



**Small rock-slope failures conditioned by Holocene permafrost degradation: a new approach and conceptual model based on Schmidt-hammer exposure-age dating in Jotunheimen, southern Norway**

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1 Small rock-slope failures conditioned by Holocene permafrost  
 2 degradation: a new approach and conceptual model based on Schmidt-  
 3 hammer exposure-age dating, Jotunheimen, southern Norway

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 24 Rock-slope failures (RSFs) constitute significant natural hazards but the geophysical  
 25 processes which control their timing are poorly understood. However, robust  
 26 chronologies can provide valuable information on the environmental controls on RSF  
 27 occurrence: information which can inform models of RSF activity in response to  
 28 climatic forcing. This paper uses Schmidt-hammer exposure-age dating (SHD) of  
 29 boulder deposits to construct a detailed regional Holocene chronology of the  
 30 frequency and magnitude of small rock-slope failures (SRSFs) in Jotunheimen,  
 31 Norway. By focusing on the depositional fans of SRSFs ( $\leq 10^3 \text{ m}^3$ ), rather than on the  
 32 corresponding features of massive RSFs ( $\sim 10^8 \text{ m}^3$ ), 92 single-event RSFs are targeted  
 33 for chronology building. A weighted SHD age-frequency distribution and probability  
 34 density function analysis indicate four centennial- to millennial-scale periods of  
 35 enhanced SRSF frequency, with a dominant mode at  $\sim 4.5 \text{ ka}$ . Using change detection  
 36 and discreet Meyer wavelet analysis, in combination with existing permafrost depth  
 37 models, we propose that enhanced SRSF activity was primarily controlled by  
 38 permafrost degradation. Long-term relative change in permafrost depth provides a  
 39 compelling explanation for the high-magnitude departures from the SRSF background  
 40 rate and accounts for (1) the timing of peak SRSF frequency, (2) the significant lag  
 41 ( $\sim 2.2 \text{ ka}$ ) between the Holocene Thermal Maximum and the SRSF frequency peak  
 42 and (3) the marked decline in frequency in the late-Holocene. This interpretation is  
 43 supported by geomorphological evidence, as the spatial distribution of SRSFs is  
 44 strongly correlated with the aspect-dependent lower altitudinal limit of mountain  
 45 permafrost in cliff faces. Results are indicative of a causal relationship between  
 46 episodes of relatively warm climate, permafrost degradation and the transition to a  
 47 seasonal-freezing climatic regime. This study highlights the importance of permafrost  
 48 degradation as a conditioning factor for cliff collapse, and hence the importance of  
 49 paraperiglacial processes; a result with implications for slope instability in glacial and  
 50 periglacial environments under global warming scenarios.

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*Key words:* small rock-slope failures (SRSFs), Schmidt-hammer exposure-age dating (SHD), permafrost degradation, Holocene Thermal Maximum, climate-change impacts, paraperiglacial processes, southern Norway

Rock-slope failures (RSFs) are indicative of instability in the landscape and reflect several geophysical processes and potential trigger factors related to rock mechanics, geomorphology, hydrology and environmental change. Moreover, RSFs constitute significant natural hazards. As a result, understanding the environmental controls on RSF occurrence provides crucial information which can inform modelling of future RSF activity in response to climate forcing (Rapp 1960a, 1960b; Brunsden & Prior 1984; Evans et al. 2006; Clague & Stead 2012; Davies 2015).

Numerous RSFs have been investigated in regions of high relief and, in some cases, RSF deposits have been dated (e.g. Korup et al. 2007; Ballantyne et al. 2014a, 2014b). However, previous research has primarily focused on modern examples, spectacular cases or small numbers of massive rock-slope failures (MRSFs;  $\sim 10^8 \text{ m}^3$ ). Which, in combination with uncertainty associated with current geochronological approaches, limits our understanding of the fundamental geophysical processes and environmental controls that determine RSF occurrence. Particular studies of RSFs have used a variety of techniques and, on some occasions, a combination of geochronological methods (Lang et al. 1999; Hermanns et al. 2000; Crosta & Clague 2009; Deline & Kirkbride 2009; Prager et al. 2009; Pánek 2014; Böhme et al. 2015; Moreiras et al. 2015; Mercier et al. 2017), but the opportunities for accurate dating are relatively rare.

The primary method for numerical-age dating of RSF deposits is terrestrial cosmogenic nuclide dating (TCND;  $^{10}\text{Be}$ ,  $^{26}\text{Al}$ ,  $^{36}\text{Cl}$ ) as this technique permits direct sampling and age determination of the exposed rock surfaces associated with RSFs (Hermanns et al. 2001, 2004, 2017; Cossart et al. 2008; Dortch et al. 2009; Ivy-Ochs et al. 2009; Penna et al. 2011; Ballantyne & Stone 2013; Ballantyne et al. 2013, 2014a, 2014b; Böhme et al. 2015; Schleier et al. 2015, 2017). However, the high financial cost of this technique limits its routine application which, in turn, often prevents statistically robust identification and rejection of erroneous results (Tomkins et al. 2018b). Consequently, there are still few reliable chronologies of RSFs which limits our understanding of the environmental factors determining their spatial and temporal occurrence.

In this paper we develop a methodology for the investigation and dating of RSFs, with targeted study of 'small rock-slope failures' (SRSFs;  $< 10^3 \text{ m}^3$ ). This focus has the advantage over MRSFs of permitting the dating and study of a relatively large sample of simple, likely single-event RSFs within a specified region. The methodology has been developed in conjunction with the relatively new calibrated-age dating technique of Schmidt-hammer exposure-age dating (SHD) (Shakesby et al. 2006, 2011; Matthews & Owen 2011; Matthews et al. 2015; Matthews & Wilson 2015; Winkler et al. 2010, 2016; Wilson et al. 2017). SHD has the potential to estimate the numerical age of rock-surface exposure at low cost with comparable accuracy and precision, and greater representativeness, than TCND over the Late

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3 101 | Glacial and Holocene (cf. [Winkler 2009](#); [Winkler & Matthews 2010](#); Matthews &  
4 102 | Winkler 2011; Matthews et al. 2013; Wilson & Matthews 2016; [Tomkins et al. 2016](#),  
5 103 | [2018a](#), [2018b](#), [2018c](#)).

6 104  
7 105 | Specific objectives of this paper are three-fold:

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- To establish a Holocene chronology of SRSF events [in the alpine zone of Jotunheimen, southern Norway](#) and identify any [phases of](#) instability;
  - To explore relationships between the timing of Holocene SRSF events and regional environmental changes, including climatic changes; and
  - To develop [further](#) the potential of SHD as a calibrated-age dating technique in the context of RSFs.

### 115 Study area and environmental context

117 SRSFs were investigated in a broad area of northern Jotunheimen, the highest  
118 mountain massif in southern Norway, which culminates in Galdhøpiggen (2469 m  
119 above sea level; a.s.l.). The study area extends from Sognefjell in the west to  
120 Veodalen in the east (Fig. 1). Most SRSFs were found in Leirdalen, Bjørndalen (a  
121 western tributary valley to upper Leirdalen) and Gravidalen. The SRSFs occurred over  
122 an altitudinal range of 600 m (950-1550 m a.s.l.), mainly above the tree line, which  
123 lies at ~1000-1100 m a.s.l., in the alpine zone, and mainly in the low- and mid-alpine  
124 belts (Moen 1999). Examples of SRSFs from the study area are shown in Fig. 2.

126 Climatic data from the Sognefjell meteorological station (1413 m a.s.l.)  
127 indicate a mean annual air temperature of +3.1 °C (mean July temperature +13.4 °C;  
128 mean January temperature -10.7 °C), and a mean annual precipitation of 860 mm,  
129 much of which occurs as snow (climatic normals AD 1961-1990; Aune 1993; Førland  
130 1993). These data are consistent with a [lower](#) altitudinal limit of discontinuous  
131 permafrost at ~ 1450 m a.s.l. in the Galdhøpiggen massif (Ødgård et al. 1992; Isaksen  
132 et al. 2002; Farbrot et al. 2009; Lilleøren et al. 2012) with permafrost limits rising  
133 eastwards as continentality increases (Etzelmüller et al. 2003; Ginås et al. 2017).  
134 However, Hipp et al. (2014) have demonstrated a large difference of several hundred  
135 metres in the lower limits of permafrost between north- and south-facing rock walls.  
136 In the Galdhøpiggen massif, the lower altitudinal limit of rock-wall permafrost is  
137 located at 1500-1700 m a.s.l. in south-facing rock walls but only 1200-1300 m a.s.l. in  
138 shaded, north-facing rock walls (Hipp et al. 2014). Small valley glaciers, cirque  
139 glaciers and ice caps are common at and above these altitudes on the surrounding  
140 mountain peaks and plateaux (Andreassen & Winsvold 2012).

