Holocene alluvial fan evolution, Schmidt-hammer exposure-age dating and paraglacial debris floods in the SE Jostedalsbreen region, southern Norway

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The evolution of several subalpine alluvial fans SE of the Jostedalsbreen ice cap was investigated based on their geomorphology and Schmidt-hammer exposure-age dating (SHD) applied to 47 boulder deposits on the fan surfaces. A debris-flood rather than debris-flow or water-flow origin for the deposits was inferred from their morphology, consisting of low ridges with terminal splays up to 100 m wide without lateral levees. This was supported by fan, catchment, and boulder characteristics. SHD ages ranged from 9480±765 to 1955±810 years. The greatest number of boulder deposits, peak debris-flood activity and maximum fan aggradation occurred between ~9.0 and 8.0 ka, following regional deglaciation at ~9.7 ka. The high debris concentrations necessary for debris floods were attributed to paraglacial processes enhanced by unstable till deposits on steep slopes within the catchments. From ~8.0 ka, fan aggradation became progressively less as the catchment sediment sources tended towards exhaustion, precipitation decreased during the Holocene Thermal Maximum, and tree cover increased. After ~4.0 ka, some areas of fan surfaces stabilized, while Late-Holocene climatic deterioration led to renewed fan aggradation in response to the neoglacial growth of glaciers, culminating in the Little Ice Age. These changes are generalized within a conceptual model of alluvial fan evolution in this recently-deglaciated mountain region and in glacierized catchments. This study highlights the potential importance of debris floods, of which relatively little is known, especially in the context of alluvial fan evolution.

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Alluvial fans are fan-shaped depositional landforms created where steep, high-powered channelized flows deposit their material load on entering a zone of flow expansion and reduced power (Harvey 2004; Owen 2014; Ventra & Clarke 2018). Typically, they are classified according to the predominant depositional process, into (i) fluvial fans, where stream flows (water flow or water flood) deposit bedload, and (ii) colluvial fans dominated by mass movement processes, particularly debris flow (also known as gravity-flow fans) (Rachocki & Church 1990; Crosta & Frattini 2004; Harvey *et al.* 2005; De Haas *et al.* 2015, 2019; Bowman 2019). Whereas most studies have emphasised these two types of alluvial fans, there is increasing recognition of the existence of a continuum of landforms, which reflect interactions between processes and intermediate-type flows (Wells & Harvey 1987; Hungr *et al.* 2001; Germain & Ouellet 2014). Terms for the flows that are intermediate in character between water floods and debris flows include fluid (wet or watery) debris flows (Sletten & Blikra 2007; Harvey *et al*. 2013), hyperconcentrated flows (Matthews *et al*. 1999; Pierson 2005; Sletten & Blikra 2007; Calhoun & Clague 2018), debris torrents (Slaymaker 1988) and debris floods (Hungr *et al*. 2001; Wilford *et al.* 2004; Mayer *et al*. 2010; D'Agostino 2013; Ouellet & Germain 2014). However, the nature of these flows, which are characterised by sediment concentrations of 40–70% by weight according to Costa (1984), and their role in fan development, are still poorly understood.

In order to understand better the development of alluvial fans, the long-standing problem of precise numerical dating (of the fan surface) needs to be overcome. Several techniques ranging from historical analysis to dendrochronology and lichenometry have been applied to the dating of fan development over annual to decadal timescales (e.g. D'Agostino 2013; Jomelli 2013; Schneuwly-Bollschweiler & Stoffel 2013; Stoffel 2013). Far fewer techniques, including those based on radiocarbon, optically stimulated luminescence (OSL) and terrestrial cosmogenic nuclides are applicable over longer, centennial to millennial timescales (e.g. Harvey *et al.* 2005; Schneuwly-Bollschweiler *et al.* 2013; Schürch *et al.* 2016). Here we apply the relatively new technique of Schmidt-hammer exposure-age dating (SHD) to fan surfaces. SHD is appropriate for providing numerical ages for boulders exposed during the Lateglacial and Holocene (see, for example, Winkler 2009; Matthews & Owen 2010; Shakesby *et al.* 2011; Matthews *et al*. 2013, 2015, 2018; Stahl *et al*. 2013; Tomkins *et al*. 2016, 2018; Wilson & Matthews 2016; Winkler *et al*. 2016; Wilson *et al*. 2019).

Various temporal patterns and activity phases have been recognised in records of floods, debris flows and other colluvial processes (ranging from snow flows to rock falls) in southern Norway (Blikra & Nesje 1997; Blikra & Nemec 1998; Blikra & Selvik 1998; Sletten *et al.* 2003; Bøe et al. 2006; Sletten & Blikra 2007; Matthews *et* *al.* 2009, 2018; Vasskog *et al.* 2011). Detailed case studies of two alluvial fans have, moreover, revealed contrasting histories. Radiocarbon dating and lichenometry show that development of the subalpine Nystølen fan in the Jostedalsbreen region (Lewis & Birnie 2001; McEwen *et al.* 2011) was dominated by deposition in the Little Ice Age of the last few centuries, whereas SHD shows that the alpine Illåe fan in Jotunheimen is largely a relict paraglacial landform that developed before ~8.0 ka (McEwen *et al.* 2020). Differences in the evolution of these two fans were accounted for largely by the extent to which their catchments were glacierized in the past.

In this study, the aim is to generalize further by analysing the development of subalpine alluvial fans in the SE Jostedalsbreen region of southern Norway (Fig. 1), based on their geomorphology and the exposure age of their surface boulder deposits. There are three main objectives: (i) To date the numerous boulder deposits on the fan surfaces using SHD and hence provide a firm chronology; (ii) To assess the origin of the boulder deposits with reference to processes of debris flow, water flow (floods) and debris floods; and (iii) To reconstruct the evolution of several fans and hence develop a regional conceptual model of fan evolution in recently-deglaciated mountain catchments.

