	0.032	D	0.25	D	15.88	
S15	13 June 2011	13:17:30	200	Y	0.283	71.400	144.08	3.00	123.1	46.3	0.013	D	0.27	D	5.36	
S14	14 June 2011	13:01:00	200	Y	0.583	1.300	0.06	2.35	284.5	83.9	0.438	V	1.24	V	2.39	
S3	15 June 2011	10:36:00	80	Y**												
S5	27 July 2011	13:00:40	200	Y	0.229	1.980	9.91	1.98	207.5	51.6	0.031	D	0.13	D	4.38	
S15	28 July 2011	15:28:30	240	Y	0.439	1.570	0.22	2.85	196.4	58.6	0.010	D	0.27	D	3.25	
S14	29 July 2011	15:12:00	160	Y	0.470	2.600	0.83	1.87	74.7	21.9	0.874†	V	2.13	V	2.92	
S12	30 July 2011	14:45:40	160	Y	0.487	9.300	5.23	2.16	68.6	23.2	0.341	D	0.43	D	2.56	
S7	31 July 2011	13:13:30	200	N	Background very variable.							0.028	D	0.24	D	
S14	1 Aug 2011	12:07:30	120	Y	0.731	1.240	0.26	4.47	41.0	38.3	0.888†	V	2.16	V	3.66	
S15	1 Aug 2011	14:43:00	120	Y	0.576	1.230	0.35	4.47	42.9	40.1	0.014	D	0.30	D	3.89	
S12	2 Aug 2011	14:45:30	160	Y	0.699	1.440	0.22	4.47	69.7	48.8	0.147	AdD	0.50	D	3.30	
S7	3 Aug 2011	13:50:00	190	N	No trace detected.							0.032	D	0.28	D	

S5	4 Aug 2011	11:19:35	195	N	Fluorometer removed before dye detected.						0.028	D	0.14	D	
S5	12 Sep 2011	17:10:00	200	Y	0.063	0.09*	1.16*	7.22	179.9	163.0					
S15	13 Sep 2011	13:28:30	240	Y	0.578	4.50	0.47	5.16	134.6	72.6	0.022	D	0.50	D	4.43
S14	14 Sep 2011	12:26:00	120	Y	0.697	1.40	0.27	6.02	45.0	56.6	0.378†	V	0.92	V	4.60
S12	14 Sep 2011	15:15:00	160	Y	0.593	3.54	1.16	6.34	71.4	71.0	0.196	D	0.49	D	5.54
S7	15 Sep 2011	14:19:00	200	Y***							0.006	D	0.25	D	
Mean (all)					0.404	7.30	26.22	3.49	133.0	48.0	0.203		0.63		5.44
Mean (upper)					0.554	8.35	12.76	3.77	134.9	54.1	0.279		0.80		3.66
Mean (lower)					0.105	5.19	53.14	2.94	129.2	33.6	0.022		0.21		10.77