142 The metamorphic [geology](#) of the region consists primarily of pyroxene-  
143 granulite gneiss with peridotite intrusions [and quartzitic veins](#) (Battey & McRitchie  
144 1973, 1975; Lutro & Tveten 1996), [and gabbroic gneiss in the area investigated on](#)  
145 [Sognefjell \(Gibbs & Banham 1979\). Only boulders and bedrock of pyroxene-](#)  
146 [granulite gneiss and gabbroic gneiss were used in this study, as described below.](#)  
147 [Although these broad lithological categories include quite variable mineralogy, any](#)  
148 [differences in surface R-values due to lithology will likely be significantly smaller](#)  
149 [than the effect of variable exposure age given the relatively long Holocene](#)  
150 [timescales of exposure and limited climatic variability within the study region.](#)

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3 151 Topographically, most of the valley-side slopes have experienced a considerable  
4 152 degree of glacial erosion, although elements of ancient palaeic surfaces are  
5 153 preserved in the landscape (Ahlmann 1922; Gjessing 1967; Lidmar-Bergström et al.  
6 154 2000) due, at least in part, to non-erosive, cold-based conditions during glaciations.  
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9 156 Jotunheimen was located near the position of the main ice-divide and ice-  
10 157 accumulation area of the Scandinavian ice-sheet at the maximum of the Last  
11 158 (Weichselian) Glaciation. Deglaciation of the main valleys is likely to have occurred  
12 159 by ~9.7 ka, following the Erdalen Event, late in the Preboreal chronozone (Dahl et  
13 160 al. 2002; Matthews & Dresser 2008; Velle et al. 2010). Most glaciers appear to have  
14 161 melted away during the Holocene Thermal Maximum (Nesje 2009) when permafrost  
15 162 limits were also higher than today (Lilleøren et al. 2012), but regenerated during  
16 163 neoglaciation, certainly by 5.5 ka and possibly as early as 7.6 ka (Ødgård et al.  
17 164 2017). Both neoglaciation and lowering of permafrost limits occurred as a result of  
18 165 climatic deterioration (cooler and wetter) in the late Holocene, culminating in the  
19 166 Little Ice Age glacier maximum of the eighteenth century (Matthews 1991, 2005;  
20 167 Matthews & Dresser 2008). Future predicted mean annual warming of 0.3-0.4 °C  
21 168 per decade in Scandinavia (Benestad 2005) is likely to lead to unprecedented glacier  
22 169 retreat (Nesje et al. 2008) and a continuing rise in permafrost limits (Lilleøren et al.  
23 170 2012).  
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## 26 172

### 27 173 Methodology

#### 28 174

#### 29 175 *Definitions and criteria for recognition of SRSFs*

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31 177 | The term ‘rock-slope failure’ (RSF) refers to both (1) a mass-movement process  
32 178 involving the deformation and loss of integrity of a volume of intact bedrock followed  
33 179 by its *en masse* collapse and downslope movement under gravity and (2) the resulting  
34 180 landform. This definition is used here to distinguish RSF from ‘rockfall’ – the  
35 181 smaller-scale process involving the piecemeal detachment and free fall of individual  
36 182 rock particles – even though the term rockfall is commonly used at all scales,  
37 183 including the largest landslides and rock avalanches (MRSFs), which are often  
38 184 complex and multiphase (cf. Bates & Jackson 1987; Cruden & Varnes 1996; Braathen  
39 185 et al. 2004; Luckman 2004; Evans et al. 2006; [Hermanns et al. 2006](#); Jarman 2006;  
40 186 Frattini et al. 2012; Hermanns & Longva 2012; Shakesby 2014; Brideau & Roberts  
41 187 2015).  
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44 189 | Fundamental to this study was the selection of SRSF landforms that  
45 190 represented, as far as it was possible to ascertain, the product of single events. Criteria  
46 191 for recognition of such SRSFs were as follows:  
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- 49 193 • a compact and coherent depositional fan of predominantly angular boulders  
50 194 located close to a bedrock cliff.
- 51 195 • a simple erosional scar in the cliff, immediately upslope of the fan, which is  
52 196 comparable in scale to the fan and therefore represents the likely source of the  
53 197 failed rock material;
- 54 198 • an absence of alternative sources of boulders up-slope of the scar.  
55 199

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3 200 Although no upper limit was placed on the size of the SRSFs recognized in this study,  
4 201 these criteria become less easily satisfied as RSFs increase in size. The lower size  
5 202 limit was the practical one of sufficient boulders for reliable Schmidt hammer  
6 203 measurement. Thus, the size range included in the study was determined by the RSFs  
7 204 in the region. Furthermore, the 92 investigated cases represent the whole population  
8 205 of SRSFs that satisfied the above criteria [in the study area](#).

#### 9 206 10 207 *Measurement of SRSF characteristics*

11 208  
12 209 Estimates were made in the field of the length and average width of the depositional  
13 210 fan of each SRSF. Aspect and the altitude of the fan apex were estimated from  
14 211 topographic maps at a scale of 1:50,000 with a contour interval of 20 m,  
15 212 supplemented by altimeter and GPS measurements in the field. Fan volume was  
16 213 calculated from the length and average width measurements, assuming an average fan  
17 214 thickness of 1 m and a voids fraction (volume of voids/total fan volume) of 40%.  
18 215 Although some of the largest fans are thicker than 1 m in places, all are thinly spread  
19 216 across and down slope and rarely involve piles of debris. Lower voids fractions have  
20 217 generally been used for MRSFs, rock avalanches, talus and other mass movement  
21 218 types involving mixed particle sizes, fine matrix and/or compacted material (Owen et  
22 219 al. 2010; Sass & Wollny 2001; Hungr & Evans 2004; Wilson 2009; Stock &  
23 220 Uhrhammer 2010; Sandøy et al. 2017). The value of 40% is justified given the  
24 221 absence of fine matrix (Fig. 2) and lack of compaction, and its compatibility with  
25 222 similar values for clean, open-graded, angular aggregate material used as backfill in  
26 223 foundation engineering (StormTech 2012; cf. Dann et al. 2009).

#### 27 224 28 225 *Measurement of Schmidt-hammer R-values*

29 226  
30 227 N-type mechanical Schmidt hammers (Proceq 2004; Winkler & Matthews 2014) were  
31 228 used to measure rebound (R-) values from 100 boulders in each depositional fan. R-  
32 229 values reflect lithologically-determined rock hardness and the compressive strength of  
33 230 the rock surface: hence, R-values decline following exposure of a rock surface to  
34 231 subaerial weathering. For boulder surfaces of the same lithology but differing age, R-  
35 232 values therefore reflect the exposure age (time elapsed since exposure) of the rock  
36 233 surface. Use of one impact per boulder from a large sample of boulders ensures that  
37 234 the R-value frequency distribution can be used to approximate the boulder-age  
38 235 distribution (Matthews et al. 2014, 2015).

39 236  
40 237 Precautions taken to eliminate or reduce possible sources of uncertainties and  
41 238 errors in Schmidt-hammer measurement included avoiding unstable or small boulders,  
42 239 boulder or bedrock edges, joints or cracks, unusual lithologies and lichen-covered or  
43 240 wet surfaces (cf. Shakesby et al. 2006; Matthews & Owen 2010; Viles et al. 2011).  
44 241 Rock surfaces were not cleaned or artificially abraded [prior to impact with the](#)  
45 242 [Schmidt hammer](#) (cf. the carborundum treatment of Viles et al. 2011) because such  
46 243 treatment would [likely](#) remove age-related weathering effects. [However, there is](#)  
47 244 [continued debate as to whether rock surfaces should be abraded prior to testing](#)  
48 245 [\(Moses et al. 2014\) although a consistent sampling approach may enable age-related](#)  
49 246 [information to be retained \(c.f. Tomkins et al. 2018b\)](#). Where possible, horizontal  
50 247 boulder surfaces were impacted but only vertical rock faces were available on cliffs.  
51 248 The two hammers used had been recently re-calibrated at a recognised service centre  
52 249 and were tested frequently on the manufacturer's test anvil throughout the study to

250 ensure there had been no deterioration in instrument performance following large  
 251 numbers of impacts (cf. McCarroll 1987, 1994; Winkler & Matthews 2016).  
 252 Measurements at 84 sites were restricted to rock surfaces of pyroxene-granulite  
 253 gneiss. At [the](#) 8 sites on Sognefjell, gneissic rocks with gabbroic textures were used,  
 254 which necessitated a separate calibration equation (see below).