Study sites and environment

The alluvial fans are located to the SE of the Jostedalsbreen ice cap on valley floors at 300–400 m above sea level at the foot of steep tributary valleys descending from a glacierized plateau at >1600 m a.s.l. (Fig. 2). Four fans, from south to north, are termed here: (i) the Erikstølsdalen fan; (ii) the Kvamsdalen fan; (iii) the Snøskreda fan; and (iv) the Kupegjelet fan, the latter two being located in Austerdalen. Kupegjelet fan (Fig. 3), in many ways similar to the others, was previously investigated by Innes (1985a, b). Data from a fifth fan (Nystølen fan) in lower Langedalen (Fig. 2), previously investigated by Lewis & Birnie (2001) and McEwen *et al.* (2011) are included in some of our analyses. The four fans (Fig. 4) were selected because of their extensive boulder deposits suitable for dating by SHD, using field and aerial photographic evidence. Other neighbouring fans were unsuitable: those south of Veitastrond have been greatly modified by land clearance, while those further north in Austerdalen (e.g. at the mouth of Røysedalen) have been eroded by the historical advance of the glacier Austerdalsbreen.

All five fans are subalpine in character: the Kvamsdalen and Kupegjelet fans are largely covered in *Betula pubescens* woodland, whereas the Erikstølen, Snøskreda and Nystølen fans have much larger areas of grassland, which are partly the result of snow-avalanche activity, and partly a response to grazing animals associated with the agricultural settlement of Veitastrond and sæters such as Tungestølen. Climatic data from the meteorological station Bjørkhaug, in the neighbouring valley of Jostedal (324 m a.s.l.), indicate a mean annual air temperature of +3.7 °C, with a July mean of +13.4 °C, a January mean of –4.9 °C and a mean annual precipitation of 1380 mm (Aune 1993; Førland 1993). The local lithology is predominantly granite with some areas of granitic gneiss (Lutro & Tveten 1996).

Morphometric data from the fans, their catchments, and their surface boulder deposits are summarised in Table 1. The five catchments are small (1.17–3.44 km^2), high relief, steep and rugged, with a Melton ratio (relief/√area; e.g. Melton 1965) of 0.70–1.08. The fans are correspondingly small (0.16–0.51 km^2) with gradients of 9–17°, but fan toes have been truncated by the main river or obscured by the growth of peat mires. The boulder deposits on the fan surfaces form broad, irregular ridges, up to 200 m in length with a mean width of 24–34 m (maximum width 100 m), most with terminal splays, some with finger-like extensions (Fig. 5).

Three of the investigated catchments are currently 8–24% glacierized by the Kvitekoll ice cap (Fig. 1) which, together with the Tverrdalsbreen glacier, occupy the plateau and extend onto the lee-slopes to the east. The catchment of the Nystølen fan is 56% glacierized. However, all catchments have late-lying snowbeds on their upper slopes, and are likely to have been affected by expanded plateau glaciers during Late-Holocene neoglaciation and especially in the Little Ice Age.

Rapid Early-Holocene deglaciation of the main valleys of SE Jostedalsbreen occurred during the Preboreal, and by ~10.1-9.7 ka glaciers had receded to the valley heads, close to their Little Ice Age limits (Dahl *et al.* 2002; see also Mottershead &

Collin 1976; Aa 1982; Nesje 1991, 2009). Subsequently, further rapid warming and glacier shrinkage resulted in the total melting of the Jostedalsbreen ice cap by ~7.3 ka (Nesje & Kvamme 1991; Nesje *et al*. 2000, 2001). Centennial- to millennial-scale glacier variations interrupted neoglacial re-growth of glaciers after ~6.1 ka, which culminated in the Little Ice Age maximum of extant glaciers around AD 1750 (Grove 1988; Bickerton & Matthews 1993). Latero-terminal moraines indicate the down-valley limits of several of these glaciers in the Little Ice Age (Fig. 2).

Methodology

Field research focused on 47 boulder deposits, which are located on Fig. 4A-D. These represent integral geomorphological units each of which can be attributed to single depositional events. They also represent the entire statistical population of boulder deposits from each fan. SHD was carried out on these deposits, supplemented by lichenometric dating and measurements of boulder roundness and boulder size.

SHD

As a basis for SHD dating, R-values were recorded from a minimum of 100 boulders on each deposit (one impact per boulder) using a mechanical N-type Schmidt hammer (Proceq 2004). Use of one impact per boulder ensured that the R-value frequency distribution approximates the boulder-age distribution (Matthews *et al*. 2014). In order to minimise variability and measurement errors, small or unstable boulders, edges, joints and cracks, and lichen-covered or wet boulder surfaces were avoided, and measurements were confined to near-horizontal surfaces and granitic lithologies (cf. Shakesby *et al*., 2006; Matthews & Owen, 2010; Viles *et al*., 2011). No cleaning or artificial abrading of the boulder surfaces was carried out prior to measurement as this would have removed age-related weathering effects. The Schmidt hammer was regularly tested on the manufacturer’s test anvil during the fieldwork to ensure no deterioration in instrument performance following prolonged use (cf. McCarroll 1987, 1994).