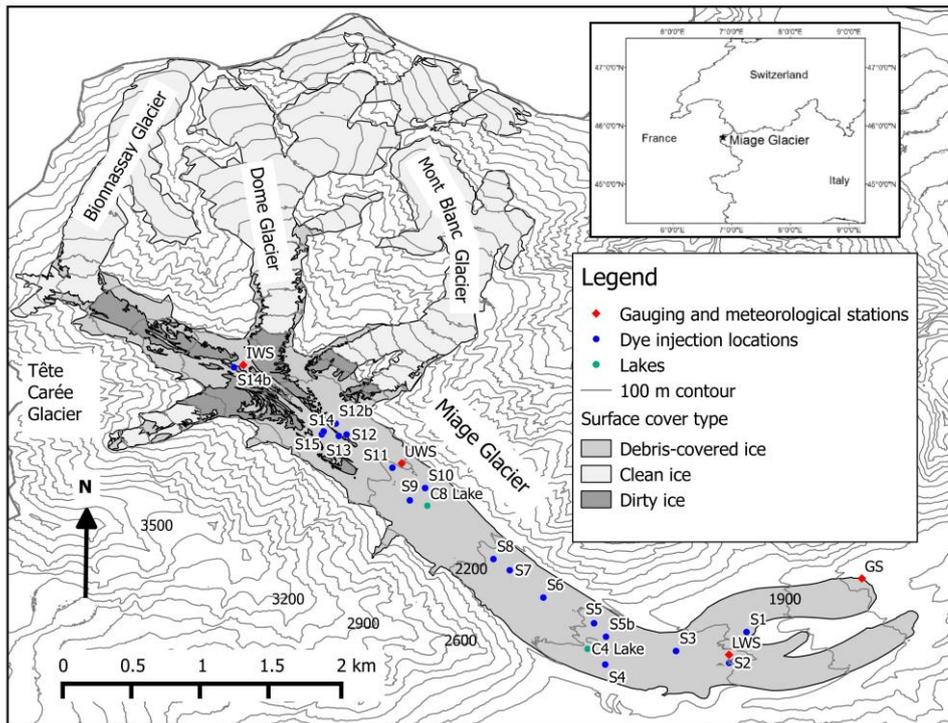


Figure 1 Map of Miage Glacier showing location of monitoring stations, lakes and dye tracing points. Inset shows location of Miage Glacier in the Alps. 'IWS' is the ice weather station, 'UWS' the upper weather station, 'LWS' the lower weather station and 'GS' the gauging station. S_i denote injection site locations.

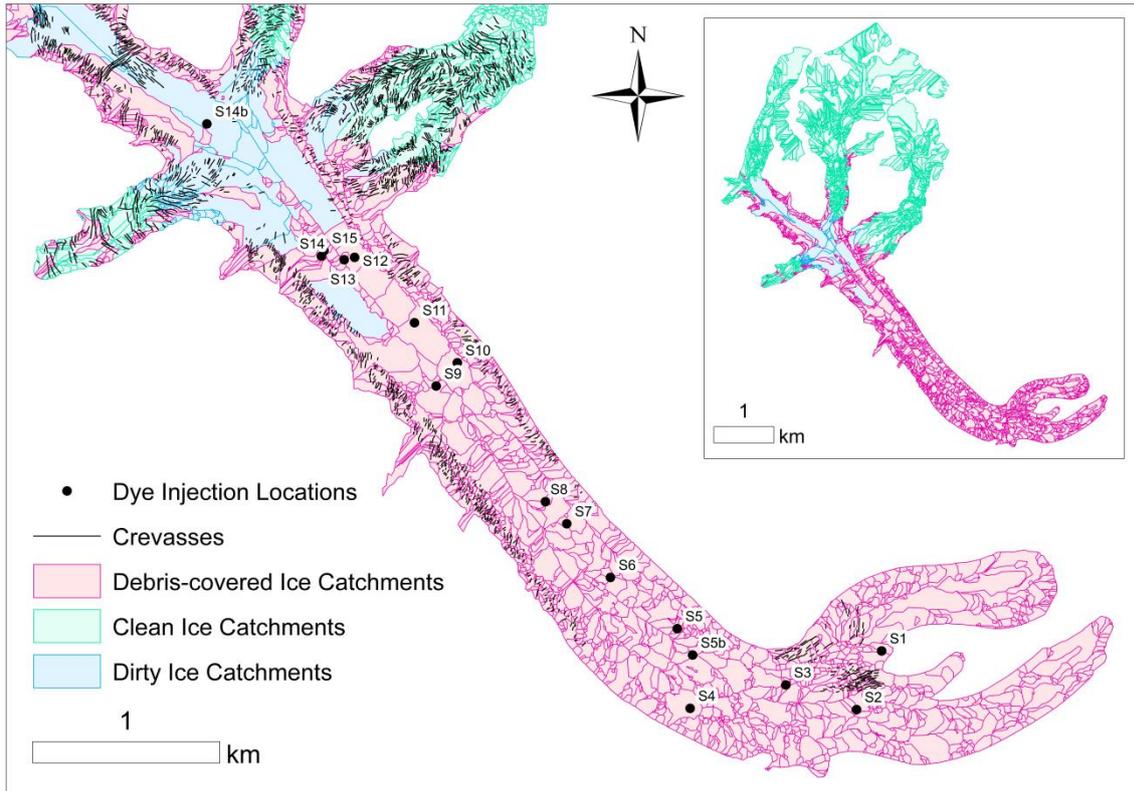


Figure 2 Supraglacial catchments distinguished by surface cover type. The inset map gives an overview for the entire glacier. The dye injection locations are given for context only since the DEM (and therefore catchment outlines) are relevant to 2008 whereas the injection locations are from 2010 and 2011.