#### 255 *Testing the validity of the approach*

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 258 In order to test the validity of our approach and especially whether the boulders  
 259 comprising the depositional fans actually represent single rock-failure events and  
 260 whether the local source of the boulders had been correctly identified, R-value  
 261 distributions associated with six fans and their corresponding scars were investigated.  
 262 Two separate tests of validity were conducted.

263  
 264 First, in the *fan-scar comparison test*, a comparable sample of R-values ( $n =$   
 265 100) from the surface of the corresponding scar was compared with the R-value  
 266 distribution of the fan to identify whether or not the scar was the likely source of the  
 267 boulders in the fan. If the scar was indeed the source of the boulders, the expectation  
 268 would be no significant difference in the R-values derived from the scar and its  
 269 corresponding fan because both would have experienced exposure over the same  
 270 period of time.

271  
 272 Second, the *unfailed-cliff test* required a comparable sample of R-values ( $n =$   
 273 100) from the adjacent intact (unfailed) bedrock cliff and also aimed to establish that  
 274 the cliff was the bedrock source for the fan boulders. If this was the case, it would be  
 275 expected that R-values from the unfailed cliff would be similar to or lower than the R-  
 276 values of both the scar and the fan. Any departure from these expectations would  
 277 indicate possible flaws in our approach.

278  
 279 The principles behind the fan-scar comparison test and the unfailed-cliff test  
 280 are illustrated in Fig. 3, which also shows the expected relationships between R-  
 281 values from the fans and R-values from the rock surfaces used as control points in the  
 282 calibration equations.

#### 283 *Calibrated-age dating using SHD*

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 286 Although there was earlier use of the Schmidt hammer for dating purposes ([e.g.](#)  
 287 [Matthews & Shakesby 1984; Nesje et al. 1994; Aa & Sjøstad 2000; Aa et al. 2007](#)),  
 288 SHD [has been](#) developed [more recently](#) as a calibrated-age dating technique (Colman  
 289 et al., 1987) [incorporating measures of uncertainty based on statistical confidence](#)  
 290 [intervals](#) (cf. Shakesby et al. 2006; Matthews & Owen 2011; Matthews & Winkler  
 291 2011; Matthews & McEwen 2013). Critically, this involves the derivation of a  
 292 calibration equation and confidence limits for age.

293  
 294 The calibration equation is based on linear regression of surface age ( $Y$ ) on  
 295 mean R-value ( $X$ ):

$$296 \quad Y = a + bX \quad (1)$$

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3 299 | A linear relationship can be justified on both theoretical and empirical grounds.  
4 300 | Although chemical weathering rates are likely to decline over longer timescales  
5 301 | (Colman 1981; Colman & Dethier 1986; Stahl et al. 2013; Tomkins et al. 2018a,  
6 302 | 2018b), near-linear rates can be expected over the Holocene timescale, especially  
7 303 | where relatively resistant lithologies are subject to relatively slow rates of chemical  
8 304 | weathering in a periglacial environment (André 1996, 2002; Nicholson 2008, 2009;  
9 305 | Matthews & Owen 2011; Matthews et al. 2016). Although physical (freeze-thaw)  
10 306 | weathering is well known in periglacial environments, it is highly dependent on  
11 307 | moisture availability for ice-lens growth (Hallet et al. 1991; Hall et al. 2002; Murton  
12 308 | et al. 2006; Matsuoka & Murton 2008) and there is no evidence that it has affected the  
13 309 | well-drained surfaces used in this study (neither boulders in the dated depositional  
14 310 | fans nor bedrock control surfaces).

15 311 |  
16 312 |       Furthermore, Shakesby et al. (2011) specifically tested the linearity  
17 313 | assumption in relation to granite boulders on independently-dated staircases of raised  
18 314 | beaches deposited since 10.4 ka in northern Sweden, with the conclusion that the  
19 315 | relationship between mean R-value and age was best described by a linear function.  
20 316 | The same conclusion can be reached from age-calibration curves in the British Isles  
21 317 | (Tomkins et al. 2018a) and the Pyrenees (Tomkins et al. 2018b), which are based on  
22 318 | 54 and 52 <sup>10</sup>Be TCND-dated granitic surfaces respectively, all associated with glacial  
23 319 | depositional or erosional landforms (moraine boulders or ice-sculpted bedrock).  
24 320 | While the Pyrenean age-calibration curve is clearly non-linear over the full age range  
25 321 | of ~50 ka, both age-calibration curves evidence linearity over the last ~20 ka. Other  
26 322 | studies that have suggested non-linear relationships have involved long timescales  
27 323 | and/or have had insufficient control points to test the linearity assumption rigorously  
28 324 | over the Holocene timescale (e.g. Betts & Latta 2000; Sánchez et al. 2009; Černá &  
29 325 | Engel 2011; Stahl et al. 2013).

30 326 |  
31 327 |       Based on two control points, the *b* coefficient can be defined as:

$$32 328 |$$

$$33 329 | b = (y_1 - y_2) / (x_1 - x_2) \quad (2)$$

$$34 330 |$$

35 331 | where  $x_1$  and  $x_2$  are the mean R-values of the older and younger control points,  
36 332 | respectively, and  $y_1$  and  $y_2$  are their respective ages. Once the *b* coefficient is known,  
37 333 | the *a* coefficient is found by substitution in equation (1). Only two control points of  
38 334 | widely differing age are available from Jotunheimen (see below). Provided they are of  
39 335 | good quality, however, two control points are sufficient for accurate R-value  
40 336 | calibration provided the underlying relationship between R-value and age is  
41 337 | approximately linear.

42 338 |  
43 339 |       For a landform produced by a single event, the SHD age resulting from this  
44 340 | calibration is the average age of the surface boulders and hence the landform age  
45 341 | (Matthews et al. 2015). Confidence intervals for the SHD age (95%) are calculated as  
46 342 | the total error ( $C_t$ ) by combining the error associated with the calibration equation ( $C_c$ )  
47 343 | with the sampling error associated with the surface to be dated ( $C_s$ ):

$$48 344 |$$

$$49 345 | C_t = \sqrt{C_c^2 + C_s^2} \quad (3)$$

$$50 346 |$$

$$51 347 | C_c = C_o - [(C_o - C_y) (R_s - R_o) / (R_y - R_o)] \quad (4)$$

$$52 348 |$$



$$C_s = b[ts / \sqrt{(n-1)}] \quad (5)$$

where  $C_o$  and  $C_y$  are the 95% confidence intervals of the older and younger control points (in years); and  $R_o$ ,  $R_y$  and  $R_s$  are the mean R-values of the older control point, the younger control point and the surface to be dated, respectively.  $C_s$  depends on the number of R-value impacts on the surface to be dated (sample size,  $n$ ), the standard deviation of those impacts ( $s$ ), and Student's  $t$  statistic. Thus, the confidence interval ( $C_t$ ) associated with any SHD age depends not only on the sample sizes used to establish the calibration equation and characterize the surface to be dated but also the natural variability exhibited by all the rock surfaces involved.

#### *Control points for calibration equations*

For this study, we constructed separate calibration equations for rock surfaces composed of pyroxene-granulite gneiss and gabbroic gneiss (each equation based on two control points). Data for the older control points, which relate to glacially-scoured bedrock surfaces, were taken from Matthews & Owen (2010). Their data from four sites in Leirdalen and Gravdalen (S and E Smørstabbtindan) were used for the pyroxene-granulite gneiss calibration equation; four sites near Leirbreen and Bøverbreen, close to Sognefjell (W Smørstabbtindan) supplied the data for the gabbroic gneiss calibration equation (Fig. 1).

Evidence for deglaciation of these sites is provided by basal  $^{14}\text{C}$  dates from peat bogs and lakes in Leirdalen, Bjørndalen, and on Sognefjell (Table 2). These  $^{14}\text{C}$  dates were recalibrated to calendar age ranges with the OxCal online program (v.4.3) using the IntCal13 calibration dataset (Reimer et al., 2013). Although one of the calibrated age ranges is significantly older, 9.7 ka is the only date for deglaciation that is compatible with the other four  $^{14}\text{C}$  dates. Use of 9.7 ka as the age of the old control points for SHD calibration can be justified on the further grounds that it is the expected date for termination of the Erdalen Event in neighbouring regions (Dahl et al. 2002) and is consistent with empirical evidence for and large-scale modeling of deglaciation in southern Norway (Dahl et al. 2002; Goehring et al. 2008; Nesje 2009; Mangerud et al. 2011; Hughes et al. 2016; Stroeven et al. 2016). Thus, the potential errors in the old control points appear to be small in relation to the calibration errors ( $C_c$  and  $C_s$ ) that are taken fully into account in this study.

