

1 **LANDFORM TRANSITIONS FROM PRONIVAL RAMPARTS TO**
2 **MORAINES AND ROCK GLACIERS: A CASE STUDY**
3 **FROM THE SMØRBOTN CIRQUE, ROMSDALSALPANE,**
4 **SOUTHERN NORWAY**

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20
21 **Abstract**

22
23 Landform transitions are defined as intermediate forms that represent transient
24 developmental stages between conventional landform types. This study evaluates
25 possible cases of landform transitions from pronival (protalus) ramparts to moraine
26 ridges, and from pronival ramparts to lobate rock glaciers (protalus lobes) at the foot of
27 the headwall of Smørbotn cirque in southern Norway. The five landforms had been
28 previously classified as pronival ramparts. We conclude that only two (Smørbotn 2 and
29 3) are undisputed, active pronival ramparts, which developed under the seasonal-
30 freezing regime of the Holocene. It is inferred that a third (Smørbotn 1) represents the
31 transition to a moraine ridge formed during the ‘Little Ice Age’ of the last few centuries
32 as a semi-permanent snowbed grew into a small temperate glacier. The two others
33 (Smørbotn 7 and 8) appear to be relict embryonic rock glaciers that developed between

34 the Last Glacial Maximum and the Younger Dryas Stadial under a permafrost regime
35 and benefited from enhanced debris supply as a result of rock-slope instability affected
36 by glacier debuitressing and permafrost degradation. Variable landscape settings and
37 distinctive environmental histories contribute to the differences in the morphology of
38 these landforms. We highlight continuing controversies over the modes of formation
39 and diagnostic characteristics of pronival ramparts by positioning them, together with
40 push/dump moraines, ice-cored moraines and rock glaciers, in a conceptual model of
41 the periglacial-glacial landform continuum. The model links snow, ice and debris fluxes
42 under seasonal freezing and/or permafrost climatic regimes to the process thresholds
43 between landform types.

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46 *Key words:* Landform transitions; periglacial and glacial landform continuum; pronival
47 ramparts, rock glaciers and moraines

48

49

50 **Introduction**

51

52 A set of remarkable depositional landforms – debris accumulations forming elongated
53 ridges of various sizes and complexity – are located parallel to the foot of the
54 headwall of the Smørbotn cirque in the Romsdalsalpane mountains of Møre og
55 Romsdal, southern Norway. These landforms were mapped as pronival (formerly
56 termed protalus) ramparts in the 1980s (Carlson et al. 1983) and have since been the
57 subject of detailed studies focusing on their morphology, formative processes and age
58 (Shakesby et al. 1995, 1999; Matthews et al. 2011; Matthews and Wilson 2015).

59

60 Pronival ramparts are ridges, ramps and mounds composed of boulders and
61 finer debris that have accumulated at the downslope margins of semi-permanent
62 snowbeds (Shakesby 1997, 2014). The main process involved in their formation has
63 conventionally been regarded as a simple one, involving rockfall combined with the
64 rolling or sliding of individual clasts across the snowbeds. The previous investigations
65 at Smørbotn and related studies of pronival ramparts elsewhere (Blagborough and
66 Breed 1967; Ballantyne and Kirkbride 1986; Harris 1986; Ono and Watanabe 1986;
67 Ballantyne 1987; Pérez 1988; Hedding et al. 2007, 2010; Wilson 2009; Hedding

68 2011; Margold et al. 2011; Brook and Williams 2013; Hedding and Sumner 2013)
69 have led, however, to the conclusion that other processes, including debris flow, snow
70 avalanche, solifluction, and snow push, may be important at particular sites. Their
71 ridge morphology can be confused in the field, moreover, with such landforms as
72 moraine ridges, rock glaciers and rock-slope failures, which involve an even wider
73 range of processes. These poorly understood landforms are therefore not as simple to
74 identify as may appear at first sight.

75

76 The classification of landforms represents a common problem in
77 geomorphology, where different landforms may have characteristics in common,
78 particular landforms are not necessarily produced by single processes, and the concept
79 of equifinality has been used to account for the formation of similar landforms by
80 different processes (Haines-Young and Petch 1983; Beven 1996; Temme et al. 2011).
81 Berthling (2011) considered the limitations of classification schemes and the concept
82 of ‘natural kinds’ in a wide ranging discussion of the terminology and
83 interrelationships between glaciers, debris-covered glaciers and rock glaciers (see
84 also, Lewin 2016).

85

86 A related problem is whether landform transitions can be recognized, not only
87 in the sense of an *intermediate form* that lies between recognised landform types but
88 also as a *transient stage* in the transformation of one recognised landform type into
89 another in response to environmental change. Transitional landforms are therefore
90 important for palaeoclimatic interpretation, especially at major climatic transitions
91 such as Termination 1 from Weichselian to Holocene.

92

93 Some of these questions have been raised before in relation to pronival
94 ramparts. Ballantyne and Benn (1994) speculated, for example, on the possible
95 existence of ‘protalus moraines’ between protalus ramparts and moraines, while
96 Haeberli (1985), Haeberli et al. (2003), Kääh (2013) and others have suggested that
97 pronival ramparts can be a transient early stage in the development of rock glaciers.
98 Explicit discussions of the nature and implications of landform transitions are,
99 however, more common in relation to glaciers, debris-covered glaciers and rock
100 glaciers (Ackert 1998; Etzelmüller and Hagen 2005; Whalley 2009; Berthling 2011;

101 Lilleøren et al. 2013; Emmer et al. 2015; Janke et al. 2015; Monnier and Kinnard
102 2015; Seppi et al. 2015; Bosson and Lambiel 2016).

103

104 In this paper we continue these debates over pronival ramparts and related
105 deposits with reference to the landforms in Smørbotn cirque. The specific objectives
106 are as follows:

- 107 • To review the Smørbotn landforms while supplementing existing published
108 data with improved geomorphological maps and additional information on
109 morphology, composition, environmental setting and age.
- 110 • To highlight possible causes of landform transitions between pronival
111 ramparts and moraines, and between pronival ramparts and rock glaciers.
- 112 • To propose a conceptual model that links these conventional landform types
113 and their landform transitions to snow, ice and debris fluxes, formative
114 processes and environmental change.