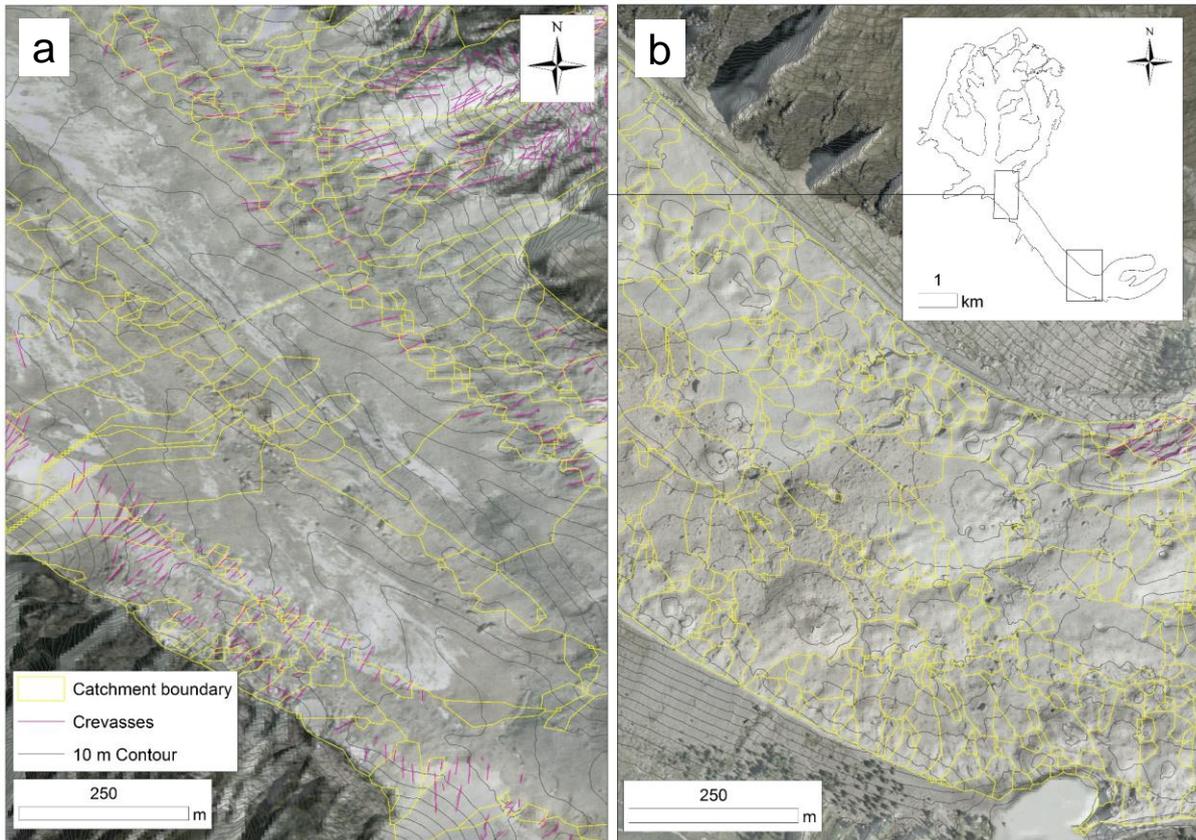


Figure 3 Topographic influence on supraglacial hydrology. Panel a) shows the clear along-glacier ridge and valley topography associated with the central, eastern and western moraines on the upper tongue which results in relatively large catchments, with panel b) showing the hummocky topography on the lower glacier. Both panels show contours at 10 m intervals. Source: Regione Autonoma Valle d'Aosta DEM.