Calibration equations given in Matthews & Owen (2010) for these rock types could not be used because their younger control points were derived from glacially-abraded surfaces from glacier forelands. Such smooth surfaces are not appropriate as a source of young control points for dating the exposure-age of boulders originating from SRSFs, which are rougher in texture yielding lower R-values than abraded surfaces of the same age (Shakesby et al. 2006; Matthews & McEwen 2013; Matthews et al. 2015). In contrast, after prolonged weathering, originally smooth surfaces are expected to yield similar R-values, and hence SHD ages, to initially rough surfaces.

Young control points with similar roughness properties to fresh boulder surfaces derived from SRSFs were therefore sought. These included: (1) boulders and bedrock surfaces produced by a recent rock-slope failure in Gravdalen and (2) bedrock exposed recently in road cuts in Gravdalen and on Sognefjell (Fig. 1). Both

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3 399 types of surfaces have been shown in previous studies to yield R-values that are  
4 400 statistically indistinguishable from each other provided sufficient care is taken to  
5 401 impact only truly fresh rock surfaces (Matthews & Wilson 2015; Matthews et al.  
6 402 2016). Furthermore, both types of recent rock surfaces used as young control points in  
7 403 this study were lichen-free and hence were assigned a maximum exposure age of 25  
8 404 years based on various estimates of the time required for the establishment (ecesis) of  
9 405 crustose lichens on bedrock surfaces in this environment (Matthews 2005; Matthews  
10 406 & Owen 2008; Matthews & Vater 2015). Errors in the age of the young control point  
11 407 are therefore considered to be negligible in the context of this study.  
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#### 409 *Chronology construction and analysis*

14 410  
15 411 Holocene chronologies of SRSF events were constructed from the SHD ages of the 92  
16 412 SRSF fans using a number of statistical approaches. First, graphical analysis of age-  
17 413 frequency distributions used 2000-yr, 1000-yr, 500-yr and 200-yr time intervals to  
18 414 define major clusters of SHD ages and hence possible multi-centennial to millennial  
19 415 phases of enhanced SRSF frequency (Matthews et al. 2009; Matthews & Seppälä  
20 416 2015). Based on the same events weighted according to their rock volume, a second  
21 417 chronology was constructed showing the changing magnitude of SRSF events through  
22 418 the Holocene.  
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25 419  
26 420 In order to take account of dating uncertainty, a weighted age-frequency  
27 421 distribution was constructed in which each SHD age was plotted over five 200-yr age  
28 422 classes: a weight of 4 was used for the central class; the second and fourth classes  
29 423 were weighted 2. Thus, the SHD age was plotted over a range of 1000 yr, consistent  
30 424 with the average 95% confidence interval of  $\pm 991$  yr calculated for the 92 SRSF fans  
31 425 (see below). One-sample  $\chi^2$  tests were used to test the hypothesis that the dated events  
32 426 were sampled from an underlying population of events with an even distribution  
33 427 through time.  
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36 429 To support weighted age-frequency analysis, the distribution of calculated  
37 430 SRSF ages was analysed using probability density function analysis. Probability  
38 431 density estimates (PDEs) were produced and modelled to separate out individual  
39 432 Gaussian distributions using the KS density kernel in MATLAB (2015) and a  
40 433 dynamic smoothing window based on age uncertainty (c.f. Dortch et al. 2013). The  
41 434 sum of individual Gaussian distributions integrates to the cumulative PDE at 1000  
42 435 iterations to obtain a good model fit. The goodness of fit between the re-integrated  
43 436 PDE, which is derived from individual Gaussian distributions, and the cumulative  
44 437 PDE, which is derived from the full age dataset, is indicated graphically. PDE  
45 438 analysis was repeated using a number of individual Gaussian distributions ( $n = 1-10$ ).  
46 439 To avoid over-interpretation of SRSF modes, the PDE model with the minimum  
47 440 number of individual Gaussian distributions, which also achieved a good model fit,  
48 441 was selected. This analytical method has primarily been employed in studies using  
49 442  $^{10}\text{Be}$  (cf. Dortch et al. 2013; Murari et al. 2014) or SHD (Barr et al. 2017; Tomkins et  
50 443 al. 2018a; 2018b; 2018c) to account for negative or positive skew of moraine boulder  
51 444 datasets and to identify and reject ages that are compromised by moraine degradation  
52 445 (Briner et al. 2005; Heyman et al. 2011) or nuclide inheritance (Hallet & Putknonen  
53 446 1996). In these applications, PDE analysis and interpretation of individual Gaussian  
54 447 distributions (cf. Fig. 3 in Dortch et al. 2013) is based on the assumption that analysed  
55 448 ages relate to a single event e.g. moraine deposition. This assumption is clearly not  
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3 449 applicable to the analysis of SRSF ages, as each numerical age relates to a distinct  
4 450 event and an individual landform. As a result, individual Gaussian distributions are  
5 451 interpreted as reflecting the temporal clustering of events. The characteristics of  
6 452 individual Gaussian distributions, i.e. the peak probability density, width of PDE tails,  
7 453 1 $\sigma$  uncertainties and the number of contributing ages (Fig. 7), were used to assess the  
8 454 significance and temporal clustering of SRSF events in Jotunheimen over the last ~10  
9 455 ka.

10 456  
11 457 The individual distributions resulting from the PDE analysis indicated that  
12 458 further analysis was necessary. Thus, a change detection analysis approach was  
13 459 undertaken in MATLAB (2015) to identify statistically unique events. Change  
14 460 detection analysis utilizes the cumulative sum algorithm (cusum), which is commonly  
15 461 used to detect abrupt change in time series data in fields ranging from seismology  
16 462 (Dera & Shumwayb 1999), remote sensed imagery (Lu et al. 2016), and GPS  
17 463 monitoring (Goudarzi et al. 2013). Parameters were set by using the average  
18 464 frequency and occurrence (~1 occurrence per 100 years) of SRSFs throughout the  
19 465 Holocene to filter out 'background' SRSF occurrence. The alarm limit was set at  $\geq 2$   
20 466 standard errors above background. To further explore the temporal pattern of SRSFs,  
21 467 discreet Meyer wavelet analysis was undertaken in MATLAB (2015) to decompose  
22 468 SRSF occurrence through time. Wavelets are discreet oscillations in both time and  
23 469 amplitude and, as such, are useful for identifying discreet events. Wavelet analysis  
24 470 has been used to identify climate signals from various records including  $\delta O^{18}$  (Lau &  
25 471 Weng 1995), and sea surface temperature (Torrence & Compo 1998). The 100 yr  
26 472 binned SRSF age data was passed through the discreet Meyer wavelet with six levels  
27 473 of deconvolution.

28 474  
29 475 Major and minor changes in SRSF activity were then compared with changes  
30 476 in regional Holocene climatic and other geo-environmental indicators to infer possible  
31 477 causes. Specific analyses were performed to investigate relationships between the  
32 478 occurrence of SRSF events and the lower altitudinal limits of discontinuous  
33 479 permafrost using aspect-dependent limits determined for rock walls in the  
34 480 Galdhøpiggen massif by Hipp et al. (2014). The current (AD 2010-2013) lower limits  
35 481 that were used for rock walls facing north, east, south and west were 1250 m, 1450 m,  
36 482 1600 m and 1450 m, respectively.

## 483 484 485 Results

### 486 487 *Data on the SRSFs*

488 489 Data on the size and environmental characteristics of the SRSFs are summarized in  
490 490 Table 1 and Fig. 4. The volume of the fans (Fig. 4A) ranges from 12 to 2520 m<sup>3</sup>, with  
491 491 90% <1000 m<sup>3</sup>, 40% <100 m<sup>3</sup> and a median size of only 180 m<sup>3</sup>. The altitudinal range  
492 492 is 960 to 1550 m a.s.l. (Fig. 4B), with a mean altitude of 1340 m a.s.l. There is a  
493 493 preferred aspect with 43% facing east, 34% facing south and 17% facing west, but  
494 494 only 5% facing north (Fig. 4C).