Calibration of R-values followed the approach developed by Matthews & Owen (2010), Matthews & Winkler (2011) and Matthews & McEwen (2013), full details of which are given in Matthews et al. (2018). The calibration equation is a linear regression of surface age (y) on mean R-value (x) derived from two local control points: 'old' and 'young' surfaces of known age. Use of a linear relationship has been specifically tested over the Holocene timescale (Shakesby *et al.* 2011), and is justified also by comparison over similar relatively short timescales with terrestrial cosmogenic nuclide dating both in southern Norway (Wilson *et al.* 2019) and elsewhere (e.g. Tomkins *et al.* 2016, 2018). A linear or near-linear relationship, which results from the slow rate of chemical weathering of rock surfaces, is therefore considered appropriate, particularly in alpine and subalpine environments over the last ~10 ka.

Confidence intervals (95%) for SHD age (C_t) are based on combining the relatively small error term associated with the calibration equation (C_c) with the larger sampling error associated with the dated surfaces (C_s). Uncertainty associated with C_s is relatively small provided: (i) very large R-value sample sizes are used for control points; and (ii) control-point ages are accurately known. Here we used 600-750 R-values for each control point and hence can justify using precise ages for the control points.

The 'young' control point involves R-values from 600 boulders (one impact per boulder) deposited on the Erikstølsdalen and Snøskreda fans (Fig. 2) during a flash flood following intense rainfall on 14 August 1979 (cf. Gjessing & Wold 1980; Drageset 2001). Both the geomorphological integrity and lichen sizes associated with the flood deposits leave no doubt that the surface boulders sampled are representative of a synchronous surface and that their age is very tightly constrained. The rockfall deposits used previously by Matthews & Wilson (2015) as their 'young' control point were deemed unsuitable for the present study due to the roughness characteristics of such colluvial boulders (cf. Matthews & McEwen 2013; Matthews *et al*., 2015; Olsen *et al*., 2020). In contrast, the 1979 flood deposits, being characterised by relatively smooth boulders, have similar roughness to the boulder deposits on the fans, and are therefore appropriate for a study of alluvial fans.

The 'old' control point, involving 750 R-values recorded from three glacially-scoured bedrock outcrops near Tungastølen and at the mouth of Kvamsdalen (Fig. 2), was used previously by Matthews & Wilson (2015). The precise date of ~9.7 ka used for this control point is based on the age of moraine ridges deposited by Jostedalsbreen outlet glaciers in valleys on both sides of the ice cap. Evidence for the age of these moraines comes from both radiocarbon (Nesje 1984; Dahl *et al*. 2002) and cosmogenic nuclide dating (Matthews *et al.* 2008) in Erdalen on the NW side of the ice cap, and by radiocarbon dating near Nigardsbreen in Jostedalen on the SE side (Dahl *et al.* 2002). The moraines, which are of a similar size to Little Ice Age moraines and located up to ~1 km beyond the Little Ice Age limits of Erdalsbreen and Nigardsbreen, relate to the Erdalen Event, an Early-Holocene centennial-scale glacier and climatic fluctuation that involved two glacier re-advances dated by Dahl *et al*. (2002) to ~10.1 and 9.7 ka.

Although no similar moraines dating from the Erdalen Event occur in Austerdalen or Langedalen, the glaciers in these valleys are assumed to have fluctuated broadly synchronously with other outlet glaciers of Jostedalsbreen, as has been demonstrated for the Little Ice Age interval (cf. Bickerton & Matthews 1992, 1993). We attribute the absence of Erdalen Event moraines downvalley of the Little Ice Age glacier limits in Austerdalen or Langedalen to the presence of relatively large ice bodies in these valleys and correspondingly large glacier re-advances during the Erdalen Event. Combined with the occurrence of this event during an otherwise prolonged period of rapid glacier retreat, we conclude that the 'old' control surfaces in the study area were deglaciated closely following the termination of the Erdalen Event (i.e. ~9.7 ka).

Probability density function analyses were used to understand the SHD age-frequency distributions over the Holocene timescale. Separate analyses were carried out for each fan and for the combined data set. Probability density was calculated at 100-year intervals using the mean and standard deviation for each fan (R Core Team 2019). Calculation assumed a normal distribution of the data. Probability density functions for each of the four alluvial fans were obtained by averaging the density values of the relevant individual boulder deposits. A regional density function was obtained by averaging the density functions of all 47 boulder deposits from the four fans.

Supplementary measurements

Three types of supplementary measurements were made from the boulder deposits. First, the maximum diameter (longest axis) of the 5 largest lichens of the *Rhizocarpon* subgenus from each of the 47 deposits was measured in order to perform lichenometric dating. This had been previously attempted by Innes (1985a) for a fan in Austerdalen. We used updated lichenometric dating curves from the neighbouring glacier foreland of Nigardsbreen (Bickerton & Matthews 1992, 1993). Second, in order to assess potential sediment sources, the boulder roundness distribution and a numerical index of mean boulder roundness were derived from a subsample of 25 boulders from 37 of the boulder deposits using the Powers (1953) roundness chart (cf. Matthews 1987). Finally, the maximum intermediate-axis clast size (*d*) from each boulder deposit was measured to allow the calculation of palaeohydrological parameters associated with the flows that deposited the sediment (Williams 1983):

$$\text{Unit stream power } (\omega) = 0.079\, d^{1.3}\ (10 \leq d \leq 1500 \text{ mm}) \qquad (1)$$

$$\text{Bed shear stress } (\tau) = 0.17\, d^{1.0}\ (10 \leq d \leq 3300 \text{ mm}) \qquad (2)$$

$$\text{Mean flow velocity (V)} = 0.065\, d^{0.50}\ (10 \leq \mathrm{d} \leq 1500 \text{ mm}) \qquad (3)$$

Results

R-values from control surfaces and calibration equations

Combined data for the ‘old’ and ‘young ‘ control points (Table 2) show excellent agreement between each pair of ‘old’ and ‘young’ control surfaces, which justifies treating each pair of surfaces sampled from different locations as replicates drawn from the same statistical population. The R-value distributions of the control points (Fig. 6A) exhibit the symmetrical, unimodal characteristics of synchronous surfaces. Furthermore, the small standard deviations ($\sigma = 6–8$) relative to the standard deviations associated with the fan surfaces ($\sigma = 8–11$; Table 3), together with wide separation of the mean R-values, signal the potential for dating using the calibration equation shown in Fig. 6B.