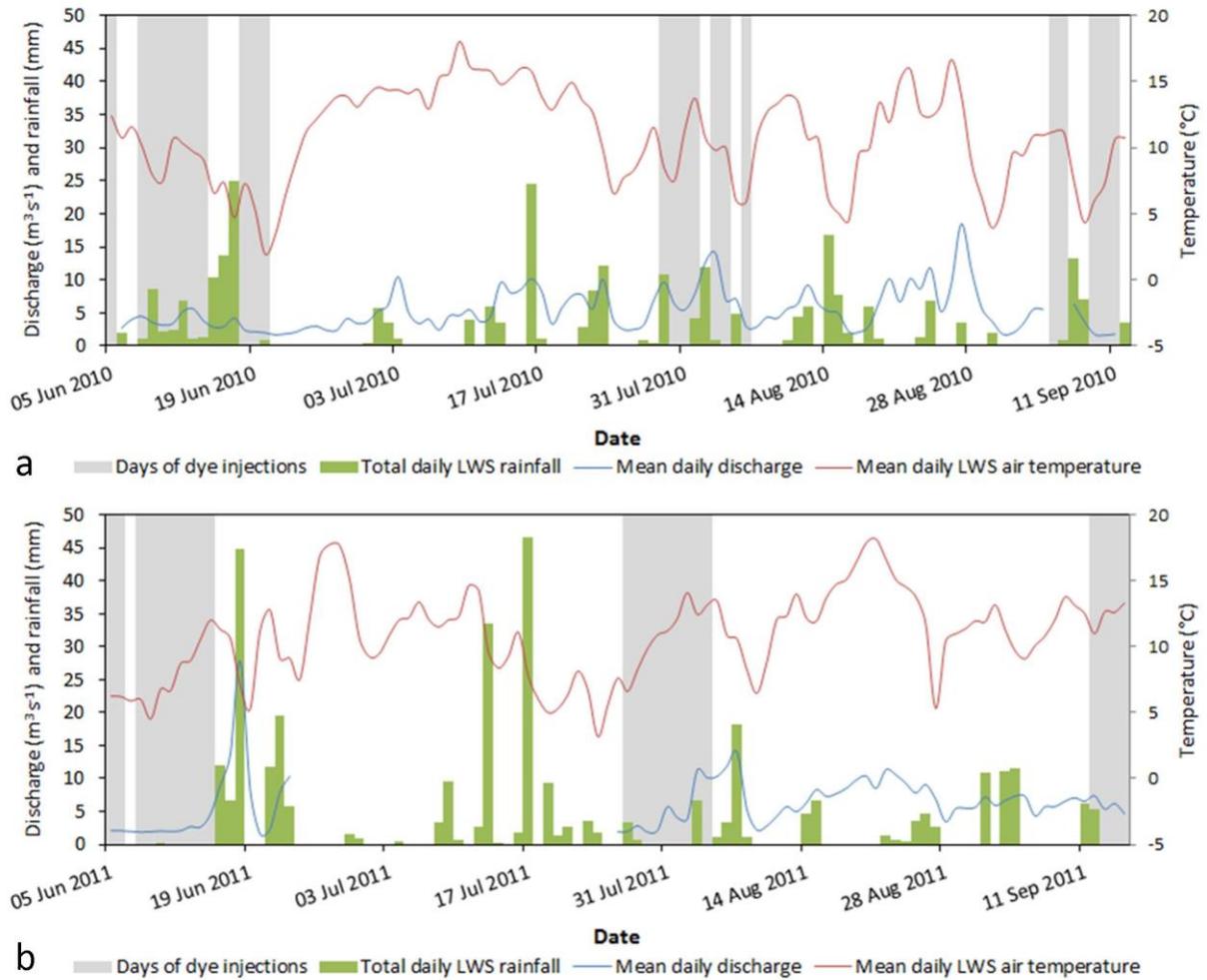


Figure 4 Meteorological conditions and proglacial discharge during the a) 2010 and b) 2011 field seasons. Grey bars indicate days when dye injections were conducted.

