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Glacial Sediment Stores and Their Reworking

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1 4 HILLSLOPE PROCESSES IN THE PROGLACIAL ZONE

2 4.1 GLACIAL SEDIMENT STORES AND THEIR RE-WORKING

3 INTRODUCTION

4 Understanding the storage and flux of sediments through the proglacial
5 zone of deglaciating catchments is an important topic within glacial geo-
6 morphology, primarily as there is great uncertainty surrounding the time-
7 scales that sediments are stored and released, which has implications for
8 fully understanding landform genesis and the response of landforms and
9 glaciated catchments to current and forecast environmental changes. The
10 steep slopes found in alpine glacial environments, combined with associ-
11 ated potential for intense geomorphological activity, offer favourable con-
12 ditions for the release, reworking and storage of glacial sediments and play
13 an important part in the geomorphic coupling and overall connectivity of
14 the down-valley sediment cascade (Cavalli *et al.*, 2013; Heckmann and
15 Schwanghart, 2013; Carrivick and Heckmann, 2017; chapter 6.2). It is no
16 surprise, therefore, that processes acting on hillslopes have formed the fo-
17 cus for much research investigating the sedimentary response to deglacia-
18 tion.

19 The concept of a so-called ‘paraglacial period’ of enhanced geomorpholog-
20 ical activity conditioned by glaciation (or indeed deglaciation) originally de-
21 tailed by Church and Ryder (1972) and subsequently developed by several
22 authors, most notably Ballantyne (2002a), has received much attention in
23 recent years, as the geomorphological consequences of deglaciation be-
24 come ever more apparent. Hillslopes and associated slope processes in

25 glaciated regions are likely to play a key role in that heightened geomor-
26 phological activity and associated release, reworking and subsequent re-
27 deposition, or storage, of glacial sediments. However, any sedimentary or
28 ‘paraglacial’, response to deglaciation is complex and different landsys-
29 tems will respond at differing rates and on differing spatial scales (Ballan-
30 tyne, 2002b) and this is certainly the case with hillslopes, which can both
31 store and release sediments at the catchment scale, with potentially signif-
32 icant spatial and temporal variability in both storage and release. Hillslopes
33 in deglaciating environments also demonstrate the attributes of both pri-
34 mary and secondary paraglacial systems. Primary paraglacial systems are
35 those where sediment release is directly conditioned by glaciation (e.g.
36 sediment-mantled slopes) while secondary paraglacial systems are those
37 where rates of sediment release are additionally controlled by reworking
38 of paraglacial sediment stores (e.g. debris cones and fans at the base of
39 drift-mantled slopes. Ballantye, 2002b). The presence, operation and in-
40 teraction of these primary and secondary systems inevitably add complexi-
41 ty to the deglaciation response.

42 Direct glacial influence on sediment release and reworking is likely to de-
43 crease as distance down valley increases, but smaller glaciers have been
44 observed to play an important role in efficiently coupling slope and fluvial
45 processes (e.g. Lukas *et al.*, 2005), ensuring that sediment fluxes from gla-
46 ciated alpine catchments are dominated by fluvial processes (e.g. Orwin *et*
47 *al.*, 2010; Carrivick *et al.*, 2013). However, critical to that fluvial reworking
48 and redistribution is the supply of sediments from adjacent slope units and
49 the degree of connectivity between slope and fluvial systems, both of
50 which will also vary in space and time (e.g. Cavalli *et al.*, 2013; Heckmann

51 and Schwanghart, 2013; Lane *et al.*, 2017; chapter 6.2). Potential supply
 52 limitation from those slope units to fluvial systems therefore becomes an
 53 important parameter partially dictating the extent to which fluvial systems
 54 act as an effective mechanism of sediment redistribution into the down-
 55 valley sediment cascade (Cavalli *et al.*, 2013). Processes acting upon
 56 hillslopes and associated storage, release and reworking of sediments in
 57 the hillslope domain therefore play a significant role in controlling the ba-
 58 sin-scale flux and yield of sediments during deglaciation.

59 It could be argued however, that in comparison with the wealth of re-
 60 search conducted in high-latitude environments, investigation of the
 61 paraglacial response to deglaciation in alpine environments is relatively
 62 under-studied. This is perhaps unsurprising, given the very obvious and
 63 dynamic geomorphological activity and landscape modification associated
 64 with processes of thermo-erosion and mass movement at high latitudes
 65 (e.g. Etzelmüller, 2000; Lyså, and Lønne, 2001; Irvine-Fynn *et al.*, 2005;
 66 Porter *et al.*, 2010; Irvine-Fynn *et al.*, 2011; Ewertowski and Tomczyka,
 67 2015). In comparison, many deglaciating alpine environments, at least su-
 68 perfcially, appear to exhibit relative stability following Little Ice Age glacier
 69 retreat and subsequent stabilisation of many proglacial areas through, for
 70 example, relatively undisturbed vegetation colonisation well beyond pio-
 71 neer stages (e.g. Eichel *et al.*, 2013; Eichel *et al.*, 2015). However, as degla-
 72 ciation gathers pace in alpine regions (e.g. Barry, 2006, Radić and Hock,
 73 2011; Klaar *et al.*, 2015), the requirement for a fuller understanding of sed-
 74 iment fluxes becomes enhanced, both due to a need to understand the
 75 complex geomorphological and sedimentological responses to deglaciation

76 and to assess any potential impacts on, or from, human activities in alpine
77 areas (e.g. Moore *et al.*, 2009; Otto *et al.*, 2009; Carrivick *et al.*, 2013).

78 OVERVIEW

79 Perhaps the most visually obvious manifestation of hillslope glaciogenic
80 sediment storage, release and redistribution within deglaciating alpine en-
81 vironments are the large, often dissected, lateral moraines that flank many
82 systems throughout alpine regions (e.g. Curry *et al.*, 2005; Curry, *et al.*,
83 2009; Lukas *et al.*, 2012, Figure 1). Although perhaps less visually obvious,
84 the extensive suites of sedimentary ice-marginal landforms and features
85 found in proglacial zones also represent an important store and source of
86 glacial and related sediments associated with deglaciation, likely subject to
87 extensive sediment redistribution and re-working that may lead to both
88 modification of landscape morphology and impacts on sediment fluxes
89 through the down-valley cascade during deglaciation.