115

116

117 **Smørbotn cirque and its environment**

118

119 Smørbotn is a north-facing, elongated cirque close to Innfjorden in Møre og Romsdal
120 Fylke (county) (Fig. 1). The floor of the cirque rises from ~700 to 800 m a.s.l. towards
121 the steep, 500-600 m bedrock headwall, which culminates in Smørbotntinden (1520 m
122 a.s.l.). Towards the east, the headwall becomes a less steep, colluvial slope rising
123 towards the higher peak of Finnan (1786 m a.s.l.). On both flanks of the cirque,
124 prominent lateral moraines slope gradually northwards and spill over the lip of the
125 cirque. These moraines, mapped by Carlson et al. (1983), are presumed to have been
126 deposited by the local glacier that occupied the cirque in the Younger Dryas Stadial
127 until at least 11,700 cal. BP (Sollid and Sørbel 1979; Nesje et al. 1987, 2009;
128 Rasmussen et al. 2006; Goehring et al. 2008; Winguth et al. 2009; Mangerud et al.
129 2011; Stroeven et al. 2015). Apart from Smørbotn 1, to be discussed later, glaciers are
130 absent from Smørbotn at present but exist on the east-facing slope of Finnan at an
131 altitude of about 1300 to 1700 m a.s.l. (See Andreassen and Winsvold 2012).

132

133 This study focuses on the depositional landforms that are draped around the
134 foot of the headwall at an altitude of ~750-850 m a.s.l. (Fig. 2a) and cross the lower
135 part of the colluvial slope to the east at ~900-1000 m a.s.l. (Fig. 2b). Particular
136 attention is given to the landforms numbered 1, 2, 3, 7 and 8 in Fig. 1. It should be
137 noted that Smørbotn 7 and 8 are located above the Younger Dryas glacier limit
138 whereas the sites of Smørbotn 1-3 would have been ice covered at that time. Today,
139 large snowbeds occur up-slope of all of these landforms but those associated with
140 Smørbotn 1-3 survive for much longer into the summer than those associated with
141 Smørbotn 7 and 8 (compare, for example, Figs 2a and 2b).

142

143 Smørbotn and its depositional landforms lie in the alpine zone above the local
144 birch (*Betula pubescens*) tree line (Moen 1999), which here attains an altitude of ~700
145 m a.s.l. (Fig. 1). The landscape within the cirque is typical of the low- to mid-alpine
146 zones in maritime western Norway and is characterised by a periglacial environment
147 with a seasonal freezing regime. Based on the meteorological station at Åndalsnes (20
148 m a.s.l.; 12 km to the NW; Aune 1993; Førlund 1993), and allowing for a temperature
149 lapse rate of 0.6 °C per 100 m and a precipitation increase with altitude of 8 % per 100
150 m (Dahl and Nesje 1996), the mean annual air temperature (MAAT) at 800 m a.s.l. in
151 Smørbotn is about 1.5 °C and the mean annual precipitation (MAP) is about 1400
152 mm, more than half of which falls as snow during the six months when mean monthly
153 temperatures lie below zero. Alpine permafrost is patchy within the region but is
154 likely to exist at or near the summit of Finnan (*cf.* Etzelmüller et al. 2003; Lilleøren et
155 al. 2012). The local bedrock is predominantly migmatitic gneiss, though gneissic
156 lithologies are more variable on the eastern side of Smørbotn, beneath Finnan (Tveten
157 et al. 1998).

158

159

160 **Methods**

161

162 Geomorphological mapping of the landforms was based on aerial photographs
163 (available from <http://www.norgebilder.no>) and field visits. This was complemented
164 by the cross-profiles, clast roundness, clast size and particle size measurements and
165 Schmidt-hammer exposure-age estimates available from previous investigations
166 (Shakesby et al. 1995, 1999; Matthews et al. 2011; Matthews and Wilson 2015). Also

167 available for this paper was a new cross-profile from Smørbotn 7 and Schmidt-
168 hammer exposure-age dating (SHD) of Smørbotn 8. Cross-profiles were based on
169 breaks of slope using a tape and Suunto clinometer. Clast roundness measurement was
170 based on the roundness chart of Powers (1953) and samples of 50 boulders: use of
171 clast roundness histograms was supplemented by calculation of percentage angularity
172 and mean roundness (Matthews et al. 2011; see also Matthews 1987). SHD techniques
173 were the same as those detailed in Matthews and Wilson (2015) apart from the use of
174 one impact per boulder (rather than three) from each site associated with Smørbotn 8.
175 Their ‘Romsdalsalpane/Valldalen’ linear age-calibration equation, $y = 37022.951 -$
176 $656.28571x$, was used, where $y =$ SHD age and $x =$ mean R-value. This uses two
177 control points derived from a modern landslide and bedrock road cuttings (young
178 control point) and Younger Dryas moraines (old control point). SHD ages and their
179 95% confidence intervals take account of the sampling error of the landform (C_s) as
180 well as the error associated with the calibration curve (C_c) and have been rounded to
181 the nearest 5 years. The main sources of error are the relatively small sample sizes
182 used to derive C_s and uncontrolled lithological variations within the gneissic bedrock
183 of Smørbotn, which may not be reflected in C_c . Snow, ice and debris fluxes affecting
184 each landform were assessed qualitatively by comparative observations of the sites
185 made over several field seasons with particular reference to the local nature of the
186 cirque headwall and surrounding slopes, the extent of extant snowbeds, and the nature
187 and distribution of debris associated with these.

188

189

190 **Results**

191

192 *Smorbotn 1*

193

194 Smørbotn 1 is an arcuate depositional ridge that has been extensively investigated by
195 Shakesby et al. (1995, 1999) and was SHD dated by Matthews and Wilson (2015).

196 This landform extends over a horizontal distance of about 380 m close to the foot of
197 the cirque headwall (Figs 3a and 4a).

198

199 The principal morphological feature is a prominent main ridge, which consists
200 of two lateral limbs separated by a large central breach eroded by the snow meltwater

201 stream. The main ridge is up to 30 m wide at its base and its crest extends up to an
202 estimated 100 m from the bedrock slabs of the backwall and stands up to 5 m above
203 the outer edge of the large area of snow accumulation. The area of this semi-
204 permanent snowbed/residual ice body shown in Fig. 3a is typical for mid-August but
205 depends in part on the amount of snow avalanching annually from the upper parts of
206 the headwall. As is the case for all the landforms in the cirque, the distal slope of the
207 main ridge (maximum slope angle 36 °) is steeper than the proximal slope (maximum
208 slope angle 32 °) and the down-slope limit of the distal slope is more difficult to
209 define as it merges into the valley-side slope below (Fig. 4a).

210

211 Important minor morphological features include a group of several small
212 ridges up to 1.2 m high located towards the south-eastern end of the landform where
213 the lateral limb takes on a ramp-like cross-section without a proximal slope (Fig. 5a).
214 Shakesby et al. (1999) recognised five of these small-scale ridges on the surface of the
215 ramp and described their asymmetrical cross-profiles with relatively low-angle,
216 smooth and compacted proximal slopes. These minor ridges were interpreted as snow-
217 push ridges by Shakesby et al. (1999). A similar secondary ridge, which contains
218 more boulders, runs parallel to the main ridge low on the proximal slope of the
219 western limb (Fig. 5b).