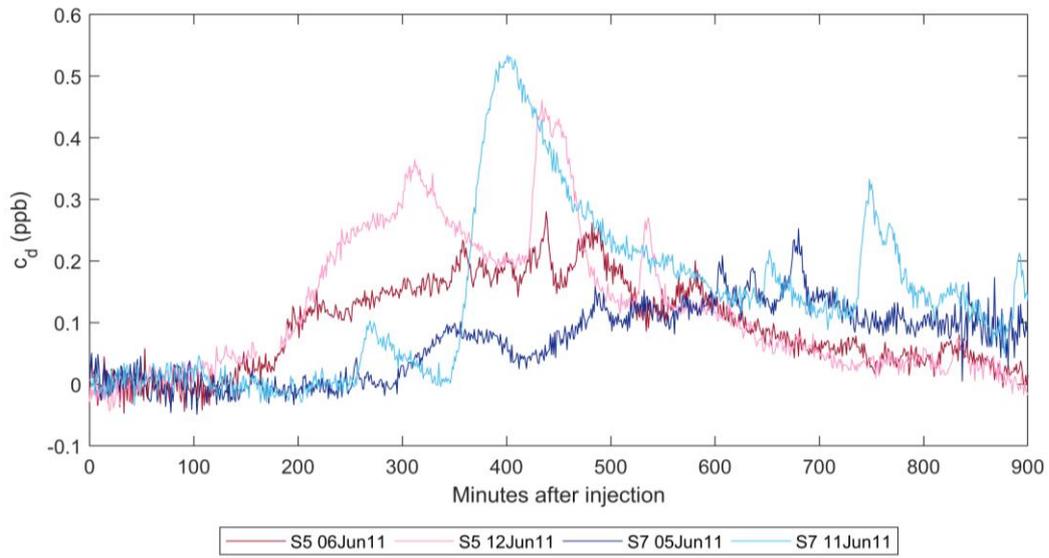


Figure 5 Dye breakthrough curves from S5 and S7 only in June 2011, where c_d is the dye concentration. Note the close correspondence in the rising and falling limb of the S5 traces which gives confidence that these traces are due to dye rather than background fluctuations.