495 496 Schmidt-hammer R-values vary widely between SRSFs (Table 1) and the  
497 497 frequency distribution of mean R-values reveals several important features (Fig. 4D).  
498 498 Mean R-values exhibit a very wide range of >20 units from 37.0 to 57.5. The overall

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3 499 mean R-value across the 92 SRSFs is 48.2 but those R-values associated with  
4 500 gabbroic gneiss (overall mean R-value 39.4, n = 8) are appreciably lower than the  
5 501 remainder involving pyroxene-granulite gneiss (overall mean R-value 49.1, n = 84).  
6 502 The latter value corresponds closely with the 49-50 modal class for the distribution.  
7 503

#### 8 504 *Control-point data and calibration equations*

9 505  
10 506 | Data from the control points (Table 3) indicate widely different mean R-values  
11 507 (differing by at least 20 units) for surfaces that differ in age by ~9700 years. It should  
12 508 also be noted that the overlapping 95% confidence intervals associated with each pair  
13 509 of replicates for particular control points indicate that their mean R-values do not  
14 510 differ significantly from each other. Control surfaces of the same age on different  
15 511 lithologies are, however, characterized by non-overlapping confidence intervals, and  
16 512 | thus show significantly different mean R-values and justify the use of separate  
17 513 calibration equations for SRSFs developed in pyroxene-granulite gneiss and gabbroic  
18 514 gneiss. The calibration equations derived from these data for the two lithologies are  
19 515 shown in Figure 5 alongside the linear relationships they represent.  
20 516

#### 21 517 *Fan-scar-cliff comparison tests*

22 518  
23 519 Mean R-values for three of the six fans tested did not differ significantly from the  
24 520 mean R-values of the corresponding scars, in accordance with expectation (Fig. 3 and  
25 521 | Table 4). However, three fans (Nos 51, 58 and 81) are characterized by mean R-  
26 522 values that are significantly lower than the mean R-values from their scars. This  
27 523 | suggests one or more of four possible explanations: (1) rock surfaces of some  
28 524 boulders in these fans are more weathered because they include the products of older  
29 525 rock failures than those that produced the measured bedrock faces of the scars; (2)  
30 526 some of the measured R-values from boulders in the fans reflect the incorporation of  
31 527 | bedrock surfaces that were pre-weathered on the cliff face before the failures  
32 528 occurred; (3) some of the R-values from boulders in the fans reflect the incorporation  
33 529 of inherited structures (e.g. joint planes) that were pre-weathered at depth before the  
34 530 failures occurred; and (4) at least part of the cliff bedrock is more resistant to  
35 531 weathering than the boulder surfaces measured in the fans. Interestingly, no fan  
36 532 exhibits a mean R-value that is significantly greater than that of its corresponding  
37 533 scar. This shows that even where more than one phase of activity seems possible, any  
38 534 blocks that were later removed from the scars were insufficient in number to affect  
39 535 appreciably the mean R-values of the fans.  
40 536

41 537 Comparisons between scars and unfailed cliffs or between fans and unfailed  
42 538 cliffs are entirely in agreement with expectation. In three cases (fan Nos 5, 51 and 58)  
43 539 neither the mean R-values for scars and unfailed cliffs nor the mean R-values for fans  
44 540 | and unfailed cliffs differ significantly, suggesting that all the exposed surfaces are of  
45 541 the same age (and relatively old). In the other three cases (fan Nos 46, 47 and 81) the  
46 542 mean R-values of the scars and the fans are both significantly higher than the mean R-  
47 543 values of the unfailed cliffs, confirming the SRSFs are younger than the exposure age  
48 544 of the unfailed cliffs.  
49 545

50 546 Comparison of the mean R-values from unfailed cliffs with the values from  
51 547 | the older control points given in Table 3 indicates that unfailed cliff surfaces were  
52 548 exposed during or immediately after deglaciation at ~9700 cal. BP. As all surfaces  
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yielded mean R-values lower than those characteristic of the younger control points (Table 4), it appears that fan deposition and scar exposure occurred throughout the Holocene and, in some cases, thousands of years after regional deglaciation. As a result, the temporal distribution of fan mean-R-values likely reflects the timing of single-event SRSF activity.

#### *Temporal variations in SRSF activity*

The age of each SRSF event, including its 95% confidence interval, is summarized graphically in Fig. 6A. Although there is some evidence of differences in the age distributions between the different valleys, there is no statistically significant correlation between SRSF age and altitude and no significant difference in age between aspects. The overall mean age of all 92 SRSF events is 5124 years, which equates with an average regional frequency of 1 in 105 years.

Simple age-frequency distributions of the SRSF events within the region as a whole are shown in Fig. 6B. Although these events occurred without any prolonged break in activity, their frequency varied considerably over the last ~10,000 years. The distribution based on 2000-yr time intervals has a single mode indicating an increase in the frequency of events through the early Holocene, a distinct peak in activity in the 6.0-4.0 ka time interval, and a consistent decline in activity thereafter. The use of 1000-year time intervals reveals two modes – at 8.0-7.0 and 5.0-4.0 ka, respectively. At least three modes can be recognized when 500-yr time intervals are used (at 9.0-8.5, 7.5-7.0 and 4.5-4.0 ka) and many more can possibly be discerned in the distribution based on 200-year time intervals. However, analysis of SRSF modes based on 200-year time intervals is not advisable, as this time interval (0.2 ka) is significantly smaller than the typical uncertainty of SRSF ages (~1 ka). Despite this, the hypothesis of an even distribution of SRSF events through time can be rejected at  $p < 0.01$  irrespective of the age classes used.

The weighted age-frequency distribution (Fig. 6C) has four modes (at ~ 8.9, 7.3, 5.9 and 4.5 ka), which suggests that only four minor phases of enhanced SRSF frequency are meaningful. Furthermore, according to the weighted distribution, the frequency of events declines steadily after ~4.5 ka with no marked fluctuations.

The temporal pattern in the magnitude of the SRSFs (rock volume), as shown in in Fig. 6D, is substantially the same as the frequency distribution (compare with use of a 200-yr interval in Fig. 6B). In particular, the age-volume distribution has a similar major peak between 4.8 and 4.2 ka, and relatively little activity before 9.0 ka or after 1.0 ka.

Probability density function analysis indicates that the spread of SRSF ages does not conform to a normal distribution (Fig. 7A) and instead, is best explained by 5 individual Gaussian age distributions (Fig. 7B). The sum of individual Gaussian distributions produces a re-integrated PDE which achieves a good model fit with the cumulative PDE. PDE analysis using < 5 individual Gaussian age distributions returns a poor ( $n \leq 3$ ) or sub-optimal ( $n = 4$ ) model fit. PDE analysis using > 5 individual Gaussian age distributions does not therefore significantly improve the model fit and instead risks over-interpretation of the number of SRSF modes. PDE analysis returns peak Gaussian ages (Fig. 7C) of  $9.00 \pm 1.13$  ka ( $n = 14$ ),  $7.38 \pm 0.99$  ka ( $n = 17$ ),  $6.40$

± 0.77 ka (n = 14), 4.50 ± 1.42 ka (n = 42) and 1.90 ± 1.42 ka (n = 18). Although these modes overlap with adjacent modes within 1 $\sigma$ , statistically significant differences between sequential Gaussian age distributions are revealed by two-sample Students t-tests ( $p < 0.01$ ).

These Gaussian age distributions closely match the four modes identified in weighted age-frequency analysis, with a dominant mode at ~4.5 ka (Fig. 7B). This mode is the highest probability Gaussian distribution, comprises a significant number of SRSF events (n = 42; Fig. 7D) and accounts for a large proportion of total SRSF volume over the last ~10 ka (18,744 m<sup>3</sup>). In contrast to weighted age-frequency analysis, PDE analysis returns an additional Gaussian age distribution during the late Holocene at ~1.9 ka. However, this is unlikely to reflect a period of enhanced SRSF activity as there is no clear clustering of SRSF ages (Fig. 7A), as evidenced by weighted age-frequency analysis. Instead, late Holocene ages likely reflect declining SRSF activity after the mid-Holocene peak.

The combined results of the age-frequency analyses and the Gaussian separation achieved for PDEs demonstrate that SRSF occurrence through time is non-uniform and multi-modal. Most notable is the high level of occurrence during the mid-Holocene, the clear statistical significance of which is confirmed by the results of change detection analysis. The cumulative sum change detection graph (Fig. 8A) shows a clear peak in the rate of SRSF intensity between 4.8 and 2.6 ka, significantly exceeding the 2 $\sigma$  threshold, with the largest departure from background occurring at 4.3 ka. Conversely, SRSF intensity is significantly reduced beyond the negative 2 $\sigma$  threshold during the late Holocene at 0.6–0.1 ka. These peaks are a significant departure from the normal rate of occurrence during the Holocene. The three other modes identified above as statistically significant must be regarded as relatively small departures from background SRSF periodicity.

Meyer wavelet analysis was used to explore the two statistically significant departures (> 2 $\sigma$ ) from the background SRSF rate, as identified by change detection analysis. The lowest frequency decomposed signal ( $d_6$ ) is shown in Fig. 8C. The full analysis record is provided in Supplementary Fig. 1.