220

221 Both the major (Shakesby et al. 1995) and the minor ridges (Shakesby et al.
222 1999) are composed of diamictons with a predominantly sandy matrix and variable
223 quantity of larger clasts. Samples of clasts from the ridges are mostly characterized by
224 an angular modal class but are distinctly less angular than supranival clasts sampled
225 from the snowbed surface, which have a very angular modal class. Ridge clasts are
226 also marginally more angular than subnival clasts sampled from beneath the snowbed
227 (a substantial number being subangular).

228

229 SHD ages for Smørbotn 1 are close to modern: -1210 ± 945 years and $815 \pm$
230 1000 years, for the western and eastern limbs of the main ridge, respectively
231 (Matthews and Wilson 2015), which imply the landform is currently active.

232

233 *Smørbotn 2 and 3*

234

235 Smørbotn 2 and 3 are linear to sub-arcuate ridges composed of diamictons, and they
236 extend over horizontal distances of about 300 m and 360 m, respectively. Both lie
237 closer to the foot of the bedrock headwall of the cirque than Smørbotn 1 and are
238 separated from the headwall by relatively narrow semi-permanent snowbeds fed by
239 snow avalanches (Figs 3b and 4b). The estimated maximum distance from the foot of
240 the headwall to the crest of the main depositional ridge of both landforms is no more
241 than 30 m.

242

243 The geomorphology of Smørbotn 2 was investigated by Shakesby et al. (1995,
244 1999) but it has not been dated. It consists of a single main ridge at its west end,
245 which takes on a ramp-like cross section (proximal slope absent) at its eastern end.
246 The ridge, which stands up to about 3 m above the lower edge of the proximal
247 snowbed, has a basal width of up to 60 m, largely accounted for by the steep but
248 relatively long distal slope (maximum distal slope angle 36° , length about 45 m;
249 maximum proximal slope angle 26° , length about 10 m) (Fig. 4b). Two meltwater
250 streams drain from the snowbed and cross the ridge at low points but notably have not
251 incised into the ridge. Several fragments of minor ridges (<1 m high) occur proximal
252 to the main ridge (Fig. 5c), which have been described by Shakesby et al. (1999).

253

254 Smørbotn 3 is the narrowest of the landforms in Smørbotn. The best developed
255 ridge form is associated with the eastern half of the landform, adjacent to where the
256 proximal snowbed tends to be most extensive (Fig. 3c). The main ridge, which stands
257 up to about 2.5 m above the snowbed edge and is no more than 30 m wide, is
258 breached in two places by meltwater streams. Several partial breaches exist in the
259 western section of the landform, which forms a ramp rather than a ridge (Fig. 4c).
260 Slope profiles in Shakesby et al. (1995) and Matthews et al. (2011) indicate a
261 maximum distal slope angle of 33° and a maximum proximal slope angle of 27° .
262 Another notable feature of Smørbotn 3 is a distinct secondary ridge with a maximum
263 height of about 0.5 m which extends for a distance of about 30 m parallel to the crest
264 of the main ridge on its proximal side (Fig. 5d). All clast samples from major and
265 minor ridges examined by Shakesby et al. (1995) had an angular mode with
266 appreciable numbers of subangular clasts.

267

268 SHD dating by Matthews and Wilson (2015) yielded ages of -1355 ± 1240
269 years and 140 ± 1890 years for the western and eastern parts of the main ridge,
270 respectively.

271

272 *Smørbotn 7 and 8*

273

274 Smørbotn 7 and 8 are lobate rather than linear or arcuate in plan (Figs 3d and 3e).
275 They take the form of mid-slope ‘bulges’ on the mountain-side with relatively flat,
276 largely bench-like, boulder-strewn upper surfaces and long, steep distal slopes, which
277 reveal a fine matrix as well as clasts up to boulder size (Figs 6a and 6b). They extend
278 for horizontal distances of 470 m and 310 m, respectively, and the outer edges of their
279 upper surfaces lie up to 60 m from the toes of the associated colluvial fans. A high
280 concentration of boulders produced by the gravity sorting of debris, characterises the
281 lower fringe of the distal slopes (Fig. 6b).

282

283 Unlike the largely smooth bedrock slabs of the headwall behind Smørbotn 1-3,
284 the bedrock on the upper slopes of Finnan is highly weathered and eroded into rocky
285 pinnacles and gullies. These upper slopes are the source of debris transported towards
286 Smørbotn 7 and 8 by rock fall and snow avalanche and reworked by debris flows and
287 meltwater streams on the colluvial fans. Although boulder roundness was not
288 measured on these landforms, observation clearly confirms an angular modal class,
289 similar to the majority of sites where roundness has been measured at Smørbotn 1-3.

290

291 A cross-profile through the northern limb of Smørbotn 7, the larger of the two
292 landforms, is shown in Fig. 7. The maximum angle of the distal slope on the profile of
293 38° is representative of the distal slope elsewhere on Smørbotn 7. This contrasts with
294 the maximum angle of the proximal slope of 15° , which is similar to the maximum
295 slope of the toe of the alluvial fan to the east. The cross-profile includes three
296 transverse ridges of low relief separated by shallow furrows (illustrated in Fig. 9a),
297 which are about 7-8 m apart and separated from the toe of the colluvial fan by a
298 trough that is up to 5 m deep measured against the crest of the proximal ridge (see
299 also Fig. 6a). Elsewhere, a single broad ridge occurs along parts of the edge of the
300 bench. Proximal troughs are less well developed or absent away from the cross-
301 profile: the absence of troughs and ridges likely indicates where past snow-avalanche

302 activity has overtopped the landform and may have eroded ridges, filled in troughs
303 and/or prevented their formation. Another distinctive minor feature associated with
304 the area of transverse ridges is upstanding elongate boulders shown in Fig. 9b: these
305 were not observed on any of the other landforms in the cirque.

306

307 Smørbotn 8 is in many ways a smaller version of Smørbotn 7 with a similar
308 lobate form in plan, a bench-like upper surface with ridges near the outer margin (Fig.
309 3e) and a long, steep distal slope (Fig. 6b). Three poorly-developed transverse ridges
310 20-30 m apart, which are separated by shallow troughs, can be recognized on the
311 upper surface near the centre of the bench (Fig. 3e).