90 In this chapter we therefore consider storage within, and supply of sedi-
91 ments from, slopes within ice-marginal and proglacial environments. Inevi-
92 tably however, there is the potential for overlap with the content of other
93 contributions in this volume, as hillslopes represent a fundamental and
94 pivotal connection between geomorphic systems, controlling both the
95 supply and storage of sediments and being subject to the full range of ge-
96 omorphological processes. We therefore limit our consideration to land-
97 forms in the proglacial zone, as these features not only represent sizeable
98 and important stores of glaciogenic and other sediments, but represent a
99 highly dynamic geomorphological environment that impacts upon and in-

- 100 teracts with the multiple geomorphic and biogeomorphic systems within
101 the basin sediment cascade that are considered elsewhere in this volume.
102

104 Figure 1. *Deglaciating proglacial alpine landscapes, Feegletscher Nord, Va-*
 105 *lais Switzerland. A. Aerial view showing recent rockfall and slope debris to*
 106 *the left of the image, proglacial lake dammed by moraines and rockfall de-*
 107 *bris and heavily dissected lateral moraine sequence to the top right of the*
 108 *image. B. Proximal face of lateral moraines showing extensive dissection*
 109 *and gullyng, with debris cone build-up progressively burying slope units*
 110 *with increasing distance from the glacier snout (out of shot behind the pho-*
 111 *tographer). C. Gullied lateral moraine slopes showing distal dipping fabric*
 112 *and slope foot debris accumulation. Note the mature vegetation at the*
 113 *slope crest, which extends down the distal slope away from the camera.*

114 Although subglacial and supraglacial sediments also represent an im-
 115 portant source from which the sediments contained within ice-marginal
 116 landforms and features may ultimately be derived (Barr and Lovell, 2014),
 117 these systems and sediments are considered in sections 3.3 and 3.4 re-
 118 spectively and are therefore not discussed in detail here. Similarly, the re-
 119 duction in lateral slope support associated with deglaciation, known as de-
 120 buttressing, is likely to enhance the likelihood of rockfall activity (e.g.
 121 Stoffel *et al.*, 2014; Vehling *et al.*, 2016) and associated delivery of pre-
 122 dominantly larger calibre sedimentary debris to moraine sequences (e.g.
 123 Shulmeister *et al.*, 2009; Cossart *et al.*, 2008; Reznichenko *et al.*, 2011),
 124 while periglacial systems such as rock glaciers also represent an effective
 125 store and potential source of slope sediments (Stoffel and Huggel, 2012;
 126 Müller *et al.*, 2014) that may be subsequently reworked. Rockfall activity
 127 however is discussed in detail in section 4.1, while rock glaciers are consid-
 128 ered in section 3.5.

129 Therefore, rather than considering the broader scale catastrophic and/or
130 slow mass movements which may arise as a result of the same climatic
131 forcing that drives deglaciation and contribute additional sediments for
132 paraglacial reworking and redistribution, here we limit our consideration
133 to those sedimentary slope units directly related to glacial activity and sub-
134 ject to paraglacial modification. These comprise lateral and forefield glacial
135 and glaciofluvial sedimentary landforms, their component slope units and
136 processes operating thereon.

137 LATERAL SLOPES

138 Formation and fluxes

139 Although large suites of lateral moraines commonly exhibiting numerous,
140 narrow, parallel ephemeral channels, hereafter referred to as gullies, are a
141 characteristic of many deglaciating alpine systems (Figure 1), the details of
142 lateral moraine formation and internal structure remain incompletely un-
143 derstood. This situation has arisen in part, due to moraine spatial location
144 with respect to the contemporary glacier front and patterns of glacier fluc-
145 tuation based thereon, being the focus for much research to date (Lukas
146 and Sass, 2011; Lukas *et al.*, 2012). Early theories of lateral moraine for-
147 mation generally assumed that subaerial weathering and resultant erosion
148 was primarily responsible for accumulation of sediment at the ice margin,
149 with additional contribution of sediment from englacial sources (e.g. Eyles
150 and Rogerson, 1978; Rothlisberger and Schneebeli, 1979; Eyles, 1983). Lat-
151 ter theories invoke processes of repeated ‘stacking’ of ice-derived debris
152 flows at the glacier margins, with sedimentary stratification, gently sloping
153 distal morphology and limited dating providing evidential support for such

154 incremental formation (e.g. Small, 1983, 1987). However, substantial sub-
155 glacial and glaciofluvial sediments observed within lateral moraines at
156 Findelengletscher, Switzerland, contrast with observations made else-
157 where and highlight the potential importance of sediment transfer via en-
158 glacial pathways (Lukas *et al.*, 2012), while work conducted in geomorpho-
159 logically active alpine areas has highlighted the rapid rate at which rockfall
160 debris may become incorporated in the englacial environment and advec-
161 ted towards the glacier margin for subsequent deposition (e.g. Dunning *et*
162 *al.*, 2015). More recent geophysical investigations indicate complex ice-
163 marginal moraine depositional history and evidence of polygenesis (e.g.
164 Midgley *et al.*, 2013; Tonkin *et al.*, 2017). These observations combined,
165 highlight the potential genetic complexity of lateral moraine features. Irre-
166 spective of the precise modes of formation, it is clear that lateral moraine
167 complexes represent a substantial store of glacial and/or glaciofluvial sed-
168 iment (Otto *et al.*, 2009) that has the potential to be extensively re-worked
169 and re-distributed during deglaciation.

170 The extent to which any given slope unit will yield sediments which are
171 then incorporated into the down-valley transfer of sediments, or sediment
172 'cascade', will depend on multiple factors, including, but not restricted to,
173 slope geotechnical properties such as lithology and degree of consolida-
174 tion (e.g. Curry *et al.*, 2009; Lukas *et al.*, 2012), topographic setting (e.g.
175 Barr and Lovell, 2014) and the consequent nature and efficacy of geomor-
176 phological processes (e.g. Curry, 1999; Curry *et al.*, 2006). These factors
177 will clearly vary in space and time (Ballantyne, 2002b, Orwin and Smart
178 2004a) with the result that the detailed sedimentological consequences of
179 paraglacial activity are uncertain (Curry and Ballantyne, 1999), with rela-

180 tively few studies to date directed towards identifying a sedimentological
181 signature associated with paraglacial reworking (e.g. Benn and Ballantyne,
182 2005; Curry *et al.*, 2009). This uncertainty is in no small part due to the fact
183 that a given slope unit can act as both a store and a source of sediment,
184 with secondary paraglacial activity furthering the complexity of the land-
185 scape response. Lateral moraines represent an interesting example of this
186 dual 'sink' and 'source' role. Being commonly located away from highly dy-
187 namic proglacial fluvial systems, lateral moraines are more usually dissect-
188 ed, reworked and modified on an episodic basis through the operation of
189 debris flow activity, facilitated by, for example, extreme precipitation
190 events, spring thaw (e.g. Blair, 1994; Kellerer-Pirklbauer *et al.*, 2010) and
191 thermo-erosion of dead ice bodies (e.g. Kjær & Krüger 2001; Schomaker &
192 Kjær, 2008; Lukas *et al.*, 2012).