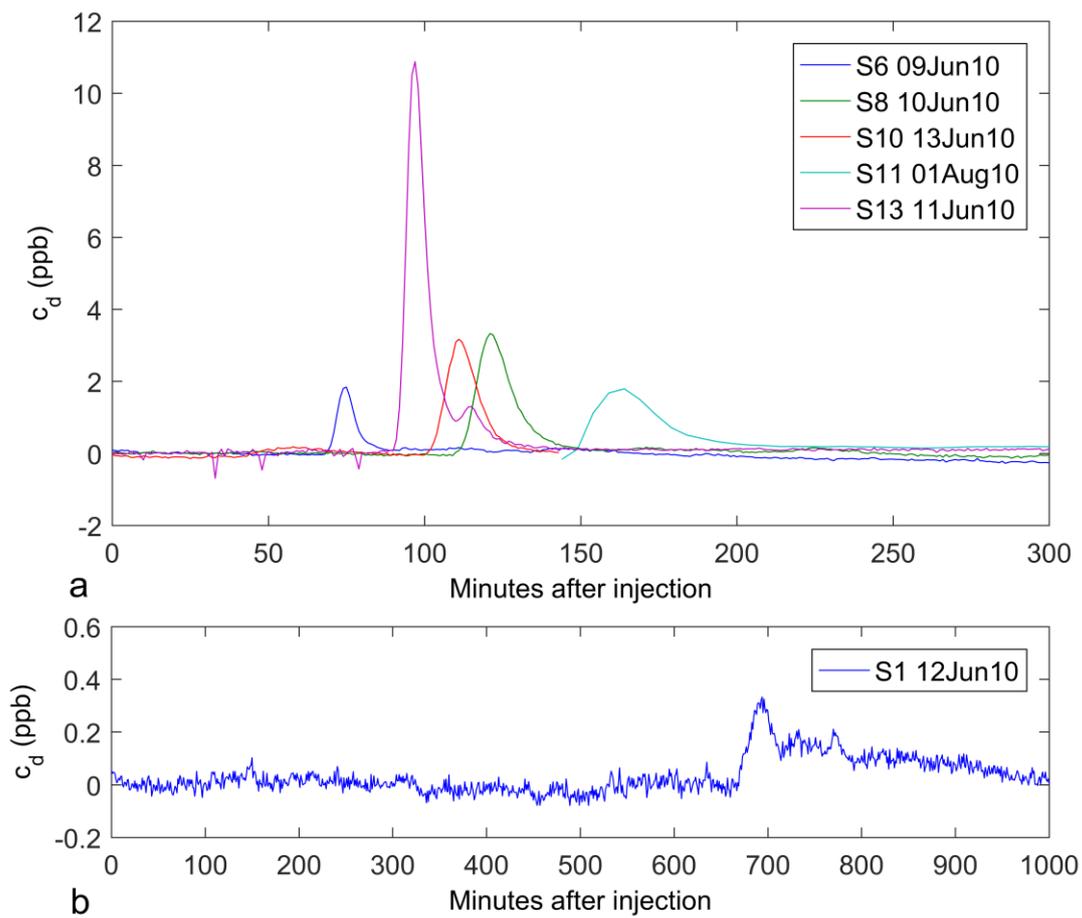


Figure 6 Dye return curves from streams that were only injected once (injections conducted in 2010), where c_d is the dye concentration. Note that vertical and horizontal scales differ between subplots.

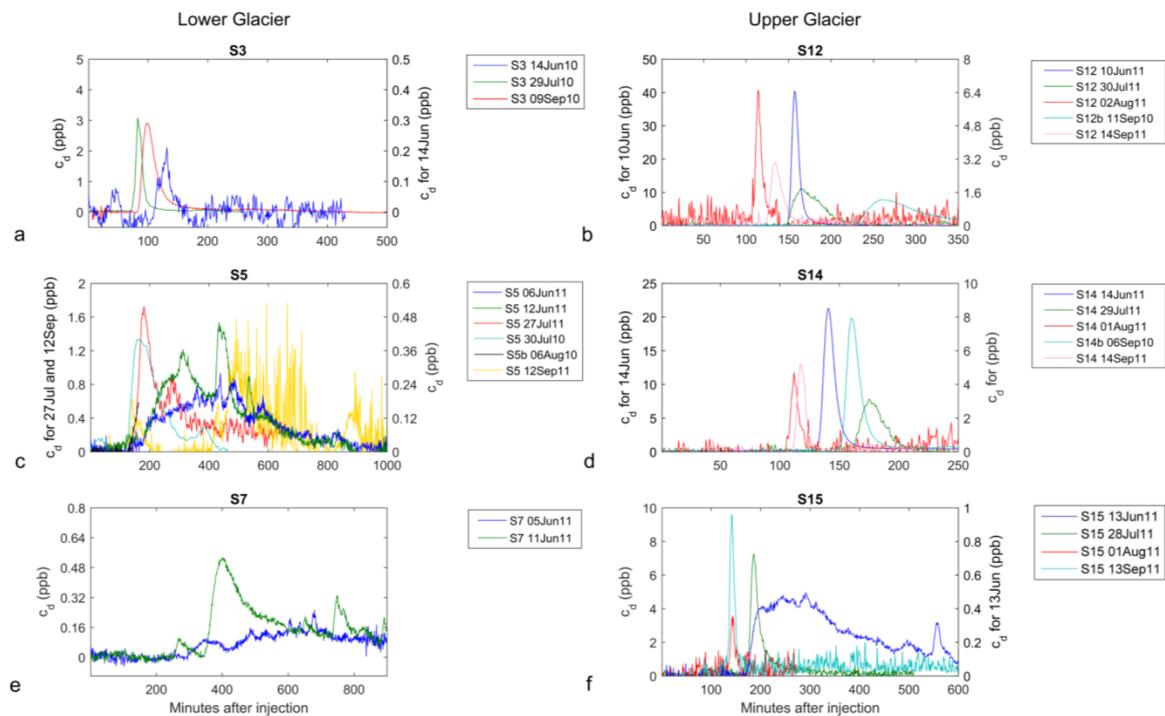


Figure 7 Repeat dye return curves from single injection points (including injections performed in both 2010 and 2011), where c_d is the dye concentration. The injection points S3, S5 and S7 (a, c and e) are on the lower glacier, while injection points S12, S14 and S15 (b, d and f) are on the upper glacier. Note that vertical and horizontal scales differ. The S3_15Jun11 and S7_15Sep11 return curves are not shown due to being poor quality curves which were not interpreted.