R-values and SHD ages from the boulder deposits

R-value distributions for 47 boulder deposits are generally symmetrical and unimodal, which is again indicative of synchronous surfaces produced here by single depositional events (Fig. 6). Whereas mean R-values (Table 3) vary widely between 38.5 (Sa 4) and 58.8 (En 7) most are closer to the characteristic of the 'old' control point than to those of the 'young' control point. SHD ages are correspondingly wide ranging but with a large majority of the boulder deposits dating from early in the Holocene (>70% before ~6.0 ka) and only two dating from the last 2.0 ka (Table 3). The sampling error (Cs) resulting from the high natural variability of weathered boulder surfaces is the dominant control on the 95% confidence intervals for age, which range from ~700-900 years.

SHD ages and probability density distributions indicate significant clustering of events and notable similarities and differences between the chronology of boulder deposits from each fan (Fig. 8), which are discussed below. Amalgamation of the age data from the four fans in the combined record emphasises the overall concentration of dates shortly after deglaciation and the long-term declining frequency of depositional events through the Holocene.

Lichen sizes and lichenometric ages

Mean lichen size on the boulder deposits varies from ~50–300 mm, which corresponds to a lichenometric age of ~70–1500 years (Fig. 9). At three sites from Kupegjelet and one from Snøskreda, single largest lichens reached 300–320 mm, which are comparable to the largest lichens (270–290 mm) measured from the same sites by Innes (1985). As a result of using the up-dated calibration equation of Bickerton & Matthews (1992, 1993), our results suggest that >50% of deposits are characterised by mean lichen sizes >150 mm and date from pre-Little Ice Age times. In contrast, Innes (1985) concluded that all the deposits fell within the Little Ice Age. However, as there is no correlation between our SHD and lichenometric ages, it can be deduced that most of the latter are large underestimates of the true age of the deposits, resulting from extrapolation of surface ages beyond the range of lichenometric dating, combined with the limited lifespan of lichens of the *Rhizocarpon* subgenus in this environment (cf. Matthews & Trenbirth 2011).

Boulder characteristics and palaeohydrological parameters

Mean boulder roundness from boulder deposits across all four fans lies consistently between values characteristic of sub-angular and sub-rounded clasts (Fig. 10). Furthermore, the majority of sites (70%) have a sub-rounded modal class with a negligible proportion of angular and very angular clasts. These roundness values are consistent with a till source for the boulders and are consistent with a relatively small degree of abrasion during transportation in relatively small catchments (cf. Boulton 1978; Matthews 1987; Evans & Benn 2004).

Maximum boulder size and the median size (D_{50}; intermediate axis) of the 10 largest boulders from the boulder deposits (Fig. 11) approximate to 2.0 m and 1.0 m, respectively. The largest boulder (2.5 m) occurred at Erikstølsdalen (En 1) while the largest D_{50} (1.7 m) was recorded from Kupegjelet (Kt 8). Such sizes require high competent flows and imply major floods as large as the 1979 flash-flood event, the deposits of which, on the Snøskreda fan, involved maximum boulder sizes and D_{50} values of 2.1 m and 1.2 m, respectively. The flash flood had a return period of ~1 in 100 years, estimated from its magnitude on the main river in Jostedal (Gjessing & Wold 1980; Drageset 2001), but this may have increased to ~1 in 1000 years in smaller catchments within the region (Matthews & McEwen 2013).

Minimum boulder-transport conditions for the largest boulder in deposits for each fan are summarised in Table 4. Lowest unit stream power for entrainment (ω) varies from 1646 W m^{-2} (Kvamsdalen) to 2065 W m^{-2} (Erikstølsdalen), lower than values for the coarsest debris-flood deposits on the upper Illåe fan, Jotunheimen (up to 2850 W m^{-2}; McEwen *et al*. 2020). A large number of deposits (47%) have largest clasts beyond the upper range of clast sizes used by Williams (1983). Lowest bed shear stresses for entrainment (τ) of the largest clast ranged from 357 to 425 N m^{-2}.

Discussion

Water floods, debris flows or debris floods?

The morphology and sedimentary characteristics of the boulder deposits, and the characteristics of the fans and their catchments, tend to be intermediate in terms of established criteria for recognising the products of water flow and debris flow (Table 5). Using all these criteria, the boulder deposits can be attributed with some confidence to debris floods, which are now recognised in the most widely-used genetic classification of landslide types (Hungr *et al*., 2014).

Previous studies by Innes (1985a, 1985b) assumed that these boulder deposits were debris-flow lobes, which also tend to have boulder concentrations in their terminal areas and on lateral levées. However, their morphologies differ from debris-flow lobes in several respects. They are commonly irregular, broad ridges, which are raised above the general level of the adjacent fan surfaces by up to several metres (Fig. 5A, B). With a mean width of 24-34 m and a maximum width of up to 100 m (Table 1), they are considerably wider than typical debris-flow deposits and, crucially, levées are absent. They terminate in several different plan shapes ranging from simple, steep-fronted tongues (similar to debris-flow lobes) to single or multiple splays (less thick as well as wider than debris-flow lobes), the latter sometimes with finger-like extensions (Fig. 5C-F). Similar broad ridges without levées occur in Iceland, where they were described as 'debris flow-like' (Decaulne *et al*. 2007). Debris flood ridges and splays also differ from the thin gravel sheets with bars and braided channels deposited on fans by water floods.