## Discussion

### *Previous models of the timing of RSFs*

Widely different conceptual models can be proposed to describe and explain the temporal distribution of Late Pleistocene and Holocene RSFs. A schematic representation of several models, each of which links a distinctive pattern of change in the frequency and/or magnitude of RSFs to one or more specific causes or triggers, is shown in Fig. 9. Although they have been based mainly on MRSFs, these models are introduced here as a basis for discussion of our Holocene SRSFs. It should be emphasised, moreover, that RSFs may be multicausal and that most if not all of the models have yet to be rigorously tested against data sets with a large number of consistently dated RSFs.

*Model 1.* – The ‘continuity-of-activity model’, proposes that there are no

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3 649 significant temporal variations in the frequency and/or magnitude of RSFs throughout  
4 650 the Holocene. Despite the small number of dated RSFs available in most studies, few  
5 651 authors have advocated this model. However, the model does appear to be consistent  
6 652 with the temporal distribution of about 60 RSFs located in an extensive area of the  
7 653 Alps centred on the Austrian Tyrol (Prager et al. 2008), which exhibits only limited  
8 654 evidence of temporal clustering at ~10.5-9.4 ka and 4.2-3.0 ka. Prager et al. (2008)  
9 655 attributed the continuity of activity to complex interactions between the processes  
10 656 characterizing models 2-5 together with rock-strength degrading processes such as  
11 657 time-dependent progressive fracture propagation that can both prepare and trigger  
12 658 slope instabilities.  
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15 |       *Model 2.* – The ‘*intermittent-earthquakes model*’, which is applicable to  
16 661 | tectonically active regions [and](#) assumes that RSFs are triggered directly by large-  
17 662 | magnitude earthquakes generated by tectonically-driven uplift or other crustal  
18 663 | stresses. [Such](#) earthquakes are essentially randomly [distributed](#) in time and therefore  
19 664 | bear little or no relationship to deglaciation, climate or any of the other potential  
20 665 | causative factors in models 3-5 that are effective in tectonically stable regions (see,  
21 666 | for example, [Fjeldskaar et al. 2000](#); [Hermanns et al. 2001](#); [Keefer 2002, 2015](#); [Hewitt](#)  
22 667 | [et al. 2008](#); [Antinao & Gosse 2009](#); [Stock & Uhrhammer 2010](#); [Penna et al. 2011](#);  
23 668 | [McPhillips et al. 2014](#); [Marc et al. 2015](#); [Murphy 2015](#)).  
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26 |       *Model 3.* – The ‘*deglaciation-close-tracking model*’, is characterised by a  
27 671 | dominant peak in RSF activity immediately (i.e. within the first millennium)  
28 672 | following regional deglaciation, with subsequent asymptotic decline in activity. The  
29 673 | temporal pattern of activity is therefore a typical paraglacial response (cf. [Ballantyne](#)  
30 674 | [2002](#)). Causal factors that may account for such a pattern include glacial unloading,  
31 675 | glacial debuttressing, stress-release fracturing, enhanced groundwater pressure in rock  
32 676 | joints and permafrost degradation, all closely associated in time with deglaciation  
33 677 | ([Fischer et al. 2006](#); [Cossart et al. 2008](#); [McColl 2012](#); [McColl & Davies 2012](#);  
34 678 | [Ballantyne et al. 2014a, 2014b](#); [Böhme et al. 2015](#); [Deline et al. 2015](#); [Mercier et al.](#)  
35 679 | [2017](#)). [Hermanns et al. \(2017\)](#) found nearly half of 22 dated rock avalanches in  
36 680 | southwest Norway occurred within the first millennium following local deglaciation.  
37 681 | Although the majority of RSF events occur shortly after deglaciation, some occur  
38 682 | much later, due to time-dependent fracture propagation and progressive failure (e.g.  
39 683 | [Eberhardt et al. 2004](#); [Krautblatter et al. 2013](#); [Phillips et al. 2017](#)). The occurrence of  
40 684 | recent RSFs on glacier forelands following the retreat of mountain glaciers from their  
41 685 | Little Ice Age maximum limits provides some support for this model ([Evans &](#)  
42 686 | [Clague 1994](#); [Holm et al. 2004](#); [Matthews & Shakesby 2004](#); [Arsenault & Meigs](#)  
43 687 | [2005](#); [Allen et al. 2010](#); [Stoffel & Huggel 2012](#)).  
44 688

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46 |       *Model 4.* – The ‘*deglaciation-lagging model*’, features a significantly delayed  
47 690 | response to deglaciation. Peak RSF activity typically occurs within a few millennia of  
48 691 | deglaciation and corresponds with maximum glacio-isostatic rebound ([Hicks et al.](#)  
49 692 | [2000](#); [Ballantyne & Stone 2013](#); [Ballantyne et al. 2013, 2014a, 2014b](#); [Cossart et al.](#)  
50 693 | [2014](#); [Decaulne et al. 2016](#)). The cause of RSF events is seen as fault reactivation and  
51 694 | fracture propagation triggered by earthquakes, the frequency of earthquakes and RSFs  
52 695 | generally diminishing through the Holocene as the rate of glacio-isostatic uplift  
53 696 | declines.  
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56 |       *Model 5.* – The ‘*cool/wet-climate-response model*’, which applies particularly  
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699 | to the Holocene, reflecting several possible effects of climatic variations on RSF  
 700 | activity. Field monitoring, historical documentation and palaeo-studies indicate that  
 701 | precipitation variations can be a dominant trigger factor in the timing of RSFs but  
 702 | both cooler conditions and indirect effects such as variations in cleft water pressure,  
 703 | frost shattering and permafrost degradation have also been implicated in rock-slope  
 704 | instability (Eisbacher & Clague 1984; Matthews et al. 1997; [Trauth et al. 2000, 2003](#);  
 705 | Dapples et al. 2003; Soldati et al. 2004; Prager et al. 2008; Crozier 2010; Borgatti &  
 706 | Soldati 2010; Blikra & Christiansen 2014; Zerathe et al. 2014; Johnson et al. 2017).  
 707 | Furthermore, Evans & Clague (1994), Huggel et al. (2010, 2012) and Stoffel &  
 708 | Huggel (2012) highlighted the possible effects of recent climate warming on RSFs,  
 709 | and direct solar heating of rock faces has also been examined as a possible trigger (cf.  
 710 | Allen & Huggel 2013; Collins & Stock 2016). In Fig. 7, model 5 assumes cool/wet  
 711 | conditions produce an increase in RSF activity, resulting in a strong rising trend  
 712 | through the late Holocene with fluctuations culminating in a Little Ice Age maximum  
 713 | of RSF activity.

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 715 | *A new model of Holocene SRSF activity in Jotunheimen*

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 717 | Based on analysis of Holocene SRSF activity in Jotunheimen and comparison with  
 718 | regional climatic and geo-environmental indicators, a new thermally-driven,  
 719 | permafrost-degradation model is proposed (Fig. 7, model 6). This model is  
 720 | characterized by several key elements:

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- 722 | • minimal activity following deglaciation in the early Holocene;
- 723 | • maximum activity late in the mid Holocene on the multi-millennial timescale;
- 724 | • declining activity through the late Holocene with a second minimum close to
- 725 | the present;
- 726 | • secondary fluctuations on multi-centennial to millennial timescales throughout
- 727 | the Holocene;
- 728 |

729 | This pattern of change bears little relationship to any of the previous models,  
 730 | which are clearly inappropriate in the context of [these](#) data. Model 1 can be rejected  
 731 | for Jotunheimen on the basis of the  $\chi^2$  tests in Table 5. Although there is an element of  
 732 | randomness in our data, and earthquakes do occasionally occur in this part of southern  
 733 | Norway, their magnitudes [tend to be](#) too low to be effective in triggering SRSFs  
 734 | inland from the seismically more active coastal and off-shore areas (cf. Bungum et al.  
 735 | 2000; [Fjeldskaar et al. 2000](#); [Hicks et al. 2000](#); Olesen et al. 2000; Blikra et al. 2006).  
 736 | Moreover, there is no sign of a dominant early-Holocene activity peak in our  
 737 | [histogram or change detection analysis](#), which is the characteristic feature of the two  
 738 | deglaciation-related models (3 and 4). Absence of an early peak may well be  
 739 | accounted for by considerable thinning of the Late Weichselian Ice Sheet prior to final  
 740 | deglaciation in Jotunheimen (Goehring et al. 2008; Mangerud et al. 2011; Hughes et  
 741 | al. 2016; Stroeven et al. 2016), which is likely to have reduced the scale of any  
 742 | paraglacial effects on RSFs after ~10.0 ka. For example, over half (56%) of the  
 743 | estimated glacio-isostatic rebound of 160 m that has taken place in Jotunheimen since  
 744 | 12.0 ka was completed prior to 10.0 ka and a further quarter (26%) by 6.0 ka (Lyså et  
 745 | al. 2008). Finally, the temporal pattern of SRSF activity in Jotunheimen is negatively  
 746 | correlated with model 5, which indicates that cool/wet conditions should be rejected  
 747 | as the major cause of enhanced SRSF activity. Instead, this inverse pattern points to  
 748 | the counterintuitive conclusion that enhanced activity is linked to relatively warm



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3 749 | climatic conditions.