193

194 **Lateral slope stability**

195 The extent to which any slope unit releases sediments for reworking into
196 the down-valley sediment cascade will relate in part to the overall geo-
197 morphic stability of that unit. Lateral moraines may be particularly im-
198 portant in this respect, due to their ability in many cases to stand stably at
199 extreme angles. Many lateral moraines exhibit very steep ($>60^\circ$) proximal
200 slopes that appear to retain stable form, despite ongoing paraglacial re-
201 working (e.g. Curry *et al.*, 2005; Curry *et al.*, 2009) and are indicative of the
202 storage of glaciogenic sediments in a quasi-stable state. Where moraines
203 exist in such a steep, stable form, any role as a source of sediment is likely
204 to become diminished. Slope stability, or lack thereof, then becomes an

205 important determinant of the delivery and storage of sediments within a
206 deglaciating environment, as enhanced paraglacial activity and associated
207 sediment delivery from current or former ice-marginal areas is not an inev-
208 itable and immediate consequence of deglaciation where factors such as
209 time, climatic setting and geotechnical properties may well induce long-
210 term slope stability. The extent to which the stability of a given slope unit
211 is compromised, such that the unit might act as a source of sediment, will
212 be dictated by multiple factors. These may include processes such as de-
213 buttressing and resultant gravitational deformation (e.g. Hugenholtz *et al.*,
214 2008), the action of fluvial processes, which may also be exacerbated by
215 extreme events such as glacier lake drainage (e.g. Iturrizaga, 2008), ante-
216 cedent saturation and soil suction levels (e.g. Springman *et al.*, 2003; Hür-
217 limann *et al.*, 2012) and complex combinations and interactions of geo-
218 morphic processes, such as freeze-thaw, snowmelt, water infiltration and
219 consequent sub-surface flow, seepage and outflow (Hürlimann *et al.*,
220 2012). Clearly therefore, detailed genesis and long term stability of mo-
221 raines will be dictated by multiple elements, with glacial conditioning and
222 local geomorphological factors being of particular importance (Hugenholtz
223 *et al.*, 2008).

224

225 Given the likely complex genesis of lateral moraines, combined with rela-
226 tive uncertainty over their stability, preservation and reworking potential
227 and formative mechanisms (Lukas *et al.*, 2012), they present something of
228 a conundrum, having the potential to act as both a dynamic sediment
229 source where paraglacial modification and reworking is taking place (e.g.

230 Irvine-Fynn *et al.*, 2011) and a relative sediment sink where stability is evi-
231 dent (e.g. Otto *et al.*, 2009, Blair, 1994).

232 The material properties and genetic mechanisms that permit some lateral
233 moraines to stand stably at angles, that in some cases exceed 70° (e.g.
234 Whalley, 1975; Lebourg *et al.*, 2004; Curry *et al.*, 2005; Lukas *et al.*, 2012),
235 remain incompletely understood, although recent work conducted in the
236 European Alps has highlighted the importance of processes such as over-
237 consolidation (e.g. Lukas *et al.*, 2012) and the steepness of underlying bed-
238 rock (Lukas, pers. comm.). This characteristic of many lateral moraines is
239 important in the context of sedimentary deglaciation dynamics as, irre-
240 spective of the source of sediment (subglacial, englacial, glaciofluvial, sub-
241 aerial) slope stability necessarily means that sediments are held in transi-
242 ent storage, but are also potentially available for release, re-working and
243 subsequent re-deposition, as evidenced by the presence of gullies and as-
244 sociated slope-foot deposits (Figure 1B and C). This ability of steep lateral
245 moraines to hold sediments in quasi-stable form has implications for the
246 correct interpretation of paraglacial sediment dynamics. Models of parag-
247 lacial system behaviour (e.g. Church and Ryder, 1972; Matthews, 1992) of-
248 ten propose maximum sediment production at or soon after deglaciation,
249 followed by a simple and continuous, uni-directional decline. However,
250 transient storage in and stochastic release of sediments from sources such
251 as quasi-stable lateral moraine sequences has the capacity to disrupt this
252 interpretation and allow release and reworking of sediments for many
253 decades, if not centuries following deglaciation (Ballantyne, 2002b; Curry
254 *et al.*, 2009) with enhanced availability of sediments associated with degla-
255 ciation not always resulting in a consistent increase in sediment release

256 and reworking (e.g. Cossart, 2008). This ‘interrupted sediment cascade’
257 (Curry *et al.*, 2005, Figure 2) continues in an episodic manner, such that
258 paraglacial slope adjustment following retreat and wasting since the Last
259 Glacial Maximum may still impact upon erosional and mass movement
260 processes operating today (e.g. Kellerer-Pirklbauer *et al.*, 2010; Vehling *et*
261 *al.*, 2016).

262 The geotechnical properties of lateral moraine sediments offer one avenue
263 of investigation towards a fuller understanding of lateral moraine stability.
264 Whalley (1975) was able to characterise various mechanical aspects of lat-
265 eral moraines at the Feegletscher, Switzerland and calculated a friction an-
266 gle of 45°, while suggesting that soil suction may have a role to play in en-
267 hancing slope strength and therefore stability. Distally dipping fabric,
268 evident with clastic material > 0.5m long, results in imbrication of the
269 Feegletscher moraines and might be expected to enhance stability as a re-
270 sult. However, the presence of steep, stable moraines with proximally dip-
271 ping fabric observed elsewhere suggests that a distal dipping fabric is not a
272 requirement for stability and extreme slope angles (Whalley, 1975). *In situ*
273 studies carried out at the same location, combined with geotechnical la-
274 boratory tests of extracted samples from upper high-angle (< 80°) slope
275 units, indicate that imbricated and proximally dipping mica-schist clasts in-
276 hibit shallow translational shear on the proximal slopes, which may assist
277 in the retention of stable form at this site (Curry *et al.*, 2009, Figure 1C). A
278 ‘buttressing’ effect provided by the inter-gulley slopes is also speculated to
279 assist in slope stabilisation, with suction effects potentially enhancing sur-
280 face crust formation, but with limited impact on long-term stability (Curry
281 *et al.*, 2009).

282

283 Figure 2. Paraglacial models illustrating catchment sediment responses to
 284 deglaciation. The solid black curve illustrates the classic model proposed by
 285 Church and Ryder (1972), whereby basin sediment yields progressively de-
 286 cline with time since deglaciation. The dashed grey curves illustrate re-
 287 sponses to deglaciation conditioned by sediment availability/exhaustion as
 288 proposed by Ballantyne (2002b), with episodic events taking place during
 289 and after deglaciation conditioned by, for example, extreme rainfall events
 290 (adapted from Ballantyne, 2002a).

291 A comprehensive sedimentological study of lateral moraine morphology
 292 and properties carried out at Findelengletscher, Switzerland (Lukas *et al.*,
 293 2012) provides further information on lateral moraine genesis and result-
 294 ant factors that may influence slope stability and resultant potential for
 295 sediment release and reworking. Rather than a dominant supraglacial
 296 source for lateral moraine sediments, subglacial and glaciofluvial sedi-

297 ments dominate, deposited by debris flows once material has been trans-
298 ferred from the bed to the ice surface via englacial pathways. Key however
299 to proximal slope stability is the process of overconsolidation. Lukas *et al.*
300 argue that this overconsolidation arises through a combination of glacio-
301 tectonisation of pre-existing sediments and incremental ‘plastering’ of till
302 onto the proximal slopes of lateral moraines by moving ice. The resulting
303 overconsolidation is suggested to be an important factor in enhancing
304 slope stability, retarding paraglacial slope modification and consequently
305 enhancing preservation potential and is regarded as a potentially wide-
306 spread process that may offer an explanation for slope stability and the ex-
307 treme slope angles often in excess of 80° as observed at Findeleng-
308 letscher, without having to invoke the presence of a distally dipping fabric,
309 a feature not ubiquitous to all lateral moraines (Curry *et al.*, 2009; Lukas *et*
310 *al.*, 2012).