Neither the size nor slope of each of our fans, nor their catchment characteristics, are typical of either debris-flow fans, which are smaller and steeper with very small rugged catchments, or fluvial fans, which are generally larger with gentler slopes and much larger catchments. Although no sections were available through these deposits, the surface sediments of the ridges appear intermediate between unsorted diamictons and well-sorted fluvial deposits. The sediments also seem to correspond to the proximal facies of terminoglacial fans described by Zieliński & van Loon (2000), which include boulder-rich diamictons and sandy gravels deposited by sheetflows and catastrophic hyperconcentrated flows. There is little evidence of fine matrix where boulder concentrations occur, but this could have been washed out of the surface material during or after deposition. The stratigraphy of the Illåe fan (Jotunheimen), where similar boulder deposits occur, revealed a variable content of matrix with alternating, crudely-sorted and generally indistinct clast-supported and matrix-supported layers (McEwen *et al*. 2020).

In order to achieve the high debris concentrations necessary for debris floods, with sufficient large subrounded to subangular boulders (Figs 10, 11), the catchment would have had to contain a suitable sediment source. This is likely have been a till cover, deposited prior to ~9.7 ka, when the catchment was completely glacierized. We argue below that all four catchments had a substantial till cover, which was exposed to subaerial processes following deglaciation. This till cover would have been readily eroded from the steep slopes of the catchments, and provides the likely source of the sediments in the debris-flood boulder deposits.

The chronology of events

The chronology of boulder deposits from each fan (Fig. 8) shows that the earliest depositional events occurred shortly after deglaciation at ~9.7 ka. Indeed, the oldest SHD dates from Snøskreda (Sa 4) and Erikstølsdalen (En 5) are 9480±765 and 9215±720 years, respectively which are statistically indistinguishable (along with several other SHD dates) from 9.7 ka. Taking account of the confidence intervals, both of these fans have a very high proportion of SHD ages >8.0 ka, while all the boulder deposits on the Kvamsdalen fan have SHD ages >4.0 ka and the Kupegjelet fan has a relatively high proportion between 8.0 and 6.0 ka. Three fans developed rapidly within two millennia of deglaciation while the fourth (Kupegjelet) appears to have started its rapid development about two millennia later than the others. All four fans therefore underwent major aggradation attributable to debris-flood activity during the Early Holocene.

Persistence of so many boulder deposits from the Early Holocene is indicative of the decline in the frequency of debris-flood events later in the Holocene. If the frequency of such events had remained high during the Middle and Late Holocene, fan aggradation in the form of debris-flood deposits would have continued into the Late Holocene and earlier deposits would have been buried by later ones. Instead, a small number of debris-flood deposits with ages <4 ka occur only at the Erikstølsdalen and Kupegjelet fans. At the Kvamsdalen and Snøskreda fans, debris-flood deposits are confined to distal and marginal parts of the fans. In proximal- and mid-fan locations, however, these two fans exhibit evidence of late-Holocene and modern aggradation from snow-avalanche and fluvial activity, which may have buried earlier boulder deposits.

The combined chronology from the four fans (Fig. 8) suggests a relatively steady decline in frequency of debris-flood events from a maximum at ~9.0–8.0 ka. However, the interpretation is complicated by wide confidence intervals for SHD age of the order of 700–900 years, the apparent absence of events between deglaciation and ~8.0 ka at the Kupegjelet fan where activity peaks at ~7.0 ka, and the possibility of centennial- to millennial-scale variations in aggradation in the Middle to Late Holocene.

Holocene development of the alluvial fans and their environmental controls

The peak in debris-flood activity immediately following deglaciation at ~9.7 ka is clearly indicative of a paraglacial pattern of sediment deposition conditioned directly by glaciation (cf. Ryder 1971; Church & Ryder 1972; Ballantyne 2002a, 2013). During and immediately following deglaciation, the till deposits on the extremely steep slopes of these catchments would have been particularly susceptible to gully erosion triggered by rainstorms, and glacial and snow meltwater (Curry 2000). Being a diamicton, the till would have been a source of abundant large subangular to subrounded boulders and fine matrix, providing the high sediment concentrations for debris-floods. These flows would have been confined in the narrow, steep tributary valleys before they debouched onto the main valley floor where redeposition and fan aggradation occurred.

Although paraglacial processes are most effective in the unstable landscape that emerges during deglaciation, paraglacial effects may last for several millennia until the landscape stabilises or sediment sources are exhausted (Ballantyne & Benn 1994; Curry 1999; Ballantyne 2002b). A steady decline in the frequency of debris-flood deposits over the first few millennia following deglaciation (Fig. 8) might therefore be accounted for simply in terms of paraglaciation. Furthermore, the exhaustion of accessible sediment sources is a distinct possibility on extremely steep slopes, particularly at relatively high altitudes in all four catchments where extensive bedrock exposure is evidence of a more-or-less completely eroded, former till cover.