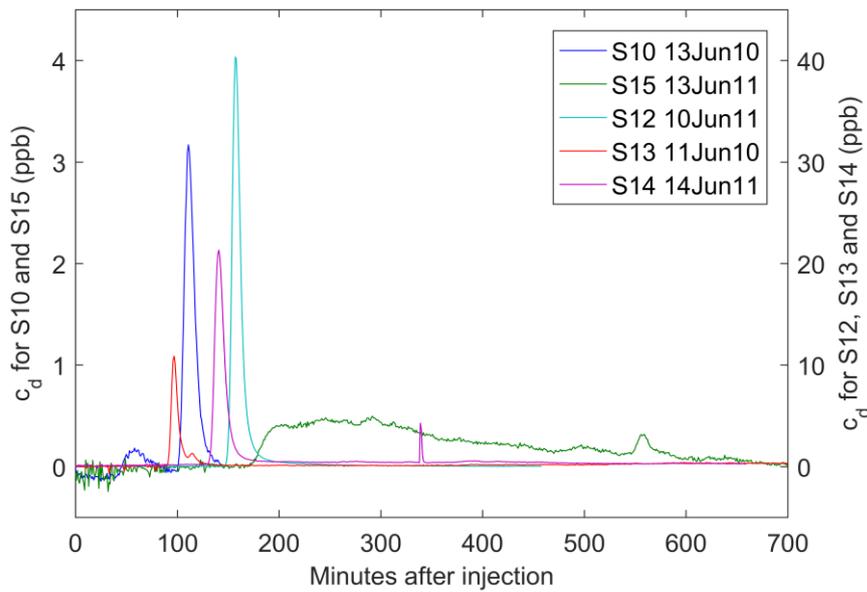


Figure 8 Dye return curves from the upper glacier streams injected in June of both 2010 and 2011, where c_d is the dye concentration. Note that vertical scales differ.

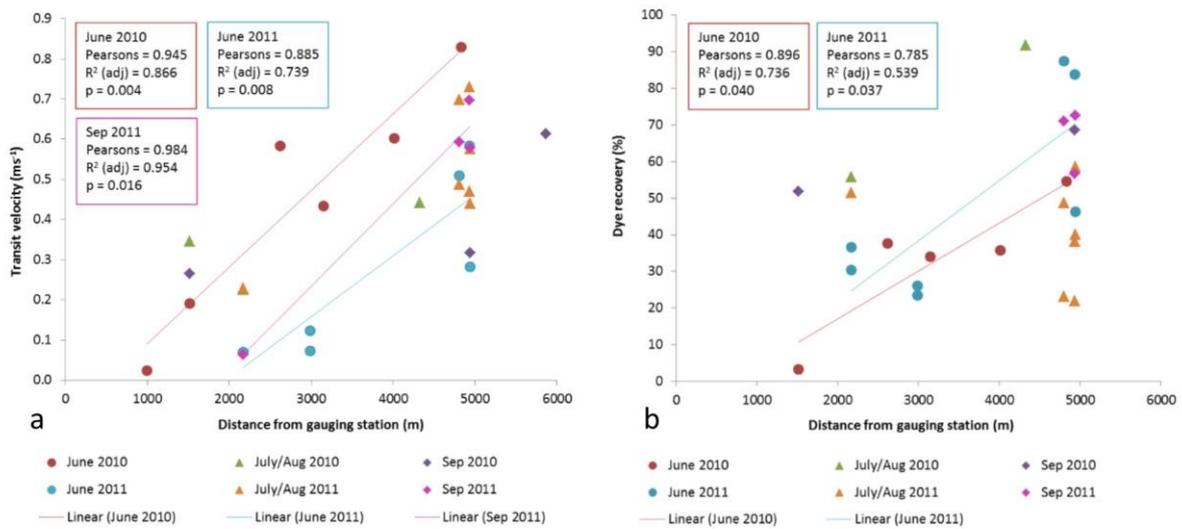


Figure 9 Relationship between the distance to gauging station and a) return curve u , and b) return curve P_r , including all 2010 and 2011 data. P_r in b) does not include values over 100%. Data have been split by field campaign with linear regression and associated parameters only shown when results were significant ($p < 0.05$). Correlation and regression was not performed on July/Aug and September 2010 P_r data because there were only two points in the dataset.