312

313 Matthews and Wilson (2015) obtained SHD ages of $13,995 \pm 1070$ years and
314 $14,635 \pm 1060$ years for the northern and southern limbs, respectively, of Smørbotn 7.
315 Schmidt-hammer R-value distributions from Smørbotn 8 are summarized in Fig. 8
316 and the resultant SHD ages are shown in Table 1. Mean R-values yielded SHD ages
317 of $10,770 \pm 2245$ years and $13,095 \pm 2120$ years for the southern and northern limbs,
318 respectively. An age of 7200 ± 1715 years was obtained for the surface of the
319 colluvial (avalanche) fan associated with Smørbotn 8.

320

321

322 **Discussion**

323

324 *Pronival rampart development at Smørbotn 2 and 3*

325

326 On the basis of their morphology, sedimentological composition and landscape
327 setting, there can be little doubt that Smørbotn 2 and 3 are undisputed pronival
328 ramparts. These characteristics agree with several proposed diagnostic criteria derived
329 from theoretical considerations (see especially Ballantyne and Benn 1994) and/or the
330 empirical evidence of actively developing pronival ramparts (see especially, Shakesby
331 1997; Hedding and Sumner 2013).

332

333 These landforms comprise single linear to sub-arcuate main ridges or ramps,
334 which lie close to the foot of the bedrock backing cliff. The maximum ridge-crest to
335 cliff-foot distance of 30 m places the whole of both landforms well within the roughly

336 estimated 30-70 m backwall to ridge crest distance necessary for a snowbed that is
337 steep enough for supranival sediment transport, which probably requires slope angles
338 $>20^\circ$, yet thin enough to avoid conversion of snow to glacier ice, which may require
339 >30 m snow thickness (Ballantyne and Benn 1994; Benn and Evans 2010). Ridge and
340 ramp morphology is consistent with snowbeds repeatedly forming in the same place,
341 though snowbed dimensions vary annually, particularly in response to the extent of
342 snow-avalanche activity, as has been observed at these sites (Fig. 10a).

343

344 Matthews et al. (2011) pointed to the intimate relationship between pronival
345 ramparts and snow-avalanche activity in Smørbotn: although pronival ramparts are
346 fed by frequent additions of snow-avalanche debris, they can also be absent, breached
347 and/or destroyed by high-magnitude events. Smørbotn 2 and 3 have formed beneath
348 steeply sloping parts of the bedrock cliffs that are not vertical and are therefore
349 conducive to snow accumulation and slab avalanches (Schweizer et al. 2003;
350 McClung and Schaerer 2006). At these sites the avalanches appear to be relatively
351 small, largely non-erosive and relatively clean, normally transporting only small
352 increments of debris onto the ramparts during each event.

353

354 Short proximal slopes to the ridges and ramp-like sections are explicable in
355 terms of infilling by long-term rockfall and snow-avalanche deposition (Fig. 10b) and
356 subsequent reworking by subnival and pronival processes (Shakesby et al. 1995).
357 Long distal slopes reflect the overtopping of the ridge crests by relatively high-
358 magnitude snow-avalanche and rockfall events. Where proximal slopes are best
359 developed, these may be steepened by snow-push processes, evidence for which
360 includes the minor ridges proximal to the main ridge crests (Shakesby et al. 1995).

361

362 Although coarse sediment textures, predominantly angular clasts and
363 associated openwork fabrics have been widely noted in relation to pronival ramparts
364 and proposed by some as diagnostic (e.g. Washburn 1979; White 1981; Ballantyne
365 and Kirkbride 1986; Ballantyne 1987; Hedding et al. 2007) other pronival ramparts
366 contain diamictons with an appreciably fine matrix consistent with the sediments at
367 Smørbotn 2 and 3 (cf. Harris 1986, Hall and Meiklejohn 1997). The former type of
368 sediments is likely where coarse rockfall debris generation by frost weathering is the
369 dominant process. This would explain the angular mode in clast roundness data but

370 not the abundant fine matrix. However, where snow-avalanche deposition is more
371 important, as in Smørbotn, the proportion of fines may be accounted for by the nature
372 of the debris being removed from the backwall, which could have been concentrated
373 further by supranival wash (Derbyshire et al. 1979) and deposited and reworked in
374 subnival or pronival positions by streams, debris flows and/or solifluction (Shakesby
375 1997).

376

377 The SHD results from Smørbotn 3 are consistent with active, modern surfaces
378 with few relatively old boulders exposed on the surface of the ridge, the apparent
379 negative (futuristic) age of the western limb being attributable to limitations of the
380 calibration curve (Matthews and Wilson 2015). However, both Smørbotn 2 and 3 are
381 probably much older than the exposure ages of their surface boulders. Indeed, there is
382 no reason to think that these landforms have not been developing since the melting of
383 the Younger Dryas glacier that occupied the cirque; their development throughout the
384 Holocene being dependent on fluctuations in the rates of rockfall and snow-
385 avalanching. It would be expected that such century- to millennial-scale fluctuations
386 in snow-avalanche activity would reflect, at least to some extent, the known temporal
387 pattern of Holocene glacier and climatic variations in southern Norway (Matthews
388 and Dresser 2008; Nesje et al. 2008; Nesje 2009; Fig. 11).

389

390 *The pronival rampart/moraine transition at Smørbotn 1*

391

392 Smørbotn 1 appears to represent the transition from a pronival rampart to a moraine
393 (i.e. a landform deposited directly by a glacier). It is not a significantly larger ridge
394 than either Smørbotn 2 or 3 but its arcuate plan form extends much farther from the
395 foot of the backwall. The proximal snowbed is therefore larger and thicker than at
396 Smørbotn 2 or 3 and, at its greatest extension from the backwall, the planform of the
397 terminal section (Fig. 3a) is reminiscent of the 'snout' of a very small glacier. The
398 relatively large terminal breach in the main ridge is moreover consistent with a
399 relatively large meltwater stream issuing from a glacier.

400

401 Another distinctive feature of the main ridge is its well-developed, steep
402 proximal slope: the ridge takes on a ramp-like form only at its southern extremity. A
403 steep proximal slope may conceivably be produced by a snow-push mechanism, as

404 advocated by Shakesby et al. (1999) for the minor ridges on the ramp-like section.
405 However, its scale is more likely to be accounted for by a combination of glacial
406 erosion, push and bulldozing by a rigid glacier tongue moving over bedrock slabs in
407 its upper part and diamicton nearer to its terminus. Although clast roundness at most
408 of the sites investigated at Smørbotn 1 (Shakesby et al. 1995) exhibits an angular
409 mode, the substantial proportion of subangular clasts is consistent with subglacial
410 abrasion (Boulton 1978; Lukas et al. 2013), which is likely to be effective even
411 beneath very small glaciers (*cf.* Matthews 1987). Thus, we suggest that moraine ridge
412 formation at Smørbotn 1 involved subglacial as well as supraglacial sediments.
413 Furthermore, rather than a ‘protalus moraine’ of the type envisaged by Ballantyne and
414 Benn (1994), which is simply the product of the deposition of supraglacial debris by a
415 dumping process, we propose a ‘push moraine’ origin that is built up incrementally
416 and incorporates a wider range of sedimentary material. Pushing (bulldozing) of
417 proglacial sediments seems to be the dominant mechanism of recent moraine ridge
418 formation at warm-based (temperate) glaciers at relatively low altitudes in southern
419 Norway (Winkler and Matthews 2010).