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5 751 *Association of SRSF activity with the thermal climate record*

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7 753 The possible associations between enhanced Holocene SRSF activity and relatively  
8 754 warm climatic conditions can be explored with reference to proxy temperature records  
9 755 | and reconstructions of temperature-sensitive geo-environmental indicators (Fig. 10A-  
10 756 G).

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12 758 The long-term annual air temperature trend for Northern Europe shown in Fig.

13 759 | 10B is a stacked pollen-based reconstruction expressed as deviations from the mean

14 760 (Seppä et al. 2009). The Holocene Thermal Maximum (HTM) is clearly expressed in

15 761 this figure from ~8.0 to 4.0 ka by mean annual temperatures consistently >0.5 °C

16 762 higher than today. Alkenone-based temperature reconstruction similarly documents

17 763 warmest sea-surface temperatures in the North Atlantic at this time (Eldevik et al.

18 764 2014; see also Jansen et al. 2008; Renssen et al. 2012). However, other

19 765 reconstructions based on chironomids (Velle et al. 2010), aquatic macrofossils

20 766 (Väliranta et al. 2015) and megafossils (Dahl & Nesje 1996; Paus & Haugland 2017),

21 767 which are not dependent on tree-pollen production or ocean temperatures, indicate

22 768 | that the highest temperatures probably occurred at 10.0–8.0 ka. Mean summer

23 769 temperatures estimated from pine-tree limits in the Scandes Mountains (Dahl & Nesje

24 770 1996), for example, peak at ~1.5 °C above present temperatures around 9.0 ka (Fig.

25 771 | 10C). An early temperature maximum at ~9.0 ka is also shown in the pollen-based

26 772 reconstruction of July air temperature from Øvre Heimdalsvatnet in the low-alpine

27 773 | belt of eastern Jotunheimen (Fig. 10D, Velle et al. 2010). At this location, a

28 774 temperature of at least 3.5 °C higher than present was attained by 9.0 ka, falling to the

29 775 | long-term Holocene average by 4.0 ka. Comparison with these reconstructions

30 776 | indicates that (1) SRSF frequency increased during the HTM and (2) maximum

31 777 activity was not reached until late in the HTM.

32 778

33 779 Three other palaeorecords can be used to focus on shorter-term warm intervals

34 780 | comparable in scale with our minor phases of enhanced SRSF frequency (Fig. 10E-

35 781 G). The first of these (Fig. 10E), based on a standardized temperature reconstruction

36 782 derived from the record of  $\delta^{18}\text{O}$  in the GISP 2 Greenland ice core (Alley 2004;

37 783 Wanner et al. 2011, their Fig. 1a), shows periods of above average air temperature.

38 784 | Fig. 10F, based on the North Atlantic standardized stacked ocean ice-rafted debris

39 785 (IRD) record (Bond et al. 2001; Wanner et al. 2011, their Fig. 3a), shows periods

40 786 between IRD events, when sea-surface temperatures are likely to have been above the

41 787 | long-term average. Both sets of warm periods demonstrate only moderate agreement

42 788 | between themselves and with our minor phases of enhanced SRSF frequency. There is

43 789 poorer agreement (particularly in the late Holocene after ~3.0 ka) with the final

44 790 record, which relates to variations in the size of mountain glaciers in the study area

45 791 | (Fig. 10G). Glacier variations are widely accepted as climate indicators that reflect, in

46 792 part, temporal variations in summer temperature, especially in the case of glaciers in

47 793 continental locations where winter precipitation variations tend to be less effective

48 794 than in maritime regions (Oerlemans 2005; Bakke et al. 2008; Nesje et al. 2008;

49 795 Winkler et al. 2010). Local glacier variations in the Smørstabbtindan massif,

50 796 Jotunheimen, which is centrally located in relation to the sites of our SRSF events in a

51 797 relatively continental region of southern Norway, exhibit at least nine Holocene time

52 798 intervals when the glaciers were smaller than they are today, including a prolonged

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3 799 | period from ~7.8 to 4.8 ka, which includes most of the HTM (Fig. 10G; Matthews &  
4 800 | Dresser 2008).

5 801  
6 802 | Thus, overall, a strong case can be made for linking millennial-scale variations  
7 803 | in SRSF activity to the thermal environment. However, causal mechanisms are  
8 804 | required to answer the following questions: (1) why was maximum SRSF activity  
9 805 | attained late in the mid-Holocene, rather than earlier in the HTM when temperatures  
10 806 | were at a maximum; and (2) why was there not a closer relationship between the  
11 807 | minor phases of enhanced SRSF activity and shorter-term warm periods, such as the  
12 808 | Mediaeval, Roman and Bronze Age warm periods, in particular during the late-  
13 809 | Holocene? We propose that permafrost degradation, and climate-dependent variation  
14 810 | in permafrost depth, can explain the temporal pattern of SRSF activity and, in  
15 811 | particular, the departure of the temporal pattern of SRSF activity from a simple  
16 812 | 'warm-climate' model.

17 813

#### 18 814 | Conditionality of SRSF activity on permafrost degradation

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20 816 | To interpret the results of both the change detection analysis and Meyer wavelet  
21 817 | analysis, a modelled permafrost record for Fennoscandia (Kukkonen & Šafanda 2001)  
22 818 | is used (Fig. 8B). This provides a basis for attributing SRSF activity in Jotunheimen  
23 819 | to permafrost degradation by focusing on relative changes to permafrost depth in  
24 820 | bedrock over the last ~10 ka. The 5% porosity model was selected for comparison as  
25 821 | this is more representative than the 0% porosity model given the numerous fractures  
26 822 | that lead to slope instability and SRSFs. The permafrost model shows a significant  
27 823 | decrease in depth beginning at ~8 ka and reaching a steady 'shallow' equilibrium by  
28 824 | ~5 ka. Permafrost is relatively stable from 5 ka until ~0.6 ka when permafrost depth  
29 825 | increases. This permafrost model is subdivided into five distinct periods and is related  
30 826 | to the SRSF record as follows:

31 827

32 828 | Phase 1: 10.0–8.1 ka ('stable phase'). – SRSF frequency is in equilibrium with  
33 829 | permafrost, which is at its maximum depth, with no alarms detected in the change  
34 830 | detection analysis and no low-order oscillations in the Meyer wavelet record. Bedrock  
35 831 | permafrost is stable throughout this period and is used to define background Holocene  
36 832 | depth. In this phase, persistent bedrock permafrost acts to stabilize slopes and limits  
37 833 | major SRSF activity.

38 834

39 835 | Phase 2: 8.1–4.8 ka ('transition phase'). – Progressive warming throughout the mid-  
40 836 | Holocene, as recorded in palaeo-climate reconstructions, acts to decrease permafrost  
41 837 | depth. In response, there is a minor progressive decrease in negative change detection  
42 838 | rates and increase in positive change detection within  $2\sigma$ . This trend is matched by  
43 839 | Meyer wavelet analysis, with a progressive increase in SRSF frequency above the  
44 840 | Holocene background rate. In this phase, a gradual (~3 ka) but clear transition from  
45 841 | 'deeper' to 'shallower' permafrost (~28% depth change) is matched by a minor  
46 842 | increase in SRSF frequency and may explain the minor phases of enhanced SRSF  
47 843 | activity identified during this period. Moreover, this gradual change in permafrost  
48 844 | depth, as opposed to a stochastic response to climate warming, provides a compelling  
49 845 | explanation for the significant lag between SRSF activity and the HTM.