311 The stability and steepness of proximal lateral moraine slopes often gives
312 rise to cross-sectional asymmetry, with distal slopes exhibiting lower slope
313 angles, although it should be noted that cross-sectional asymmetry is not a
314 ubiquitous observation (Lukas and Sass, 2011). In contrast to steep proxi-
315 mal slopes, lower-angled distal slopes, that usually represent depositional
316 fan surfaces resting at the angle of repose, are generally less prone to in-
317 tense paraglacial modification and the formation of gullies and, as such,
318 tend to become more readily stabilised by vegetation cover (Figure 3).
319 During deglaciation therefore, paraglacial modification of slope form may
320 act to release sediments for reworking, while vegetation colonisation acts
321 to decrease geomorphological activity and stabilise slope form where con-
322 ditions are conducive to that colonisation (Matthews, 1992; Ballantyne,

2002a; Eichel *et al.*, 2015). Conversely, at sites in the Norway and the central Swiss Alps, Curry (1999) and Curry *et al.* (2005) suggest that progressive vegetation colonisation on geomorphologically-active proximal moraine slopes is thought to be a response to slope stabilisation, rather than a causal factor. The role of vegetation acting to stabilise, or arising as a result of stabilisation and enhancing longer-term stability, therefore adds further complexity to the functioning of lateral moraines as both stores and sources of sediment. Feedback between vegetation and geomorphic process is likely to strongly condition slope form, but the details of this feedback are incompletely understood (Eichel *et al.*, 2015; chapter 7.2) and represent an important area for further research, as deglaciation and vegetation colonisation gather pace in many locations.

Reworking of lateral slopes

Where paraglacial reworking does take place on lateral moraines, it frequently gives rise to dissection of moraine slopes, formation of gullies and redistribution of sedimentary materials, with associated colluvial debris cones and fans accumulating at the slope foot (e.g. Curry and Ballantyne, 1999; Curry *et al.*, 2005; Curry *et al.*; 2009; Lukas *et al.*, 2012, Figure 1B and 1C). Clearly this dissection of moraines provides evidence of mass movement and resultant redistribution and transfer of sediment to the glacier surface or, where glacier retreat has left the moraine unit wholly exposed, the proglacial zone. Despite their apparent large-scale overall stability, steep lateral moraines are therefore potentially important conditioning factors for paraglacial slope modification (Curry *et al.*, 2009) and represent a dynamic environment that may provide an efficient transport link be-

348 tween hillslopes and other geomorphic systems such as channelised flow
349 in forefield areas (Cavalli *et al.*, 2013; Eichel *et al.*, 2015). However, in
350 common with studies of lateral moraine formation and internal structure,
351 paraglacial modification of slope form generally (Curry, 1999; Curry, 2000)
352 and lateral moraine form in particular (Lukas *et al.*, 2012) has received rel-
353 atively scant research attention to date.

354 Paraglacial modification of lateral moraine slope form can take place
355 through a variety of processes, including debris flowage, debris slides,
356 stream action, solifluction, snow avalanching and stream action (Ballan-
357 tyne, 2002a; Curry *et al.*, 2006; chapter 4.4). Such processes are largely re-
358 sponsible for the characteristic paraglacial landscapes of gullied lateral mo-
359 raines and valley sides, slope foot debris cones, fans and valley-floor
360 deposits (Ballantyne and Benn, 1994; Ballantyne, 1995, Curry, 1999; Curry
361 *et al.*, 2006, Figure 1A). It is clear that extensive paraglacial modification of
362 slope form through, for example, gullyng, can take place within short
363 timescales of the order of a few decades (e.g. Ballantyne and Benn, 1994)
364 and that debris flow activity is one of the prime agents responsible for the
365 redistribution of glacial sediments contained within hillslopes or moraines.
366 Obvious flow tracks, levées and debris cones are often visible (e.g. Eyles
367 and Kocsis, 1988; Owen, 1991; Ballantyne and Benn, 1994, Curry 2000),
368 and rainfall (e.g. Chiarle *et al.*, 2007) and rapid snowmelt and resultant liq-
369 uefaction (e.g. Ballantyne and Benn, 1994; Palacios *et al.*, 1999; Curry,
370 2000) are key triggers in many debris flow events. However, the precise
371 controls on the extent and efficacy of debris flowage and other paraglacial
372 hillslope activity represent a relatively under-studied aspect of deglaciation
373 terrain relaxation (Curry, 2000). Studying paraglacial modification of glaci-

374 ogenic slope sediments in western Norway, Curry (2000) utilises gully den-
375 sity as a surrogate indicator of paraglacial modification of sediments within
376 lateral moraines. In common with other studies of paraglacial slope modi-
377 fication (e.g. Curry *et al.*, 2005; Curry *et al.*, 2009; Cavalli *et al.*, 2013), de-
378bris flow activity emerges as a dominant mechanism, with snow ava-
379 lanches representing a secondary paraglacial process. Of several intrinsic
380 and extrinsic conditioning factors studied that may influence paraglacial
381 debris flow activity, slope gradient, sediment availability and water supply
382 emerge as important factors. However, a relative lack of paraglacial slope
383 modification even in high relief areas demonstrates a clear need for a
384 fuller understanding of the detailed constraints on paraglacial activity (Cur-
385 ry, 2000).

386 One obvious constraint is sediment supply and this aspect of paraglacial
387 activity warrants a re-evaluation of the original Church and Ryder model
388 that emphasises time as the key variable dictating paraglacial sediment
389 supply, release and reworking, with maximum sediment supply occurring
390 shortly after deglaciation and then steadily declining as glacier shrinkage
391 continues (Church and Ryder, 1972). Ballantyne (2002b) therefore sug-
392 gests that an exponentially-declining exhaustion model might be more ap-
393 propriate, with sediment yield being dictated by sediment availability and
394 perturbations in yield being dictated by factors such as changing base level
395 or episodic sediment release from, for example, hillslopes during extreme
396 rainfall events (Figure 2, Ballantyne, 2002b). This concept of sediment
397 availability as a key controlling factor has significance when considering
398 the role that hillslopes may play as stores and sources of sediment in a de-
399 glaciating landscape. Indeed, Curry (1999) suggests that upslope sediment

400 availability is a likely control on slope stabilisation and to a certain extent,
401 the delivery of sediments from ice-marginal hillslopes itself ensures a con-
402 tinuously diminishing supply of sediment and longer term stability, as de-
403 bris flowage delivers material to the slope foot, progressively burying up-
404 per slope units, reducing slope gradient, reducing upslope sediment supply
405 and potentially facilitating slope stability, with vegetation colonisation fur-
406 ther stability (Figure 1B).