420

421 Shakesby et al. (1999) argued strongly that Smørbotn 1 was a pronival rampart
422 produced by a snowbed rather than a moraine ridge produced by a glacier. Based
423 largely on similarities to the adjacent pronival ramparts, snowbed thickness and snow
424 density, they concluded that the minor ridges could be accounted for by the snow-
425 push mechanism and that this mechanism could be scaled-up to account for the main
426 ridge. However, the maximum ridge-crest to cliff-foot distance of 100 m lies beyond
427 the lower limit normally regarded as necessary to produce sufficient snow thickness
428 for the transformation of snow to ice. Snow densities measured at the site by
429 Shakesby et al. (1999) of up to 900 kg m^{-3} exceed the value of 830 kg m^{-3} indicative
430 of ice rather than firn (Benn and Evans 2010). The maritime location of Smørbotn is,
431 moreover, conducive to melting and refreezing of snow, which would allow snow to
432 glacier ice transformation exceptionally rapidly, perhaps at snow thicknesses as little
433 as 13 m (Paterson 1994; Benn and Evans 2010). In addition, accumulation of snow
434 from snow-avalanches is likely to produce much larger and thicker snowbeds than
435 those observed by us or shown on aerial photographs taken in mid- to late summer.
436 Indeed, evidence of the former existence of a thicker snowbed and/or glacier is
437 provided by the relatively light colouration of the bedrock (indicative of clean rock

438 surfaces devoid of lichens) that can be seen well above the snowbed edge in Figs 3a
439 and 4a. This light zone implies that extremely large snowbeds are banked against the
440 backwall early in the ablation season. Furthermore, snow-avalanche activity may have
441 been greater, and/or the snowbeds may have persisted for longer, in the recent past
442 (see below).

443

444 The SHD dates from Smørbotn 1 indicate a surface composed primarily of
445 recently deposited boulders. The dates alone are insufficient to distinguish between an
446 active pronival rampart and a recent moraine that continues to be subject to boulder
447 deposition from snow-avalanches. However, the SHD ages are compatible with the
448 existence of a Little Ice Age glacier at this site. It is firmly established that many
449 southern Norwegian glaciers were larger in the Little Ice Age of the last few centuries
450 than at any time since the early Holocene (Nesje et al. 2008; Matthews and Dresser
451 2008; Nesje 2009). Based on accurate historical evidence at several glaciers, and less
452 accurate lichenometric dating at many more, their Little Ice Age maxima can be
453 precisely dated to the mid-eighteenth century (Grove 1988, 2004; Bickerton and
454 Matthews 1993; Matthews 2005). Hence, it is proposed here that a major phase in the
455 development of the main ridge at Smørbotn 1 involved deposition by a small,
456 temperate (warm-based) glacier in the Little Ice Age when mean annual snowfall was
457 likely to have been higher than it is today and mean summer air temperature lower
458 (Nesje et al. 2007a). These climatic conditions would, in turn, have led to larger
459 and/or more frequent avalanches (Grove 1972, Nesje et al. 2007b; Vasskog et al.
460 2011), a thicker snowbed, and glacier growth at the site.

461

462 Glaciers may have first formed at the site during mid-Holocene neoglaciation
463 (Matthews 2013) and expanded several times (as suggested by the Holocene glacier
464 variations shown in Fig. 11). Each expansion would have added a new sediment
465 increment to the moraine ridge. Indeed, it is probable that during the long-term
466 development of Smørbotn 1, the main depositional processes alternated between those
467 characteristic of a pronival rampart fronting a large late-lying snowbed and a moraine
468 ridge fronting a small glacier. In this respect, the depositional ridge might be
469 interpreted as a polygenetic landform that has crossed several transitions from
470 pronival rampart to moraine, and *vice-versa*.

471

472 *The development of Smørbotn 7 and 8 and the possibility of a pronival*
473 *rampart/rock glacier transition*

474

475 For these landforms, we follow Haeberli (1985), Barsch (1992, 1996), Haeberli et al.
476 (2006), Berthling (2007), Kääh (2013) and others in defining a rock glacier as an
477 accumulation of rock debris with interstitial ice, segregation ice or an ice core that
478 deforms, fundamentally by creep, under the force of gravity. In valley-wall or talus-
479 foot locations where debris supply is sufficient, they develop distinctive lobes with
480 transverse ridges and furrows (creep or flow structures) on their upper surface. Such
481 ‘lobate rock glaciers’ were termed ‘protalus lobes’ (rather than ‘rock glaciers’) by
482 Whalley and Martin (1992), Hamilton and Whalley (1995), Whalley and Assiz
483 (2003), Harrison et al. (2008) and Whalley (2009). Active rock glaciers of this type
484 are typically indicative of a permafrost environment with a negative mean annual air
485 temperature of -2 to -6 °C and an annual precipitation of 500 to 1500 mm
486 (Etzelmüller and Frauenfelder 2009), generally in somewhat colder and drier climatic
487 conditions than glaciers (Brazier et al. 1998; Humlum 1998; Sattler et al. 2016).

488

489 The bulging lobate form of Smørbotn 7 and 8, combined with evidence of low
490 and broad transverse ridges, proximal troughs and upstanding boulders, lead us to
491 conclude they are embryonic rock glaciers, rather than large pronival ramparts.
492 Transverse-ridge formation at the surface of rock glaciers have been explained by the
493 deformation of rock debris under compressive flow, possibly involving thrusting
494 and/or buckle folding (Loewenherz et al. 1989; Kääh and Weber 2004; Springman et
495 al. 2012; Frehner et al. 2015). Well-developed multiple transverse ridges tend to be
496 best developed on relatively large rock glaciers and/or at sites promoting compressive
497 flow, such as breaks of slope at valley-floor locations. Small, bench-like rock glaciers
498 that resemble closely the small lobes of Smørbotn 7 and 8 and lack well-developed
499 transverse ridges have also been described and seem particularly common in Svalbard
500 (Sollid and Sørbel 1988; Berthling et al. 1998). The fact that transverse ridges are
501 poorly developed at Smørbotn 7 and 8 may be attributed to a reduced propensity for
502 compressional flow on a steep hillslope, combined with the embryonic nature of the
503 landforms and their relict status (see below).

