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

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Abstract

Landform transitions are defined as intermediate forms that represent transient developmental stages between conventional landform types. This study evaluates possible cases of landform transitions from pronival (protalus) ramparts to moraine ridges, and from pronival ramparts to lobate rock glaciers (protalus lobes) at the foot of the headwall of Smørbotn cirque in southern Norway. The five landforms had been previously classified as pronival ramparts. We conclude that only two (Smørbotn 2 and 3) are undisputed, active pronival ramparts, which developed under the seasonal-freezing regime of the Holocene. It is inferred that a third (Smørbotn 1) represents the transition to a moraine ridge formed during the 'Little Ice Age' of the last few centuries as a semi-permanent snowbed grew into a small temperate glacier. The two others (Smørbotn 7 and 8) appear to be relict embryonic rock glaciers that developed between the Last Glacial Maximum and the Younger Dryas Stadial under a permafrost regime and benefited from enhanced debris supply as a result of rock-slope instability affected by glacier debuttressing and permafrost degradation. Variable landscape settings and distinctive environmental histories contribute to the differences in the morphology of these landforms. We highlight continuing controversies over the modes of formation and diagnostic characteristics of pronival ramparts by positioning them, together with push/dump moraines, ice-cored moraines and rock glaciers, in a conceptual model of the periglacial-glacial landform continuum. The model links snow, ice and debris fluxes under seasonal freezing and/or permafrost climatic regimes to the process thresholds between landform types.

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

Introduction

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

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

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

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

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

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

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

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

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

Smørbotn cirque and its environment

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

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

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

Methods

Geomorphological mapping of the landforms was based on aerial photographs (available from http://www.norgeibilder.no) and field visits. This was complemented by the cross-profiles, clast roundness, clast size and particle size measurements and Schmidt-hammer exposure-age estimates available from previous investigations (Shakesby et al. 1995, 1999; Matthews et al. 2011; Matthews and Wilson 2015). Also available for this paper was a new cross-profile from Smørbotn 7 and Schmidt-hammer exposure-age dating (SHD) of Smørbotn 8. Cross-profiles were based on breaks of slope using a tape and Suunto clinometer. Clast roundness measurement was based on the roundness chart of Powers (1953) and samples of 50 boulders: use of clast roundness histograms was supplemented by calculation of percentage angularity and mean roundness (Matthews et al. 2011; see also Matthews 1987). SHD techniques were the same as those detailed in Matthews and Wilson (2015) apart from the use of one impact per boulder (rather than three) from each site associated with Smørbotn 8. Their 'Romsdalsalpane/Valldalen' linear age-calibration equation, $y = 37022.951 - 656.28571x$, was used, where y = SHD age and x = mean R-value. This uses two control points derived from a modern landslide and bedrock road cuttings (young control point) and Younger Dryas moraines (old control point). SHD ages and their 95% confidence intervals take account of the sampling error of the landform (*Cs*) as well as the error associated with the calibration curve (*Cc*) and have been rounded to the nearest 5 years. The main sources of error are the relatively small sample sizes used to derive *Cs* and uncontrolled lithological variations within the gneissic bedrock of Smørbotn, which may not be reflected in *Cc*. Snow, ice and debris fluxes affecting each landform were assessed qualitatively by comparative observations of the sites made over several field seasons with particular reference to the local nature of the cirque headwall and surrounding slopes, the extent of extant snowbeds, and the nature and distribution of debris associated with these.

Results

Smorbotn 1

Smørbotn 1 is an arcuate depositional ridge that has been extensively investigated by Shakesby et al. (1995, 1999) and was SHD dated by Matthews and Wilson (2015). This landform extends over a horizontal distance of about 380 m close to the foot of the cirque headwall (Figs 3a and 4a).

The principal morphological feature is a prominent main ridge, which consists of two lateral limbs separated by a large central breach eroded by the snow meltwater stream. The main ridge is up to 30 m wide at its base and its crest extends up to an estimated 100 m from the bedrock slabs of the backwall and stands up to 5 m above the outer edge of the large area of snow accumulation. The area of this semi-permanent snowbed/residual ice body shown in Fig. 3a is typical for mid-August but depends in part on the amount of snow avalanching annually from the upper parts of the headwall. As is the case for all the landforms in the cirque, the distal slope of the main ridge (maximum slope angle 36 º) is steeper than the proximal slope (maximum slope angle 32 º) and the down-slope limit of the distal slope is more difficult to define as it merges into the valley-side slope below (Fig. 4a).