407 Considering lateral moraine materials further up-valley, the supraglacial
408 accumulation of sediments sourced from adjacent hillslopes clearly repre-
409 sents an important input to lateral moraines. Although there is an exten-
410 sive literature on debris-covered glaciers (see also section 3.3), the extent
411 to which ice-marginal sediment delivery to a glacier surface may be re-
412 garded as strictly paraglacial in nature is open to question. Although we do
413 not consider rockfall delivery in this section, it is thought that enhanced
414 rockfall activity is a likely consequence of deglaciation (e.g. Fischer *et al.*,
415 2006; Stoffel *et al.*, 2014; chapters 4.1 and 4.2) and it seems not unrea-
416 sonable to assume that ice-marginal sediment delivery to a glacier surface
417 is also likely to increase as deglaciation continues, through processes of
418 debuttreasing (in instances where sedimentary slope units are being sup-
419 ported by the glacier), melt of permafrost that is acting to stabilise sedi-
420 mentary units and through a general increase in water supply, associated
421 with ice melt and shifting in the relative proportions of precipitation from
422 snow to rainfall. Potential links between deglaciation and delivery of ice
423 marginal sediments to the glacier surface have been demonstrated in Arc-
424 tic settings (e.g. Porter *et al.*, 2010) and there is a clear need for further re-
425 search to assess the links between slope sedimentary systems and suprag-

426 lacial systems in alpine settings, particularly considering the impact that
 427 surface debris cover can have on glacier melt, behaviour and ultimately
 428 moraine formation (e.g. Reznichenko *et al.*, 2011). It is also becoming in-
 429 creasingly clear that a genetic origin for lateral moraines based wholly up-
 430 on supraglacial accumulation of sediments derived from valley-side slopes
 431 may not appropriately explain formative mechanisms. The presence of
 432 sub- and en-glacial sediments and glaciofluvial deposits within lateral mo-
 433 raines highlights the potential importance of alternative mechanisms for
 434 supraglacial sediment accumulation and subsequent reworking through
 435 processes such as thermo-erosion and debris flow activity (e.g. Etzelmüller,
 436 2000; Irvine-Fynn *et al.*, 2011; Lukas *et al.*, 2012; Porter *et al.*, 2010; Tonkin
 437 *et al.*, 2017).

438

439 FOREFIELD SLOPES

440 Formation and fluxes

441 Although often less visually obvious than large lateral moraine and valley
 442 side glaciogenic sediment sequences, the often extensive suites of mo-
 443 raine landforms commonly observed in deglaciating forefields represent a
 444 dynamic source and store of glacial sediments and exhibit characteristics
 445 of primary and secondary paraglacial activity. Although forefield slope gra-
 446 dients are typically lower than those observed on lateral moraine slopes,
 447 the presence of often spatially extensive and high-energy fluvial activity in
 448 the proglacial area (chapters 5.1 and 5.2) offers an important means of
 449 sediment transport and re-distribution, making the forefield area the most

450 dynamic part of the alpine sediment flux system (Fenn and Gurnell, 1987;
451 Warburton, 1990; Maizels, 1993; Marren, 2005; Otto *et al.*, 2009). The
452 presence of ice cores within proglacial moraine structures adds additional
453 interest, as not only do ice cores potentially influence sub-surface drainage
454 pathways with potential slope stability implications (e.g. Langston *et al.*,
455 2011; Muir *et al.*, 2011), but thermo-erosion offers an additional mecha-
456 nism of sediment release from moraine slope units, aided by associated
457 meltwater supply. There is extensive existing literature detailing mecha-
458 nisms and processes of glaciofluvial entrainment, transport and deposition
459 within proglacial environments and useful overviews can be found in Gur-
460 nell and Clarke (1987), Warburton (1990), Marren (2005) and Lane *et al.*,
461 (2017). Likewise, the genesis, morphology and structure of ice-marginal
462 moraines is an area that has received much research attention and we
463 would direct readers to Bennett (2001) and Barr and Lovell (2014) as use-
464 ful starting points for discussion of these aspects of glacial geomorphology.
465 As with the previous section concerning lateral slope forms and processes,
466 in this section we confine our consideration to discussion of the role of
467 moraine proglacial landforms as potential stores or sources of sediment
468 and consider the mechanisms and extent of paraglacial activity operating
469 on moraine slope units to further the redistribution of glaciogenic sedi-
470 ments in deglaciating environments.

471 Glacier forefields are necessarily comprised of deposits that reflect multi-
472 ple episodes of erosion, transport and deposition, having inevitably been
473 subject to the operation of multiple geomorphological processes and can
474 be viewed as transitioning from glacial to non-glacial conditions (Heck-
475 mann *et al.*, 2013). The diversity and morphological characteristics of fore-

476 field deposits will broadly reflect processes of sediment production and
477 modification beneath the glacier, forefield geomorphological activity and
478 glacier dynamics and in particular, the glacial response to deglaciation and
479 the extent of any reworking and redistribution by meltwater (Orwin &
480 Smart, 2004a). Exposure through glacier retreat and thinning makes de-
481 posits potentially available for re-mobilisation, transport and deposition by
482 non-glacial processes, the presence of flowing water in the form of pro-
483 glacial meltwaters being a particularly effective agent of erosion in a geo-
484 morphological context of abundant sediment supply, with observations in-
485 dicating that up to 80% of total suspended sediment yield in proglacial
486 rivers may be sourced from the forefield area (Orwin and Smart, 2004b).
487 However, given the likely complexity of the genetic history of forefield sed-
488 iments, they will comprise a wide range of deposits of varying ages and
489 therefore varying stability and vulnerability to reworking (Orwin & Smart,
490 2004a). For example, negative feedback in some areas of deglaciating ba-
491 sins (e.g. transient storage of sediment in slope-foot debris cones) can act
492 to reduce, rather than increase, basin sediment yields, through a decline in
493 sediment system connectivity, despite increases in that connectivity else-
494 where in the basin in response to deglaciation (e.g. Lane *et al.*, 2017). In-
495 deed, the extent to which the recently deglaciating forefield might repre-
496 sent a source or a sink for sediments is open to conjecture. Observed in an
497 Arctic setting, Hodgkins *et al.* (2013) found that the extent to which the
498 forefield zone may act as a net sink or a source of sediments may depend
499 upon factors such as the nature of the temporally variant runoff regime,
500 while age since exposure also likely represents an important factor in dic-
501 tating vulnerability to reworking (e.g. Orwin and Smart 2004a), reflecting

502 the generalised form of the paraglacial model of declining sediment yield
503 as time since deglaciation advances (e.g. Church and Ryder, 1972, Figure
504 2).

505 At the basin scale, it is common for the forefield to be dominated by a rel-
506 atively flat glacio-fluvial plain that extends from the glacier margin (Mar-
507 ren, 2005). In alpine environments, the forefield environment is typically
508 delineated laterally by steep valley-side slopes and potentially high and
509 steep lateral moraines, while the frontal margin is often marked by a
510 prominent moraine (e.g. Figure 3), invariably breached by flowing meltwa-
511 ters. This frontal moraine is often the most morphologically distinctive
512 forefield moraine form (Benediktsson *et al.*, 2009) and in alpine environ-
513 ments terminal or end moraines are usually associated with maximum re-
514 cent glacier advance during the Little Ice Age (Figure 3A). Dependent on
515 age, spatial extent and intensity of geomorphological activity, end mo-
516 raines may support vegetation cover, particularly on distal slopes (Figure
517 3A), but equally, exposed sediments will be vulnerable to processes of re-
518 working and redistribution. The stability of these moraines is a topic that
519 has attracted much recent interest in the light of deglaciation and result-
520 ant formation of proglacial water bodies dammed behind terminal or end
521 moraines with consequent risk of glacier lake outburst floods (GLOFs)
522 should the moraine dam be breached (e.g. Stoffel and Huggel, 2012; Wes-
523 toby *et al.*, 2014; Ashraf *et al.*, 2015. See also section 5.3).