504

505 Major ridges at the outer edges of small rock glaciers and their associated
506 proximal troughs have previously been explained in three ways (Berthling et al.
507 1998). First, Swett et al. (1980) argued for ridges formed by the upward movement of
508 material associated with compressional flow. Second, Liestøl (1962) attributed such
509 ridge development to snow avalanches or rockfalls coming to rest in a narrow zone
510 beyond a proximal snowbed. The third explanation regards the proximal troughs as
511 thermokarst phenomena caused by ice melting in response to either climatic
512 amelioration (Humlum 1982) or the presence of areas of extensional flow (Berthling
513 et al. 1998). All of the above explanations seem compatible with the evidence
514 available at Smørbotn 7 and 8.

515

516 Upstanding boulders on the crest and upper-distal slope of Smørbotn 7 are
517 suggestive of internal thrusting – the specific mechanism proposed by Springman et
518 al. (2012) for the formation of transverse ridges and furrows on rock glaciers. Similar
519 boulders, thrust up at a steep angle to the slope, have been reported from ice-cored
520 moraines associated with high-alpine permafrost in Jotunheimen, southern Norway.
521 However, in the case of the ice-cored moraines, deformation was induced by glacial
522 push rather than gravity (Matthews et al. 2014).

523

524 The position of Smørbotn 7 and 8 at the toes of large-scale colluvial fans
525 appears essential to any explanation for the development of these rock glaciers. The
526 fans are regarded as snow-avalanche fans because they are not as steep as talus cones
527 and have a concave profile, characteristic of avalanche run-out (Rapp 1959; Luckman
528 1977; Jomelli and Francou 2000). Increments of snow and rock debris at the fan toes
529 especially in the spring, together with intermittent summer thawing of snow,
530 percolation and refreezing of water, and winter re-freezing, provide a potential source
531 of ice for the ice-debris mixture necessary for rock-glacier creep (Humlum et al.
532 2007)

533

534 The SHD dates indicate that Smørbotn 7 and 8 are relict landforms. Both dates
535 from Smørbotn 7 and the date obtained from the northern limb of Smørbotn 8 fall
536 within the range $14,635 \pm 1060$ to $13,095 \pm 2120$ years and hence suggest that much
537 of their development had occurred before the Younger Dryas Stadial (~12.9–11.7 ka;
538 see Fig. 11). The younger age of $10,770 \pm 2245$ years for the southern limb of

539 Smørbotn 8 can be attributed to the cluster of Schmidt-hammer R-values in the range
540 50-60 (Fig. 8), which probably reflect boulders deposited by a relatively recent snow-
541 avalanche event that exceptionally reached the rock glacier from the adjacent fan. The
542 age of 7200 ± 1715 years for the associated fan surface is similarly consistent with an
543 essentially relict landform but, in this case, exhibiting somewhat more evidence of
544 boulder deposition during the Holocene.

545

546 The thickness of the Late Weichselian Ice Sheet in southern Norway at the
547 Last Glacial Maximum (LGM) is disputed. On the basis of terrestrial cosmogenic
548 nuclide dating (TCND) of boulders and bedrock associated with similar coastal
549 mountains to those surrounding Smørbotn, however, combined with the results of ice-
550 sheet modelling, it is probable that the upper altitudinal limit of the ice-sheet at the
551 LGM was ~ 1500 m a.s.l. in the study area (Goehring et al. 2008; Stroeven et al. 2015;
552 Hughes et al. 2016). Thus, the rock pinnacles and gullies of the summit area of Finnan
553 existed as nunataks at the LGM, and the slopes of Smørbotn at the sites of the rock
554 glaciers appear to have been exposed during rapid deglaciation between about 15.0
555 and 14.0 ka. If these estimates are correct, formation of the rock glaciers took place
556 over a time interval of no more than ~ 3000 years between deglaciation and the
557 Younger Dryas-early Holocene transition.

558

559 For rock glacier development to have occurred within such a relatively short
560 time interval as a few thousand years, environmental conditions must have been very
561 different from today. Debris supply rates would have been considerably higher during
562 and immediately after deglaciation, when bedrock instability and colluvial activity are
563 likely to have been triggered by glacier debuitressing (Jarman 2006; McColl 2012;
564 McColl and Davies 2012) and permafrost degradation (Gruber et al. 2004; Gruber and
565 Haeberli 2007; Deline et al. 2015). Permafrost is likely to have been ubiquitous during
566 deglaciation, both on the nunataks and in the bedrock beneath the cold-based ice sheet
567 (Sollid and Sørbel 1994; Kleman and Hättestrand 1999; Kleman and Glasser 2007;
568 Goehring et al. 2008). Residual permafrost can be hypothesized as contributing to the
569 necessary ice content for rock glacier creep at Smørbotn 7 and 8. Stabilization of the
570 rock glaciers and the surrounding slopes would have accompanied continuing
571 permafrost thaw. Interestingly, the return of a permafrost climatic regime in the
572 Younger Dryas appears not to have produced a major resurgence in rock glacier

573 development, though activity probably continued, at least at Smørbotn 8. Thus, in
574 terms of debris supply, Smørbotn 7 and 8 are considered to be primarily paraglacial
575 and/or paraperiglacial landforms dependent on complex interactions between
576 paraglacial and/or paraperiglacial slope processes shortly after deglaciation (*cf.*
577 Ballantyne 2002; Mercier 2008). In terms of rock glacier movement, however, they
578 are the product of periglacial processes.

579

580 In rock-glacier development, a pronival rampart might form the earliest stage,
581 as suggested in the model of Haeberli (1985), Strelin and Sone (1998), Haeberli et al.
582 (2003), Scapozza et al. (2011), Kääb (2013) and Scapozza (2015). If so, the extreme
583 lateral margins of Smørbotn 7 and 8 could be viewed as evidence of this transient
584 stage, prior to the majority of both landforms having been transformed into rock
585 glaciers by permafrost creep. Although this remains a possibility, at least in theory, it
586 seems an unnecessary complication in the cases of Smørbotn 7 and 8. We prefer,
587 therefore, to view these landforms as embryonic rock glaciers developed entirely by
588 permafrost creep within the colluvial fans without a pronival rampart precursor.