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

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

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

Smørbotn 2 and 3

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

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

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

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

Smørbotn 7 and 8

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

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

A cross-profile through the northern limb of Smørbotn 7, the larger of the two landforms, is shown in Fig. 7. The maximum angle of the distal slope on the profile of 38º is representative of the distal slope elsewhere on Smørbotn 7. This contrasts with the maximum angle of the proximal slope of 15º, which is similar to the maximum slope of the toe of the alluvial fan to the east. The cross-profile includes three transverse ridges of low relief separated by shallow furrows (illustrated in Fig. 9a), which are about 7-8 m apart and separated from the toe of the colluvial fan by a trough that is up to 5 m deep measured against the crest of the proximal ridge (see also Fig. 6a). Elsewhere, a single broad ridge occurs along parts of the edge of the bench. Proximal troughs are less well developed or absent away from the cross-profile: the absence of troughs and ridges likely indicates where past snow-avalanche activity has overtopped the landform and may have eroded ridges, filled in troughs and/or prevented their formation. Another distinctive minor feature associated with the area of transverse ridges is upstanding elongate boulders shown in Fig. 9b: these were not observed on any of the other landforms in the cirque.

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

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

Discussion

Pronival rampart development at Smørbotn 2 and 3

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

These landforms comprise single linear to sub-arcuate main ridges or ramps, which lie close to the foot of the bedrock backing cliff. The maximum ridge-crest to cliff-foot distance of 30 m places the whole of both landforms well within the roughly estimated 30-70 m backwall to ridge crest distance necessary for a snowbed that is steep enough for supranival sediment transport, which probably requires slope angles >20º, yet thin enough to avoid conversion of snow to glacier ice, which may require >30 m snow thickness (Ballantyne and Benn 1994; Benn and Evans 2010). Ridge and ramp morphology is consistent with snowbeds repeatedly forming in the same place, though snowbed dimensions vary annually, particularly in response to the extent of snow-avalanche activity, as has been observed at these sites (Fig. 10a).

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

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

Although coarse sediment textures, predominantly angular clasts and associated openwork fabrics have been widely noted in relation to pronival ramparts and proposed by some as diagnostic (e.g. Washburn 1979; White 1981; Ballantyne and Kirkbride 1986; Ballantyne 1987; Hedding et al. 2007) other pronival ramparts contain diamictons with an appreciably fine matrix consistent with the sediments at Smørbotn 2 and 3 (cf. Harris 1986, Hall and Meiklejohn 1997). The former type of sediments is likely where coarse rockfall debris generation by frost weathering is the dominant process. This would explain the angular mode in clast roundness data but not the abundant fine matrix. However, where snow-avalanche deposition is more important, as in Smørbotn, the proportion of fines may be accounted for by the nature of the debris being removed from the backwall, which could have been concentrated further by supranival wash (Derbyshire et al. 1979) and deposited and reworked in subnival or pronival positions by streams, debris flows and/or solifluction (Shakesby 1997).

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

The pronival rampart/moraine transition at Smørbotn 1

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

Another distinctive feature of the main ridge is its well-developed, steep proximal slope: the ridge takes on a ramp-like form only at its southern extremity. A steep proximal slope may conceivably be produced by a snow-push mechanism, as advocated by Shakesby et al. (1999) for the minor ridges on the ramp-like section. However, its scale is more likely to be accounted for by a combination of glacial erosion, push and bulldozing by a rigid glacier tongue moving over bedrock slabs in its upper part and diamicton nearer to its terminus. Although clast roundness at most of the sites investigated at Smørbotn 1 (Shakesby et al. 1995) exhibits an angular mode, the substantial proportion of subangular clasts is consistent with subglacial abrasion (Boulton 1978; Lukas et al. 2013), which is likely to be effective even beneath very small glaciers (*cf*. Matthews 1987). Thus, we suggest that moraine ridge formation at Smørbotn 1 involved subglacial as well as supraglacial sediments. Furthermore, rather than a 'protalus moraine' of the type envisaged by Ballantyne and Benn (1994), which is simply the product of the deposition of supraglacial debris by a dumping process, we propose a 'push moraine' origin that is built up incrementally and incorporates a wider range of sedimentary material. Pushing (bulldozing) of proglacial sediments seems to be the dominant mechanism of recent moraine ridge formation at warm-based (temperate) glaciers at relatively low altitudes in southern Norway (Winkler and Matthews 2010).