524

525 Figure 3. A. *Terminal moraine of the Feegletscher Nord, Valais, Switzerland.*526 A. *Dashed line indicates approximate location of the glacier front during*527 *the Little Ice Age maximum. Note the heavily vegetated distal slopes of the*

528 *terminal moraine. B. View towards the terminal moraine from the promixal*
 529 *side. Note the mature vegetation on the stable proximal slopes of the ter-*
 530 *minal moraine to the far left of the image and the potential for slope foot*
 531 *fluvial transport where the proglacial meltwater stream impinges on the*
 532 *moraine in the middle distance.*

533 The sediments stored within the forefield area are typically characterised
 534 by substantial volumes of the products of glacier erosion, and are usually
 535 dissected by meltwater streams with resultant capacity for extensive
 536 glacio-fluvial reworking and redistribution (e.g. Warburton, 1990; Hodgkins
 537 *et al.*, 2003; Orwin & Smart, 2004a; Leggat *et al.*, 2015). It is not uncom-
 538 mon for deglaciating forefields to be further characterised by a diverse
 539 range of glacial, fluvial and non-glacial landforms, features and deposits.
 540 Setting aside the glacio-fluvial landforms, stores and processes that are
 541 considered elsewhere in this volume (chapter 5.1), it is the diverse groups
 542 of moraines that are particularly important when considering paraglacial
 543 activity operating within the forefield area, although clearly relief and
 544 slope angles are both likely to be lower on average than that observed in
 545 recently deglaciaded lateral locations.

546 Forefield moraines are an important source of information about the ex-
 547 tent and dynamics of glaciers and ice masses. Consequently, they have
 548 been extensively studied and, despite the fact that mechanisms of for-
 549 mation remain manifold (Hiemstra *et al.*, 2015), they have been widely
 550 used as an indirect proxy for climatic variation and past glacier dynamics
 551 and extents (e.g. Benn & Ballantyne 2005; Beedle *et al.*, 2009; Evans *et al.*,
 552 1999). Numerous moraine types and classifications exist, leading to a

553 complex picture in part due to ‘inconsistent terminology’ (Winkler and
 554 Matthews, 2010, p87) with sometimes ‘incompatible’ interpretations of
 555 the same landform being made (Evans *et al.*, 1999, p673) and confusion
 556 being heightened by restricting moraine analysis to geomorphology alone,
 557 without a detailed consideration of the detailed sedimentology of mo-
 558 raines that is evident in more recent work (e.g. Lukas *et al.*, 2012, Reinardy
 559 *et al.*, 2013; Chandler *et al.*, 2016). Given the dynamic nature of proglacial
 560 environments and associated potential for moraine modification and even
 561 potential eradication by a range of post-depositional processes (Kirkbride
 562 & Brazier, 1998; Kirkbride & Winkler, 2012), preservation potential is vari-
 563 able, as will be the role that moraines play in the storage and supply of
 564 sediment. The extent to which moraines may provide an accessible source
 565 of sediment for reworking and redistribution will depend upon myriad fac-
 566 tors including, but not limited to: time since exposure, slope angle, degree
 567 of glaciotectionisation and compaction, the presence of an ice core and as-
 568 sociated de-icing and thermo-erosion, the nature and extent of vegetation
 569 colonisation and the degree of transport and supply/weathering limita-
 570 tions.

571 REWORKING OF FOREFIELD SLOPES

572 Buried ice

573 The presence and subsequent prolonged degradation of buried remnant
 574 glacier ice within moraine landforms may influence moraine formation,
 575 morphology, slope activity and flow of meltwaters and groundwater in the
 576 forefield area (e.g. Kjær and Krüger, 2001; Langston *et al.*, 2011; chapter
 577 3.6). Retreating debris-covered glaciers can leave forefields of ice-cored

578 forms and ice-rich debris (e.g. Bosson *et al.*, 2015) and this is particularly
 579 true close to the ice-margins, where dead ice can be incorporated during
 580 the process of moraine formation (Lukas *et al.*, 2012). Ice-cores may sur-
 581 vive decades to millennia and at considerable distances from the active ice
 582 margins (Barr & Lovell, 2014) and their degradation is typically evident as
 583 backwasting (i.e. lateral retreat) and downwasting (i.e. thinning, Krüger &
 584 Kjær, 2000), commonly giving rise to inverted topography and the rework-
 585 ing of sediments due to slumping, fall-sorting and the formation of fea-
 586 tures such as sinkholes, extension fractures, kettles, cracks, slips and
 587 mud/debris flows (Johnson, 1971; Kjær and Krüger, 2001). The redistribu-
 588 tion and remobilisation of sediments can also be facilitated further by ice
 589 core melt and consequent sediment mobilisation, where removal of slope
 590 sediments through, for example, slumping or other slope mass movement
 591 exposes ice which is then vulnerable to thermo-erosion and backwasting
 592 (e.g. Kjær and Krüger, 2001; Schomacker & Kjaer 2007). Similarly, slow
 593 melt of buried ice can initiate sinkhole formation and associated collapse
 594 of slope units through undermining, and over-steepening of adjacent
 595 slopes that may initiate processes such as sliding and backslumping (Kjær
 596 and Krüger, 2001).

597 These processes of slope readjustment in response to a melting ice core
 598 and consequent sediment redistribution, may be enhanced through fluvial
 599 induced thermo-erosion where meltwaters are present. It is well estab-
 600 lished that fluvial activity plays an important role in shaping the detailed
 601 morphology of forefield landscapes (e.g. Carrivick *et al.*, 2013), primarily
 602 through the mechanical processes of entrainment, transport and deposi-
 603 tion and where flowing water is able to remove surficial deposits and ex-

604 pose buried ice, melt will progress rapidly. However, the presence of bur-
605 ied ice may also modify and partially dictate the location and efficacy of
606 sub-surface water flow paths and it has been established that moraines
607 can contain complex subsurface hydrological systems associated with the
608 presence and interaction of sediments, buried and ground ice and bedrock
609 (Langston *et al.*, 2011).

610 Buried ice can act in two key ways to modify sub-surface drainage. Firstly,
611 ice can act as an aquiclude, presenting a relatively impermeable barrier to
612 sub-surface waters, controlling the routing of water. Secondly, buried ice
613 may act as a source of water through the ongoing process of slow, sub-
614 surface melting, or through rapid melting where ice becomes exposed at
615 the surface. The accumulation of water in response to any sub-surface ice
616 melt also offers the potential for saturation of sediments, with resultant
617 implications for slope stability should that saturation destabilise slope
618 units or indeed penetrate to surface sediment horizons. In Arctic settings,
619 collapse of moraine sediments through melt of buried ice provides a po-
620 tentially important component of the sediment cascade as a means of re-
621 leasing sediments for redistribution, with resultant impact on moraine
622 slope form and basin sediment yield (e.g. Etzelmüller, 2000; Etzelmüller *et*
623 *al.*, 2000; Lyså, and Lønne, 2001; Lukas *et al.*, 2005; Porter *et al.*, 2010; Ir-
624 vine-Fynn *et al.*, 2011). Where subsurface ice and water are in contact
625 there is also the potential for effective thermo-erosion. However, largely
626 due to a paucity of field observations and the practical difficulties of as-
627 sessing both sub-surface water flow and ice presence and characteristics,
628 the detailed role of forefield moraine landforms in controlling subsurface
629 water storage and flow remains poorly constrained (Langston *et al.*, 2011)

630 and by implication, so do the resultant influences upon surface slope pro-
631 cesses and stability.