The Jostedalsbreen ice-cap, along with the Kvitekoll ice cap and the other glaciers that directly affected the catchments of the alluvial fans, are inferred to have melted away completely by ~7.3 ka (Nesje & Kvamme 1991; Nesje *et al*. 1991, 2000, 2001). This date coincides with the rapid development of the Kupegjelet fan which, according to our SHD dates, occurred up to two millennia later than at the other three fans. A possible explanation for later development at Kupegjelet is the survival of glacier ice for longer in its narrow catchment and in the steep, north-facing cirque-like extension to the valley head on its south side. In much the same way, the north-facing valley head of Røysedalen is currently occupied by the northern outlet glacier of the Kvitekoll ice cap (Fig. 2). A second possible explanation is that rapid fan development at Kupegjelet was triggered by the paraperiglacial degradation of permafrost in the upper catchment: i.e. it was a conditional response to the transition from permafrost to seasonal-freezing regime (cf. Mercier 2008; Scarpozza 2016; Matthews *et al.* 2018). The lower altitudinal limit of discontinuous mountain permafrost currently lies at ~1600 m a.s.l. in the Jostedalsbreen region, and could be lower in north-facing rock walls (Etzelmüller & Hagen 2005; Gisnås *et al*. 2016; Steiger *et al*. 2016).

Landscape stabilisation and hence reduced paraglacial aggradation on the fans are likely to have been accentuated by the spread of a dense tree cover onto the lower-altitude slopes of the catchments in the Early to Middle Holocene as a result of a warmer climate than today during the Holocene Thermal Maximum (HTM). Present-day tree lines attain altitudes of 850–1000 m at favourable locations within the four catchments (https://www.norgeskart.no/) and, based on pollen analyses from the valleys around Jostedalsbreen, would have been at least 200 m higher during the HTM (Nesje & Kvamme 1991; Nesje et al. 1991; see also Wilford et al. 2005; Marston 2010; Pawlik 2013).

The SHD evidence indicates that paraglacial sedimentation was the dominant control on the Early Holocene development of all four fans. However, there was greater divergence in their evolution during the Late-Holocene: the Kvamsdalen fan seems to have become an essentially relict landform when paraglacial effects effectively ceased at ~4.0 ka; the Kupegjelet fan experienced continuing deposition from debris floods at a much reduced level until at least ~2.0 ka; the Erikstølsdalen and Snøskreda fans appear to have been dominated by a different sedimentological and hydrological regime, which began sometime after ~8.0 ka and has continued to the present day. This new regime, which is attributed to the diminished sediment supply after the cessation of the debris floods of the paraglacial phase, was dominated by snow-avalanches and fluvial activity, and has left boulder deposits unburied at the margins of these fans. Evidence of the importance of snow avalanching at these sites includes the presence of extensive accumulations of snow on the fan apex and upstream, which are clearly visible on late-summer aerial photography (https://www.norgeibilder.no/), isolated angular boulders scattered over the fan surface, and the names 'Snøskreda' (which means snow avalanche in Norwegian) and 'Erikstølskreda', which are established place names used on topographic maps. Fluvial activity is indicated by gravel deposits alongside the current stream, largely vegetated distributary channels, and the boulder-rich sediments deposited by the AD 1979 flash flood.

Neoglaciation from ~6.1 ka and Late-Holocene glacier variations appear to have made significant contributions to the later phases of fan evolution, particularly at the Erikstølsdalen and Kupegjelet sites. Based on moraines dated by historical evidence and lichenometric dating, it is well established that the main glaciers in this region, including Austerdalsbreen and Nystølsbreen (Fig. 2) attained their Late-Holocene maximum extent *c.* AD 1750, in the Little Ice Age (Bickerton & Matthews 1993) and, in the case of Nystølsbreen, the glacier extended onto its fan (McEwen *et al.* 2011). Similar undated moraines in Røysedalen indicate that the northern outlet of the Kvitekoll ice cap expanded at this time (Fig. 2), and strongly suggest that both this ice cap and Tverradalsbreen overflowed into the fan catchments during the Little Ice Age. Although there is insufficient evidence from this study to detect any century- to millennia-scale responses, the existence of small glaciers in these catchments during the Little Ice Age and earlier neoglacial glacier expansion episodes (cf. Nesje et al. 2008; Nesje 2009; Matthews 2013) are likely to have affected meltwater discharge, slope processes and sediment loads, and hence variations in Late-Holocene fan aggradation (cf. McEwen *et al*. 2011; Laute & Beylich 2012, 2013). Similarly, changes in fan aggradation would be expected from any Late-Holocene variations in snow meltwater discharge and snow-avalanche frequency, the latter affecting the Erikstølsdalen and Snøskreda fans in particular.

A regional model of alluvial fan evolution in recently-deglaciated mountains

The evolution of alluvial fans in the SE Jostedalsbreen region – including the four fans reported in this study and the Nystølen fan investigated by McEwen *et al*. (2011) – can be generalized as a regional conceptual model (Fig. 12A–D) that includes local variations in the timing of four main phases of fan development, the changing nature and intensity of aggradational processes, variations in glacier size, and changes in the climatic and hence hydrological regime during the Holocene. This model extends and refines a previous model of alluvial fan development in glacierized catchments presented by McEwen *et al*. (2020) and makes a broader contribution to the rather limited understanding of alluvial fans in alpine and subalpine environments from various perspectives (cf. Kostaschuk *et al*. 1986; Eyles & Kocsis 1988; Derbyshire & Owen 1990; Blair & McPherson 1994; Cavalli & Marchi 2008; Korup & Clagues 2009; Schneuwly-Bollscheiler *et al*. 2013; Heiser *et al*. 2015; Tomczyk *et al.* 2019).

Phase 1: Intense paraglacial aggradation (9.7–8.0 ka). – The first phase begins immediately after deglaciation. Aggradation rapidly intensifies as gully propagation takes place in steep and initially unvegetated till-mantled catchment slopes. Peak paraglacial aggradation, on the basis of the frequency of dated debris-flood deposits from three fans (Kvamsdalen, Erikstølsdalen and Snøskreda), is placed at ~9.0 ka (Fig. 12A, B).