Shakesby et al. (1999) argued strongly that Smørbotn 1 was a pronival rampart produced by a snowbed rather than a moraine ridge produced by a glacier. Based largely on similarities to the adjacent pronival ramparts, snowbed thickness and snow density, they concluded that the minor ridges could be accounted for by the snow-push mechanism and that this mechanism could be scaled-up to account for the main ridge. However, the maximum ridge-crest to cliff-foot distance of 100 m lies beyond the lower limit normally regarded as necessary to produce sufficient snow thickness for the transformation of snow to ice. Snow densities measured at the site by Shakesby et al. (1999) of up to 900 kg m^{-3} exceed the value of 830 kg m^{-3} indicative of ice rather than firn (Benn and Evans 2010). The maritime location of Smørbotn is, moreover, conducive to melting and refreezing of snow, which would allow snow to glacier ice transformation exceptionally rapidly, perhaps at snow thicknesses as little as 13 m (Paterson 1994; Benn and Evans 2010). In addition, accumulation of snow from snow-avalanches is likely to produce much larger and thicker snowbeds than those observed by us or shown on aerial photographs taken in mid- to late summer. Indeed, evidence of the former existence of a thicker snowbed and/or glacier is provided by the relatively light colouration of the bedrock (indicative of clean rock surfaces devoid of lichens) that can be seen well above the snowbed edge in Figs 3a and 4a. This light zone implies that extremely large snowbeds are banked against the backwall early in the ablation season. Furthermore, snow-avalanche activity may have been greater, and/or the snowbeds may have persisted for longer, in the recent past (see below).

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

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

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

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

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

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

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

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

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

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

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

For rock glacier development to have occurred within such a relatively short time interval as a few thousand years, environmental conditions must have been very different from today. Debris supply rates would have been considerably higher during and immediately after deglaciation, when bedrock instability and colluvial activity are likely to have been triggered by glacier debuttressing (Jarman 2006; McColl 2012; McColl and Davies 2012) and permafrost degradation (Gruber et al. 2004; Gruber and Haeberli 2007; Deline et al. 2015). Permafrost is likely to have been ubiquitous during deglaciation, both on the nunataks and in the bedrock beneath the cold-based ice sheet (Sollid and Sørbel 1994; Kleman and Hättestrand 1999; Kleman and Glasser 2007; Goehring et al. 2008). Residual permafrost can be hypothesized as contributing to the necessary ice content for rock glacier creep at Smørbotn 7 and 8. Stabilization of the rock glaciers and the surrounding slopes would have accompanied continuing permafrost thaw. Interestingly, the return of a permafrost climatic regime in the Younger Dryas appears not to have produced a major resurgence in rock glacier development, though activity probably continued, at least at Smørbotn 8. Thus, in terms of debris supply, Smørbotn 7 and 8 are considered to be primarily paraglacial and/or paraperiglacial landforms dependent on complex interactions between paraglacial and/or paraperiglacial slope processes shortly after deglaciation (*cf.* Ballantyne 2002; Mercier 2008). In terms of rock glacier movement, however, they are the product of periglacial processes.

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

The periglacial-glacial landform continuum and landform transitions

Shakesby et al. (1987) introduced the idea of a morphological and developmental continuum of large scale talus-derived landforms in Rondane, southern Norway, which extended from talus slopes to pronival ramparts, rock glaciers and ice-cored 'push-deformation' moraines. In their schematic model these landforms are characterised by different transport processes (sliding of debris over snow, gravity deformation or glacier ice-push deformation, respectively) and develop along separate pathways leaving little room for transitional forms between the recognized landform types (see also Shakesby et al. 1989). In a critical appraisal of this model, Kirkbride (1989) replaced rockfall/talus with debris supply from a broader range of debris sources, and focused on the relative importance of fluxes of snow/ice and debris as the primary determinant of the landform types (see also the related concepts and models of Giardino and Vitek 1988; Whalley 2009; and Janke et al. 2013). Thus, changes in these fluxes through time (especially due to climate change) and over space (due to local site characteristics) can be seen as important dimensions of the landform continuum, and transitions of one landform type to another are accommodated. These ideas may be taken further in the light of the pronival ramparts and related landforms in Smørbotn, resulting in an improved conceptual model of part of the periglacial-glacial landform continuum and containing potential transitional landforms (Fig. 12).

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

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

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

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

Diagnostic criteria revisited

The existence of transitional landforms adds to the difficulties in differentiating pronival ramparts from related landforms. In the light of the landform transitions identified in Smørbotn and analysed within the conceptual framework of Fig. 12, none of the diagnostic criteria listed for pronival ramparts by Shakesby (1997) or Hedding and Sumner (2013), can be regarded as definitive. Single morphological or sedimentological criteria are certainly not universally applicable and, even in combination, may be misleading. It should be possible instead, to identify each type of landform using a combination of criteria as developed in this paper. These criteria relate to morphological and sedimentological characteristics that take account of the local environment and landscape context, including land-surface processes, climatic regime, and landform age and development.

Conclusions

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

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

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

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

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

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

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

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Figure captions

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

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

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

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

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

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

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

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

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

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

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

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

Table 1. Schmidt-hammer exposure-age dating of the northern (North) and southern limbs (South) of Smørbotn 8 and the associated colluvial fan toe (*Cs* = sampling error; *Cc* = calibration error)

Site	Mean R-value	Standard Deviation (σ)	Confidence Interval (95%)	Sample size (n)	*Cs* (years)	*Cc* (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