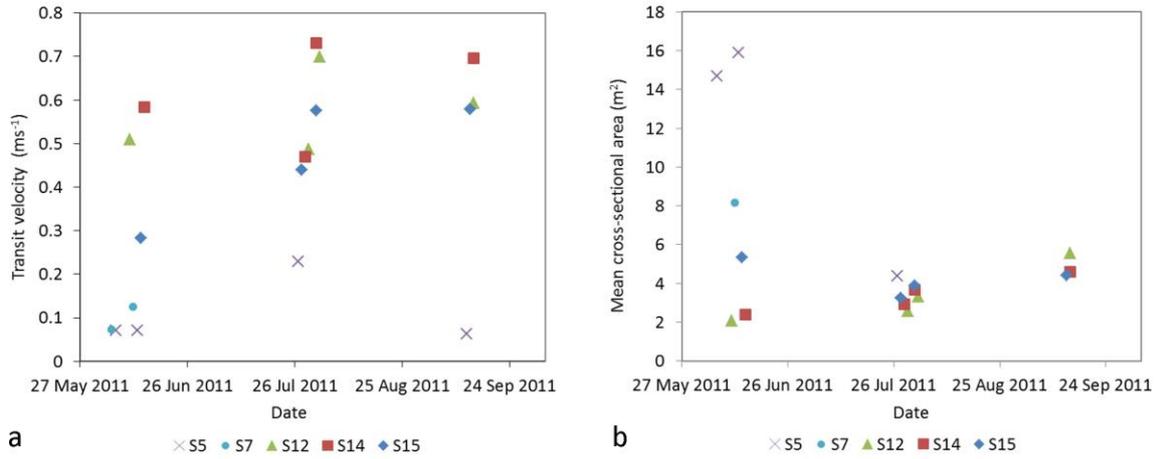


Figure 10 a) Return curve n variations over the 2011 season, and b) mean A_m variations over the 2011 season.

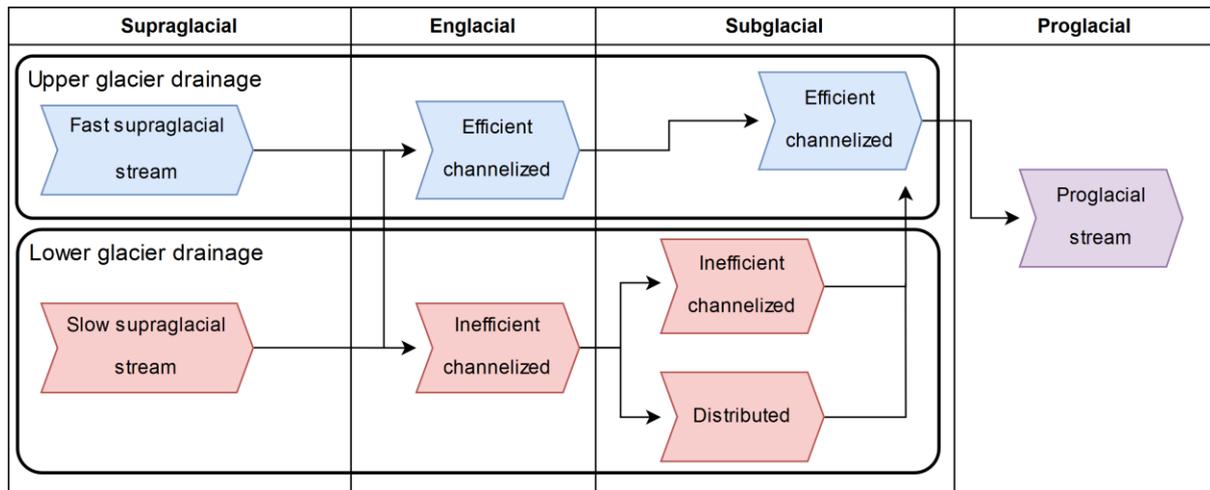


Figure 11 Implied drainage system structure of Miage Glacier. Lower glacier streams may make a more direct connection to the efficient englacial/subglacial system, as shown by the link between the englacial paths. The upper glacier drainage does pass beneath the lower glacier to reach the proglacial stream.



Supplementary Material a) dye tracing the S12 stream in September 2011 and b) dye tracing S14 in July 2011.