589

590 *The periglacial-glacial landform continuum and landform transitions*

591

592 Shakesby et al. (1987) introduced the idea of a morphological and
593 developmental continuum of large scale talus-derived landforms in Rondane, southern
594 Norway, which extended from talus slopes to pronival ramparts, rock glaciers and ice-
595 cored ‘push-deformation’ moraines. In their schematic model these landforms are
596 characterised by different transport processes (sliding of debris over snow, gravity
597 deformation or glacier ice-push deformation, respectively) and develop along separate
598 pathways leaving little room for transitional forms between the recognized landform
599 types (see also Shakesby et al. 1989). In a critical appraisal of this model, Kirkbride
600 (1989) replaced rockfall/talus with debris supply from a broader range of debris
601 sources, and focused on the relative importance of fluxes of snow/ice and debris as the
602 primary determinant of the landform types (see also the related concepts and models
603 of Giardino and Vitek 1988; Whalley 2009; and Janke et al. 2013). Thus, changes in
604 these fluxes through time (especially due to climate change) and over space (due to
605 local site characteristics) can be seen as important dimensions of the landform
606 continuum, and transitions of one landform type to another are accommodated. These

607 ideas may be taken further in the light of the pronival ramparts and related landforms
608 in Smørbotn, resulting in an improved conceptual model of part of the periglacial-
609 glacial landform continuum and containing potential transitional landforms (Fig. 12).

610

611 As many environmental factors are held constant within the Smørbotn cirque,
612 and SHD ages have been determined, it is possible to infer the range of potential
613 causal factors affecting the differences between the various landforms. The main
614 differences relate to: (1) the extent of the extant snowbeds, and hence identification of
615 glacier formation at Smørbotn 1 during the Little Ice Age and possibly at other times
616 during the seasonal freezing regime of the Holocene; and (2) the location of the
617 colluvial fans and hence recognition of the relatively high rate of paraglacial and
618 paraperiglacial debris supply to now relict Smørbotn 7 and 8 during the permafrost
619 regime that accompanied ice-sheet deglaciation.

620

621 The development of Smørbotn 2 and 3 as pronival ramparts is represented at
622 the lower left of Fig. 12. These features are active under the present seasonal-freezing
623 climatic regime and are fed by snow and debris inputs from rockfalls and snow
624 avalanches. With greater snow thickness at Smørbotn 1, and the transformation of
625 snow to glacier ice, it is inferred that a pronival rampart was transformed into a
626 push/dump moraine at this site. This landform transition therefore involved crossing a
627 threshold from the periglacial to the glacial domains. If it is assumed that a small
628 glacier formed at this site more than once during the Holocene (as suggested above), it
629 is possible that the development of pronival ramparts and moraines alternated at the
630 site. Arguably, therefore, a two-headed arrow could be inserted between push/dump
631 moraine and pronival rampart in Fig. 12. Similar pronival ramparts at the sites of
632 Smørbotn 7 and 8 may have been transformed into rock glaciers under the different
633 climatic conditions that pertained following deglaciation, when debris supply was
634 enhanced and interstitial ice or ice bodies could have developed and/or survived under
635 the permafrost regime. However, our simpler explanation is that these lobate rock
636 glaciers (protalus lobes) developed without the transitional pronival-rampart stage, as
637 indicated at the upper left of Fig. 12. Both alternatives involve landform evolution
638 wholly within the periglacial domain, albeit with enhanced paraglacial and/or
639 paraperiglacial debris supply.

640

641 Ice-cored moraines, which form a part of the periglacial-glacial landform
642 continuum not represented in Smørbotn, are introduced at the upper right of Fig. 12.
643 There has been much discussion in the literature concerning the possibility of the
644 transformation of ice-cored moraines into rock glaciers and whether ice-cored
645 moraines are rock glaciers (Østrem 1964, 1971; Barsch 1971, 1977; Etzelmüller and
646 Hagen 2005; Lilleøren and Etzelmüller 2011; Lilleøren et al. 2013; Matthews et al.
647 2014). Ice-cored moraines are deposited by glaciers and the deformation that produces
648 their transverse ridges results from glacial movement, not the force of gravity alone as
649 is the case with rock glaciers (hence the term ‘push-deformation moraine’ coined by
650 Shakesby et al. 1987). Our scheme allows for the transformation of ice-cored
651 moraines into rock glaciers (i.e. crossing from the glacial to the periglacial domain)
652 due to gravity deformation, which generally requires the moraines to have been
653 deposited on sloping terrain.

654

655 Landform transitions between pronival ramparts and rock glaciers, pronival
656 ramparts and push/dump moraines, push/dump moraines and ice-cored moraines, and
657 ice-cored moraines and rock glaciers, appear to be recognisable in nature. Other
658 transitions are likely to exist, the debris-covered glacier to rock-glacier transition
659 being a case in point. Thus, there is considerable potential to expand on our
660 interpretation of the periglacial-glacial landform continuum. Many such transitions
661 are associated with thresholds in the Earth-surface processes operating within and
662 between the periglacial and glacial domains and across permafrost and seasonal-
663 freezing regimes. As such, they are of relevance when using landforms for
664 palaeoclimatic reconstruction. Landform transitions may be rarely encountered in
665 present-day landscapes, however, because they are often short-lived and/or the
666 relevant thresholds may not have been crossed during the relatively small-scale
667 climatic fluctuations of the Holocene.

668

669 *Diagnostic criteria revisited*

670

671 The existence of transitional landforms adds to the difficulties in differentiating
672 pronival ramparts from related landforms. In the light of the landform transitions
673 identified in Smørbotn and analysed within the conceptual framework of Fig. 12, none
674 of the diagnostic criteria listed for pronival ramparts by Shakesby (1997) or Hedding

675 and Sumner (2013), can be regarded as definitive. Single morphological or
676 sedimentological criteria are certainly not universally applicable and, even in
677 combination, may be misleading. It should be possible instead, to identify each type
678 of landform using a combination of criteria as developed in this paper. These criteria
679 relate to morphological and sedimentological characteristics that take account of the
680 local environment and landscape context, including land-surface processes, climatic
681 regime, and landform age and development.

682

683

684 **Conclusions**

685

686 (1) The landforms in Smørbotn cirque, southern Norway, include two pronival
687 ramparts, a moraine ridge, and two rock glaciers. Classification has been based on the
688 similarities and differences between these landforms in terms of their morphology,
689 sedimentology, SHD age and position in the landscape in relation to inferred snow,
690 ice and debris supply.

691

692 (2) The pronival ramparts (Smørbotn 2 and 3) are linear to subarcuate ridges or
693 ramps with a steep distal slope and little or no proximal slope. They are active, lie
694 close to bedrock cliffs, and associated with late snowbeds fed by snow-avalanches.

695

696 (3) The moraine (Smørbotn 1) is an arcuate ridge, which extends up to ~100 m
697 from the foot of the bedrock cliff. The steep distal slope of the ridge is matched by an
698 equally steep proximal slope for most of its length. It is argued that this moraine
699 developed from a pronival rampart, most notably during the Little Ice Age when the
700 snowbed became permanent and was transformed into a small temperate glacier.