632 **Hydrological interactions**

633 It is well established that sediment transfer within, and from, deglaciated
634 forefields reflect the competing controls of sediment supply and the effi-
635 cacy of transport processes (e.g. Warburton, 1990; Hodgkins *et al.*, 2003).
636 Networks of high-energy streams and rivers typically dominate the down-
637 stream transfer of glacial sediments from forefield environments and
638 fluvial processes are one of the primary agents of within-catchment land-
639 scape change due to their pivotal role in forefield sediment reworking and
640 redistribution (Warburton, 1990; Marren, 2005; Carrivick *et al.*, 2013; Lane
641 *et al.*, 2017).

642 The impact of fluvial-slope interactions has its most obvious morphological
643 expression where forefield moraines have been dissected by meltwater
644 channels. Such dissection will arise through ‘normal’ processes of glacio-
645 fluvial erosion, but more extreme examples of dissection and associated
646 slope modification have been associated with outburst floods (e.g. Staines
647 *et al.*, 2014) or the failure of moraine dams in mountain environments (e.g.
648 Korup & Tweed, 2007). However, there are limited quantitative data avail-
649 able to establish the overall importance of such large-scale sediment redis-
650 tribution events in re-mobilising and redistributing sediments as deglacia-
651 tion terrain relaxation takes place, nor to establish the likely recurrence of
652 such events (Ballantyne, 2002a; Beylich & Warburton, 2007). More com-
653 monly, there will be ‘normal’ or ‘continuous’ post-depositional modifica-
654 tion of forefield landforms through the action of flowing meltwater, inter-

655 persed with discrete, extreme events (e.g. Cossart, *et al.*, 2008; Carrivick
656 *et al.*, 2013).

657 The impact of connections between slope and glaciofluvial sedimentary
658 systems and resultant landform modification are most evident at the mar-
659 gins of proglacial streams, where fluvial processes may directly interact
660 with hillslope units, presenting the most obvious sign that sediment is be-
661 ing delivered from the slopes to the valley floor (Beylich & Warburton,
662 2007; Cavalli *et al.*, 2013, Figure 3B). Any resulting erosion not only modi-
663 fies slope form, but may potentially generate slope instability through pro-
664 cesses of undercutting and resultant collapse (Iturrizaga, 2008; Winkler &
665 Matthews 2010). Direct proglacial channel margin erosion will be limited
666 to the valley floor, however, upslope instability may be induced and only
667 become apparent when slopes are destabilised and mass movements oc-
668 cur (Orwin & Smart, 2004; Schrott *et al.*, 2006).

669 It is clear that fluvial activity plays an important role in moraine morpho-
670 logical modification and associated release of sediments and that forefield
671 sediments generally can contribute significantly to total basin sediment
672 yield (e.g. Warburton, 1990; Orwin & Smart, 2004a; Leggat *et al.*, 2015)
673 and in an episodic fashion in the case of mass movements (Schrott *et al.*,
674 2006; Carrivick *et al.*, 2013). However, over time the hydrological efficacy
675 of flowing water in mobilising forefield sediments is likely to decline. This
676 arises from the eventual decline of the 'deglaciation dividend' (Kaser *et al.*,
677 2010), whereby the initial enhanced melt, or dividend, associated with de-
678 glaciation reduces. Aside from this dwindling supply of meltwaters in the
679 latter stages of deglaciation, surface sediments and slopes are likely to

680 stabilise because of eluviation of fines and progressive armouring of sur-
681 face layers caused by processes such as overland flow and rainsplash. This
682 process can take place rapidly and forefield surfaces may cease to function
683 as significant sources of sediment within decades of deglaciation and ex-
684 hibit stability to all but extreme events (Orwin and Smart 2004a).

685 The reworking and redistribution by meltwater of sediments within mo-
686 raines may eventually lead to a situation where much of the forefield will
687 comprise reworked glaciogenic material (Carrivick *et al.*, 2013) such that
688 relief becomes subdued and the area of meltwater river channels increas-
689 es (Staines *et al.*, 2014). Forefield slope stability will likely be restricted to
690 areas unaffected by fluvial activity or ice-core degradation (Ewertowski *et*
691 *al.*, 2010). This increase in slope stability and consequent decrease in sed-
692 iment availability and mobilisation with time has been evident from studies
693 of the patterns of aggradation and degradation in glacier forefields, where
694 geomorphological activity decreases in both spatial extent and in intensity
695 with distance from the glacier and by implication, time since deglaciation
696 (e.g. Orwin and Smart, 2004a). Hillslope activity will then likely become re-
697 stricted to inherently unstable slopes, or those unaffected by or resistant
698 to fluvial activity. As deglaciation continues, fluvial activity in general will
699 progressively become more important. However, the apparent stability
700 and persistence of moraine landforms and sediments observed in contem-
701 porary forefield areas suggests that the reworking and redistribution of
702 deposits by fluvial activity may not be spatially extensive and that geomor-
703 phological activity is increasingly dominated by the interactions with spe-
704 cific processes (e.g. fluvial) over relatively small areas of catchments (Car-
705 rivick *et al.*, 2013) or by the occurrence of extreme high-magnitude,

706 episodic, low-frequency events such as those induced by changes in tem-
707 perature or precipitation (Stoffel and Huggel, 2012; Blair, 1994).

708 The importance of episodic rainfall events in mobilising sediments from
709 forefield slopes has been noted in a number of studies (e.g. Richards,
710 1984; Kellerer-Pirkbauer *et al.*, 2010; Cavalli *et al.*, 2013; Leggat *et al.*,
711 2015; chapter 4.4) and can be responsible for both destabilising slopes
712 (Blair, 1994; Deline *et al.*, 2015) and contributing to the exhaustion of sup-
713 plies of fine sediments, with resultant reduction in overall sediment mobi-
714 lization (Orwin & Smart, 2004b). The role of precipitation as a sediment
715 transfer mechanism may become increasingly important during deglacia-
716 tion due to the increasing proportion of total runoff comprising liquid pre-
717 cipitation as stores of snow and ice deplete (Collins, 2008), while rainfall-
718 induced extreme discharge events clearly have the capacity to mobilise
719 even well-armoured and stabilised forefield sediments (e.g. Luckmann,
720 1981), interrupting the theoretical uni-directional decline in basin sedi-
721 ment yield as deglaciation progresses. Given forecast changes in precipita-
722 tion frequency and intensity associated with climate change (Stoffel and
723 Huggel, 2012), the potential for precipitation events to further affect basin
724 sediment yields is only likely to increase.

725 ENVIRONMENTAL CHANGE AND ALPINE PARAGLACIAL ACTIVITY

726 As the impacts of climate change are becoming more obvious in alpine re-
727 gions, an increased interest in interactions and feedbacks between parag-
728 lacial activity and associated biogeographical and hydrological processes
729 has developed. In the European Alps for example, larger magnitude debris
730 flow events may become increasingly common due to predicted increases

731 in autumn and spring rainfall, permafrost melt and enhanced sediment de-
 732 livery (Stoffel *et al.*, 2014). It is also hypothesised that, given the combina-
 733 tion of enhanced sediment availability, permafrost degradation and
 734 changes in rainfall, debris flow events with ‘little or no historical prece-
 735 dent’ could be facilitated (Stoffel and Huggel, 2012, p430).