The start of this intense phase may be delayed by the late survival of glacier ice within the catchment (or by paraperiglacial permafrost degradation), as hypothesised for Kupegjelet. Intense paraglacial aggradation takes place not only at a time of shrinking glaciers (Fig. 12C), but also in a climatic environment of high and rising temperatures and increasing precipitation (Fig. 12D). The hydrological effect of this is likely to contribute to relatively high discharges from both glacial and snow meltwater.

Phase 2: Reduced paraglacial aggradation (8.0–4.0 ka). – The transition to a phase of reduced paraglacial aggradation is considered, on the basis of the dating evidence from Kvamsdalen and Kupegjelet, to occur no more than 2000 years after the start of the intense phase. The start of this second phase is therefore placed at ~8 ka for Kvamsdalen (Fig. 12A) and this date is also used in Fig. 12B (although this phase was delayed to ~6.0 ka at Kupegjelet). The apparent absence of debris-floods for many millennia from ~8.0 ka at Erikstølsdalen and Snøskreda is attributed to their burial by later fluvial and snow-avalanche sedimentation.

Reduced aggradation after ~8.0 ka is primarily a response to the reduced availability of sediment and the possible eventual exhaustion of sediment sources within the catchment. Three other factors are seen as contributing to increasing stability within the catchment and the reduction of aggradation on the fans. First, with glaciers very small or absent from the catchments (Fig. 12C), the paraglacial sediment load of the rivers is supplemented to a negligible extent by glaciofluvial sediments direct from glaciers. Second, stabilization increases over time with the establishment of vegetation and, in particular, with the spread of trees at relatively low altitudes within the catchments. Third, temperatures remain high while precipitation is much reduced, at least until ~6.0 ka (Fig. 12D): the climatic regime therefore suggests reduced runoff from snowmelt at this time. Diminution of paraglacial aggradation is shown in Fig. 12B to continue until ~4.0 ka, though this must be regarded as an arbitrary point on the long-term declining trend.

Phase 3: Fan surface stability (4.0–0 ka). – A phase of near-zero aggradation on the fan surface is the logical outcome of the exhaustion of sediment supply within the catchment, and is recognised at Kupegjelet from ~2.0 ka and at Kvamsdalen from

~4.0 ka (Fig. 12A). Fan surface stability may also follow from flows with decreasing sediment concentrations resulting from an increase in discharge during Late-Holocene climatic deterioration and the early stages of neoglacial glacier growth. Judged in terms of the non-existence of dated debris-flood deposits, stabilization of fan surfaces did not take place before ~4.0 ka, but evidence of older stable phases could be buried by later aggradation.

The possibility of entrenchment introduces a further complication (cf. McEwen *et al*. 2020), which may itself be initiated in response to reduced sediment loads during the phase of reduced paraglacial aggradation. In this study, entrenchment is exhibited to some extent by the modern streams on the upper (proximal) parts of each fan (Fig. 4). This helps explain the tendency to asymmetrical development, at least during the later stages of fan evolution, and hence the persistence and survival of debris-flood deposits on the north side of each fan, as well as towards each fan toe. Each stream currently discharges to the south side of the fan, topographically-controlled avulsions having followed the slope of the fan (cf. De Haas *et al*. 2019), which is in turn influenced by the direction of the trunk valley, thus diverting flows away from the north side of the fans.

Phase 4: Neoglacial re-activation (4.0–0 ka). – Re-activation takes place in the Late-Holocene in response to climatic deterioration and glacier growth, provided that sufficient sediment sources are available and accessible within the catchment. The onset of this final phase is placed at ~4.0 ka on the basis of dated debris-flood deposits at Kupegjelet and Erikstølsdalen (Fig. 12A). Small glaciers regenerating as early as ~6.1 ka (Fig. 12C), and/or the associated climatic deterioration involving decreasing temperatures and increasing precipitation (Fig. 12D), are seen as unlikely to have had a major effect on aggradation initially. By ~4.0 ka, however, as neoglaciation intensifies, increasing discharge combined with greater potential for bedload generation and transport is consistent with renewed aggradation.

Re-activation greatly increases the potential for burial of older deposits, which is inferred to account for the apparent absence of debris-flood activity after ~8.0 ka at Erikstølsdalen and Snoskreda. This argument is supported by the confinement of the debris-flood deposits at the latter fan to its extreme distal fringe, the remainder of the fan surface being affected by more recent water-flood and snow-avalanche deposits. Neoglacial re-activation associated with an increase in water-flood and snow-avalanches was even more effective at the Nystølen fan (Fig. 12A), where the whole of the fan surface dates from the Little Ice Age (Lewis & Birnie 2001; McEwen *et al*. 2011).

Conclusions

- Boulder deposits from four subalpine alluvial fans in the SE Jostedalsbreen region of southern Norway were dated using SHD, demonstrating the usefulness of this technique for establishing the exposure-age of surface boulders in the context of the evolution of alluvial fans. The 47 SHD ages were established with 95% confidence intervals of ~700–900 years and were sufficient in number to determine a chronology of aggradational events during the Holocene based on age-frequency distributions and probability density functions.

- SHD ages indicated that a major phase of alluvial fan aggradation commenced immediately following regional deglaciation at ~9.7 ka and peaked at ~9.0–8.0 ka. This is attributed to paraglacial processes within unvegetated and only partially forested catchments. On three of the fans, later aggradation failed to bury the Early-Holocene deposits, which is consistent with a regional decline in the effectiveness of paraglacial processes through the Middle Holocene. The increase in glacierization of the catchments from ~6.0 ka (neoglaciation) and especially after ~4.0 ka, which accompanied climatic deterioration and culminated in the Little Ice Age of the last few centuries, accounts for the limited number of boulder deposits and reduced aggradation over the Late Holocene. Topography of the catchments, combined with differences in the timing and extent of glaciers in the catchments during deglaciation and later neoglacial glacierization, explains the local differences in fan evolution.