701

702 (4) The rock glaciers (Smørbotn 7 and 8) are somewhat larger, distinctly lobate
703 forms, with a very steep and long distal slope and a relatively flat, bench-like upper
704 surface on which transverse ridges and up-thrust boulders are notable features. The
705 rock glaciers are essentially relict and have been inactive for most, if not all, of the
706 Holocene. It is inferred that they developed under a permafrost regime following Late
707 Weichselian deglaciation, when glacier debuitressing and permafrost degradation
708 enhanced debris supply. Transformations of pronival ramparts into rock glaciers may

709 be possible in theory but are considered unlikely, at least in Smørbotn cirque.

710

711

712 (5) Landform transitions are interpreted in a conceptual model of part of the
713 periglacial-glacial landform continuum, including pronival ramparts, push/dump
714 moraines, ice-cored moraines and rock glaciers (Fig. 12). This model encompasses
715 spatial and temporal variations in the supply of snow, ice and debris across periglacial
716 and glacial domains, and emphasises the trigger factors that may lead to the crossing
717 of process thresholds under seasonal freezing and/or permafrost climatic regimes.

718

719 (6) Landform transitions may be rarely encountered in the landscape due to their
720 short-lived nature and/or the relatively small-scale of Holocene environmental
721 changes. Recognition of transitions between conventional landform types adds to the
722 complexities of determining diagnostic landform characteristics and using landforms
723 for palaeoenvironmental reconstruction. Single morphological or sedimentological
724 criteria are unlikely to be definitive. This paper recommends use of a combination of
725 criteria within the local environment and landscape context, taking account of land-
726 surface processes, climatic regime, and landform age and development.

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1230 **Figure captions**

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1232

1233 Fig. 1. Smørbotn cirque, the pronival ramparts and related transitional landforms
1234 (numbered), selected other geomorphological features, and (inset) its location within
1235 southern Norway.

1236

1237 Fig. 2. The local setting of the pronival ramparts and related transitional landforms
1238 within Smørbotn cirque. (a) Smørbotn 1-3 (right to left) beneath the northeast-facing
1239 bedrock headwall photographed from Smørbotn 7 in August 2015. (b) Smørbotn 7
1240 (centre right) and 8 (centre) on the west-facing colluvial slope beneath Finnan
1241 photographed from the cirque floor in August 2015 (note also the Younger Dryas
1242 lateral moraine in the foreground)

1243

1244 Fig. 3. Geomorphological maps of the pronival ramparts and related transitional
1245 landforms. (a) Smørbotn 1, (b) Smørbotn 2, (c) Smørbotn 3, (d) Smørbotn 7, (e)
1246 Smørbotn 8. Snow extent in August 2006 is based on aerial photographs.

1247

1248 Fig. 4. Main ridges and distal snowbeds associated with Smørbotn 1-3. (a) Smørbotn
1249 1, (b) Smørbotn 2 and (c) Smørbotn 3, all photographed from Smørbotn 7 in August
1250 2013. Note the smaller snowbeds than in 2006 and 2015.

1251

1252 Fig. 5. Minor ridges associated with Smørbotn 1-3. (a) Smørbotn 1, on the bench
1253 surface of the eastern limb (1997). (b) Smørbotn 1, near the base of the proximal slope
1254 of the western ridge (1996). (c) Smørbotn 2, on the proximal slope of the western
1255 ridge (1997). (d) Smørbotn 3, near the crest of the eastern ridge (1993).

1256

1257 Fig. 6. Major features of Smørbotn 7 and 8 photographed in August 2015. (a) The
1258 upper surface of Smørbotn 7 from Smørbotn 8 (note, from left to right, snowbed
1259 remnants on toe of colluvial fan, trough, main ridge and distal slope). (b) The distal
1260 slope and lobate form of Smørbotn 8 from Smørbotn 7 (note also the rock pinnacles
1261 and gullies of Finnan in the background).

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1263 Fig. 7. Cross profile (slope angle of slope segment given in degrees; no vertical
1264 exaggeration; T = transverse ridge) from Smørbotn 7.

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1266 Fig. 8. Schmidt-hammer R-value histograms from Smørbotn 8 (class interval = 2
1267 units; N = northern limb; S = southern limb; C = combined data).

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1269 Fig. 9. Minor features of Smørbotn 7. (a) Transverse ridges (a broad-crested ridge
1270 extends from top left to bottom right with a boulder-filled trough to its left). (b)
1271 Upstanding boulders (at least four are visible in this photograph, each appearing to
1272 have been up-thrust out of the slope at a steep angle).

1273

1274 Fig. 10. Evidence of snow-avalanches and their effects. (a) Avalanche snow spilling
1275 over Smørbotn 3 (foreground) and Smørbotn 2 (background; 1993). (b) Debris spread
1276 by rockfall and/or snow-avalanche over the proximal slope, minor ridges and
1277 proximal snowbed associated with Smørbotn 2 (1997).

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1279 Fig. 11. Summary of the SHD ages (\pm 95% confidence intervals) of Smørbotn
 1280 landforms in relation to Late Weichselian deglaciation, the Younger Dryas Stadial,
 1281 and century- to millennial-scale Holocene glacier expansion episodes reconstructed in
 1282 the Smørstabbtindan massif, Jotunheimen, southern Norway (shaded bands from
 1283 Matthews and Dresser, 2008).

1284
 1285 Fig. 12. Periglacial-glacial continuum model of possible transitions (bold arrows)
 1286 between pronival ramparts, push/dump moraines, ice-cored moraines and rock
 1287 glaciers. The model includes the main inputs and main processes involving snow, ice
 1288 and debris flux (lower case lettering) under seasonal-freezing and/or permafrost
 1289 climatic regimes (shaded areas) within periglacial and glacial domains (large
 1290 unshaded boxes).

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 1300

1301 Table 1. Schmidt-hammer exposure-age dating of the northern (North) and southern
 1302 limbs (South) of Smørbotn 8 and the associated colluvial fan toe (C_s = sampling error;
 1303 C_c = calibration error)
 1304

Site	Mean R-value	Standard Deviation (σ)	Confidence Interval (95%)	Sample size (n)	C_s (years)	C_c (years)	SHD age (years)	Confidence Interval (years)
South	40.00	11.80	± 3.39	50	2225	330	10,770	± 2245
North	36.46	11.13	± 3.20	50	2095	325	13,095	± 2120
Fan toe	45.44	8.92	± 2.56	50	1680	335	7200	± 1715

1305
 1306