736 To date, catchment-scale studies of paraglacial activity have largely fo-
 737 cussed upon the storage, release and reworking of sediments. However, as
 738 climate warms, there is also a need to consider the related issues of en-
 739 hanced biological and hydrological activity that develop in tandem with
 740 enhanced storage and transfer of sedimentary materials. It is becoming in-
 741 creasingly clear that any consideration of the landscape response to degla-
 742 ciation cannot be considered in isolation from the closely-associated hy-
 743 drological and biological responses. In addition to enhancing the
 744 vulnerability of sediments to processes of reworking and redistribution,
 745 glacier retreat and down-wasting makes those same sediments available
 746 for microbial colonisation, with resultant build up of nutrient pools that
 747 encourage pedogenesis and increased organic matter development (e.g.
 748 Bernasconi *et al.*, 2008; Schurig *et al.*, 2013). In response to increased soil,
 749 organic matter and nutrient availability, vegetation succession is initialised,
 750 leading ultimately to the growth of higher order plants that may play an
 751 important role in conditioning slope stability (Bradley *et al.*, 2014; Eichel *et*
 752 *al.*, 2015; chapter 7.2). Indeed, vegetation-landform interactions are be-
 753 coming increasingly recognised as an important, albeit under-studied, as-
 754 pect of the landscape response to deglaciation, with the result that the
 755 discipline of ‘biogeomorphology’ is currently seen to be an actively devel-
 756 oping and useful means of enhancing understanding of forefield dynamics

757 (Eichel *et al.*, 2013; Eichel *et al.*, 2015). Rapid deglaciation and revelation
758 of ice-marginal and forefield sediments also offers opportunities to study
759 in detail the mechanisms of and factors determining the efficacy of vegeta-
760 tion colonisation and succession (e.g. Jumpponen *et al.*, 1999).

761 It has been established that the extent and stage of vegetation colonisa-
762 tion is a likely strong determinant of sediment availability and catchment
763 sediment yield (Ballantyne, 2002a; Klaar *et al.*, 2015) and, in general terms,
764 vegetation succession and associated ecosystem complexity will develop
765 on a similar timescale to the progressively declining sediment yield as-
766 sumed in many paraglacial models (Ballantyne, 2002a; Klær *et al.*, 2015,
767 Figure 2). For slope units that are not subject to regular disturbance from
768 geomorphological activity, vegetation colonisation impacts paraglacial ad-
769 justment through the stabilisation of landforms and a resultant decline in
770 sediment availability as the paraglacial 'period' progresses (Ballantyne,
771 2002a; Orwin and Smart, 2004b; Marston, 2010; Klaar *et al.*, 2015;). As
772 vegetation colonisation progresses, sediment cohesion and shear strength
773 increase and rainfall interception and infiltration become enhanced,
774 thereby reducing surface runoff and likelihood of surface sediment mobili-
775 sation (Klær *et al.*, 2015).

776 However, in much the same way that a simple uni-directional decline in
777 sediment yield may not adequately describe the reality of the landscape
778 response to deglaciation (e.g. Ballantyne 2002b, Figure 2), vegetation suc-
779 cession may also be interrupted by geomorphic processes (Matthews,
780 1992) dependent on magnitude and frequency (Eichel *et al.*, 2013), result-
781 ing in the persistence of younger successional stages of vegetation coloni-

782 sation (e.g. Eichel *et al.*, 2013) with resultant impacts on sediment availa-
783 bility. Feedbacks and relationships between geomorphic systems and veg-
784 etation ecosystems are therefore potentially complex and the details of
785 microbial and subsequent vegetation colonisation and consequent impacts
786 on slope stability and sediment redistribution (or stability-induced lack
787 thereof) are presently obscure (Klaar *et al.*, 2015) and represent a current
788 research priority.

789 In terms of hydrological impacts, glacial and associated sediments and mo-
790 raines play an important role in controlling the timing and quantity of wa-
791 ter release from glaciated catchments (e.g. Langston *et al.*, 2011; Cook *et*
792 *al.*, 2013), with any paraglacial modification potentially altering the nature
793 and functioning of water flow paths, with resultant impacts on the quanti-
794 ty and timing of basin water yields and the ecology of meltwater-fed eco-
795 systems (e.g. Brown *et al.*, 2006; Milner *et al.*, 2009). Forefield surface gla-
796 cial sediments, such as those contained within moraines, can play an
797 important role in temporarily storing and thereby buffering basin meltwa-
798 ter discharge and sustaining baseflow, but the processes of water storage
799 and transfer through moraines and other depositional glacial landforms
800 remain obscure (Langston *et al.*, 2011). Given the likely increasing intensity
801 of sediment redistribution associated with deglaciation, understanding the
802 associated impacts on meltwater flow specifically and basin hydrology
803 generally represents an important area for further research, especially
804 considering the significant reliance on meltwater as a resource in many al-
805 pine regions.

806 **CONCLUSION**

807 As deglaciation in alpine regions continues, research interest in the result-
808 ant impacts on sediment storage, release and reworking is likely to be-
809 come further enhanced, adding to the growth of interest in paraglacial ge-
810 omorphology identified by Ballantyne (2002a). Great uncertainty still
811 exists, however, concerning the timescales over which the geomorpholog-
812 ical response to deglaciation persists (Dadson and Church, 2005) and the
813 subsequent patterns of sediment release from basins during deglaciation.
814 The basic notion of maximal sediment delivery immediately following de-
815 glaciation, followed by a slow, uni-directional decline as slope units and
816 other terrain surfaces stabilise, may be broadly appropriate over large
817 timescales. However, over shorter timescales, it is clear that enhanced de-
818 livery of sediments from slope units is not an inevitable or immediate con-
819 sequence of deglaciation and that a variety of geological, geomorphologi-
820 cal, hydrological and biogeomorphological factors will inevitably add a
821 somewhat stochastic element to patterns of sediment release. Under-
822 standing these patterns of sediment release and indeed spatial and tem-
823 poral variations in sediment storage, represents an important area of cur-
824 rent glaciological research, as contemporary deglaciation offers an
825 unparalleled, albeit ultimately time-limited, opportunity to directly observe
826 the genesis of deglaciation-landforms, their modification and associated
827 sediment fluxes and fluctuations in basin-scale sediment storage. This not
828 only permits a fuller understanding of the complexities of the geomorpho-
829 logical processes during deglaciation, but contributes to enhanced under-
830 standing of glacial depositional landform genesis. Furthermore, with rapid
831 warming evident in alpine regions, melt of permafrost and a predicted
832 greater frequency of extreme precipitation events have the potential to

833 remobilise the substantial glaciogenic sediment stores that are present in
834 many alpine regions. It is therefore critical, both from a geomorphological
835 and human impact point of view, that a fuller and detailed understanding
836 of the processes that drive that remobilisation is gained, such that greater
837 accuracy can be afforded to predictions of landscape development and as-
838 sociated potential impacts on human activity.

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