- Alluvial fan aggradation and boulder concentrations on fan surfaces are commonly attributed to fluvial activity (water floods) and/or debris flows. This study highlights the potential importance of debris floods, of which relatively little is known, especially in the context of alluvial fan evolution. The morphology of the boulder deposits on our fans is distinctive, consisting of broad, low ridges with distal splays but no evidence of the levées characteristic of debris flows. The degree of boulder rounding and crude sorting present in the boulder deposits, and the catchment characteristics also point to an intermediate flow-type between water flow and debris flow. Such flows require a debris concentration of 40-70% by weight, which we argue was attained during the paraglacial reworking of till deposits in these steep catchments.

- Our results have led to the development of a conceptual model of alluvial fan evolution for glacierized catchments and recently deglaciated mountains SE of the Jostedalsbreen ice cap (Fig. 12). A phase of 'intense paraglacial aggradation' is succeeded by phases of 'reduced paraglacial aggradation', 'fan surface stability' and 'neoglacial re-activation'. The model incorporates the timing of deglaciation, subsequent glacier activity, catchment topography and vegetation cover, sediment sources and climatic changes linked to the hydrological regime, all of which are effective controls on fan aggradation. The model should be applicable to some degree in other recently deglaciated mountain regions with small, steep catchments, if only as a template for comparison.

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Author contributions. – Matthews, McEwen and Owen conceived and planned the study based on previous work and carried out the fieldwork; Matthews analysed the

Schmidt-hammer and lichenometry data, calculated the SHD ages and wrote the first draft of the paper; McEwen did the palaeohydrological calculations; Los carried out the probability density analyses. All authors contributed substantially to the final draft.

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FIGURE CAPTIONS

Fig. 1. Location of the study area SE of the Jostedalsbreen ice cap in southern Norway.

Fig. 2. The study area, the alluvial fans, their catchments, and the location of sites used as ‘young’ (Y) and ‘old’ (O) control surfaces for SHD dating. Note also the location of Fig. 4 A, B.

Fig. 3. Kupegjelet fan, Austerdalen, from opposite valley side. Note extensive boulder deposits on the fan surface, and the rugged catchment with exposed bedrock on its upper slopes.

Fig. 4. Aerial photographs of the alluvial fans flown in 2017 showing numbered boulder deposits and fan outlines. A. Kvamsdalen and Erikstølsdalen. B. Snøskreda and Kupegjelet (source: https://www.norgeibilder.no/).

Fig. 5. Boulder deposits on fan surfaces. A. Typical wide boulder ridge (Kupegjellet 17, up-slope view). B. Narrow boulder ridge (Kupegjellet 13, down-slope view). C. Typical boulder splay at the distal end of a ridge (Kupegjellet 16). D Boulder ‘fingers’ extending from the distal end of a boulder splay (Kupegjellet 10). E. Steep-fronted (lobe-like) boulder splay (Kupegjellet 8). F. Boulder ‘finger’ extending from a steep-fronted boulder splay (Snøskreda 1).

Fig. 6. A. Frequency distributions of Schmidt hammer R-values for control points of ‘old’ (unshaded) and ‘young’ (shaded) control points. B. The calibration equation and calibration curve. (A) and (B) are linked by the dashed vertical lines representing the mean R-values of the control points.

Fig. 7. Schmidt hammer R-value distributions for 47 boulder deposits from four alluvial fans: Kvamsdalen (Kn), Erikstølsdalen (En), Kupegjelet (Kt) and Snøskreda (Sa). Vertical lines indicate the mean R-values from the ‘old’ and ‘young’ control

points. Sample size (*n*) was 100 boulders for each boulder deposit, except for Kn 1-4 where $n = 150$.

Fig. 8. SHD ages and probability density distributions for boulder deposits from each fan: Kvamsdalen (A), Erikstølsdalen (B), Kupegjelet(C) and Snøskreda (D). SHD age for each boulder deposit is represented by the mean boulder exposure age (circled) and 2σ confidence interval (horizontal line). The probability density function for each boulder deposit is shown as a normal distribution; combined probability density distributions are also shown for each fan (thick black lines). In (E), the frequency of SHD ages in 500-year intervals for the combined data set is shown together with the regional probability density distribution (thick black line). Regional deglaciation followed the Erdalen Event (10.2-9.7 ka), which is shown by the shaded vertical band across all parts of the figure.

Fig. 9. Lichen size (mean of the five largest lichens) and lichenometric age for 47 boulder deposits from the four fans. Lichenometric age uses the 5.1 calibration equation of Bickerton & Matthews (1991, 1992).

Fig. 10. Mean roundness (mean of 25 boulders) for 37 boulder deposits from the four fans. Mean roundness values for sub-angular (SA, 3.0) and sub-rounded (SR, 4.0) clasts are indicated.

Fig. 11. The largest boulders for 47 boulder deposits from the four fans. A. Maximum boulder size B. Median size (D_{50}) of the 10 largest boulders.

Fig. 12. Regional conceptual model of alluvial fan evolution in recently-deglaciated mountains related to Holocene glacier and climatic variations. A. Phases of fan development since deglaciation in the Jostedalsbreen region. B. Schematic intensity of paraglacial and neoglacial drivers of aggradation. C. Generalized size of the Jostedalsbreen ice cap (based on Nesje 2001) and D. Smoothed mean annual air temperature (MAAT) and annual precipitation (AP) anomalies for the normal period AD 1961-1990 in western Norway (based on Mauri et al. 2015; Hilger 2019).