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51 847 | Phase 3: 4.8–2.6 ka ('peak phase'). – Permafrost depth is more-or-less stable and  
52 848 | remains close to its minimum Holocene depth for ~2 ka. This period is matched by

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3 849 SRSF activity, as change detection analysis records a significant, sustained and  
4 850 positive rate of change ( $> 2\sigma$ ) for  $\sim 2.2$  ka, with a maximum attained at  $\sim 4.3$  ka and  
5 851 with SRSF frequency significantly exceeding the average frequency until  $\sim 3.3$  ka ( $>$   
6 852  $6\sigma$ ). This change is matched by the Meyer wavelet record, with a peak at  $\sim 4.6$  ka and  
7 853 a gradual decline to the Holocene background rate at  $\sim 2.5$  ka. In this phase, persistent  
8 854 shallow permafrost may directly influence SRSF occurrence by (1) actively  
9 855 destabilizing bedrock cliffs and causing slope failure and/or by (2) weakening bedrock  
10 856 cliffs and making them more susceptible to other trigger factors.

11 857  
12 858 Phase 4: 2.6–0.6 ka ('exhaustion phase'). – Permafrost depth remains relatively stable  
13 859 and shallow for  $\sim 2$  ka, with no significant deviation from modelled depths during the  
14 860 'peak phase'. However, there is a clear decrease in SRSF frequency after the mid-  
15 861 Holocene peak with a return to the Holocene background rate, as revealed by both  
16 862 change detection and Meyer wavelet analysis. In this phase, we propose that bedrock  
17 863 cliffs have reached a new equilibrium with permafrost, as the majority of slopes that  
18 864 can fail under these permafrost conditions have failed by this time; that is the supply  
19 865 of 'potentially failable' cliffs is exhausted. As a result, SRSF occurrence returns to an  
20 866 average frequency comparable with the 'stable phase' of the early Holocene.

21 867  
22 868 Phase 5: 0.6 - 0.1 ka ('stabilization phase'). – Contrary to the dominant Holocene  
23 869 trend, this short-term late-Holocene phase shows a clear increase in permafrost depth  
24 870 after  $\sim 0.6$  ka. This transition is coeval with a statistically significant decrease in SRSF  
25 871 frequency ( $> 2\sigma$ ) while Meyer wavelet analysis records the continued decrease in  
26 872 frequency below the Holocene background level. These data suggest that an increase  
27 873 in bedrock permafrost depth directly controls SRSF activity by stabilizing slopes and  
28 874 decreasing the susceptibility of bedrock cliffs to direct or indirect failure.

29 875  
30 876 The correlation between SRSF frequency and permafrost depth in bedrock as  
31 877 modeled by Kukkonen & Šafanda (2001) provides a compelling explanation for the  
32 878 low-frequency variations in SRSF activity during the Holocene and, in particular for:

- 33 879  
34 880 • the significant departure from mean Holocene SRSF frequency at the end of  
35 881 the mid Holocene;  
36 882 • the lag between the HTM and the SRSF frequency peak;  
37 883 • the low SRSF frequency in the early Holocene; and  
38 884 • the marked decline in SRSF frequency near the end of the late Holocene (after  
39 885  $\sim 0.6$  ka).

40 886  
41 887 These explanations are supported by change detection analysis and (d<sub>6</sub>) Meyer  
42 888 wavelet analysis. A causal link between SRSF frequency and regional permafrost  
43 889 degradation is also supported by the close match between the altitudinal distribution  
44 890 of the 92 SRSFs and the current aspect-dependent lower altitudinal limit of permafrost  
45 891 in rock faces in the Galdhøpiggen massif (Hipp et al. 2014). Approximately 87% (n =  
46 892 80) of SRSFs occur within  $\pm 300$  m of the limit and  $\sim 62\%$  (n = 57) are  $\leq 200$  m below  
47 893 this limit. A small number of SRSFs are found above the permafrost limit ( $\sim 16\%$ ; n =  
48 894 15) but the majority are restricted to within  $\leq 50$  m above this limit. These data imply  
49 895 a causal relationship between SRSF occurrence and the time-dependent degradation  
50 896 and aggradation of bedrock permafrost during the Holocene, as driven by climate and  
51 897 locally controlled by aspect. Based on an altitudinal lapse rate of  $0.6$  °C per 100 m in  
52 898 mean annual air temperatures (MAAT), this implies that all SRSF sites would have

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3 899 been in the permafrost zone when temperatures were 3.0 °C lower than today. It is  
4 900 likely, therefore, that much of the permafrost that had survived or developed in SRSF  
5 901 cliffs following deglaciation would have degraded during the HTM when MAAT is  
6 902 likely to have reached 2.0–3.0 °C warmer than at present and when permafrost limits  
7 903 would have been correspondingly higher (Lilleøren et al. 2012).  
8 904

9 905 Higher-frequency changes in SRSF activity as reflected by weighted age-  
10 906 frequency (Fig. 6C) and ( $d_1 - d_5$ ) wavelet analysis (see Supplementary Fig. 1) can be  
11 907 interpreted as representing Holocene background SRSF frequency after removal of  
12 908 the mid-Holocene peak of the change detection analysis (Fig. 8A). These higher  
13 909 frequency changes are more challenging to interpret, given the limited availability of  
14 910 palaeo-environmental records (e.g. seasonal paleo-precipitation data, storm-event  
15 911 chronologies, palaeoseismic and groundwater flux records) and the inherent SHD age  
16 912 uncertainties. The conceptual models related to deglaciation and characterized by  
17 913 early-Holocene peak activity (Fig. 9) can be discounted as these bear limited  
18 914 resemblance to the chronology of SRSF events.  
19 915

20 916 Changes in permafrost depth might be expected to play a role in explaining the  
21 917 higher-frequency changes. However, we cannot preclude a contribution to higher-  
22 918 frequency variability from the continuity, earthquake, and cool/wet climate conceptual  
23 919 models (Fig.9). Thawing permafrost may be a direct trigger factor for SRSF events  
24 920 due, for example, to loss of strength or elevated hydrostatic pressure, or it may render  
25 921 the rock slope susceptible to other triggers involving meltwater from spring snow melt  
26 922 or extreme rainfall events in summer (Gruber et al. 2004; Gruber & Haeberli 2007;  
27 923 Krautblatter et al. 2013; Blikra & Christiansen 2014; Draebing et al. 2014;  
28 924 Krautblatter & Leith 2015; Messenzehl & Dikau 2017). Extreme summer rainfall  
29 925 events, which are likely to have been more frequent during warm periods, have been  
30 926 implicated in triggering debris-flow events in Leirdalen (Matthews et al. 2009) and  
31 927 might have triggered some SRSFs.  
32 928

#### 33 929 *Further conceptual and methodological implications*

34 930  
35 931 Thus, the timing of SRSFs in this study, with fluctuating SRSF activity rising to a  
36 932 sustained peak at the transition from the mid- to late-Holocene, suggests the  
37 933 importance of progressive but intermittent permafrost degradation lagging behind the  
38 934 highest temperatures of the Holocene. Subsequent declining SRSF frequencies, in  
39 935 contrast, appear to signal exhaustion of the supply of failable cliffs and/or renewed  
40 936 aggradation of permafrost.  
41 937

42 938 These fundamental findings recognize that Holocene SRSF activity in  
43 939 Jotunheimen essentially reflects paraperiglacial processes: that is, it is a conditional  
44 940 response to the transition from a permafrost to a seasonal-freezing climatic regime as  
45 941 permafrost depth decreases (cf. Mercier 2008; Scarpozza 2016; Matthews et al. 2017).  
46 942 While this model is primarily applicable to the SRSFs sampled in this study, it could  
47 943 be tested in comparable mountain regions. In particular, links between permafrost  
48 944 degradation and enhanced slope failure may explain SRSF frequency in regions with  
49 945 comparable seismotectonics, glaciation and deglaciation histories or climatic trends.  
50 946 Robust SRSF chronologies would need to be constructed to test the model, either  
51 947 using radiometric methods (e.g.  $^{10}\text{Be}$ ) or calibrated-age dating techniques (e.g. SHD).  
52 948

949 Our new SRSF chronology indicates, moreover, that SHD can be used to  
 950 generate reliable SRSF chronologies, although further work is necessary to verify this  
 951 technique by directly comparing age estimates for individual landforms derived from  
 952 both SHD and radiometric methods.

953  
 954 Finally, the recognition of a causal link between climate, permafrost  
 955 degradation and enhanced slope instability may have important implications for  
 956 glacial and periglacial environments under global warming scenarios. In particular,  
 957 while widespread retreat of mountain ice caps and valley glaciers may trigger initial  
 958 slope instability, our data suggest that the geomorphological impact of current climatic  
 959 and deglacial trends and, in particular, the slow transition from glacial to periglacial,  
 960 and to seasonal-freezing climatic regimes, may have a long-lasting impact on  
 961 mountain environments.

## 962 963 964 Conclusions

965  
 966 (1) We have developed an approach to the exposure-age dating of a large sample of  
 967 rock-slope failures, which involves adapting Schmidt-hammer exposure-age dating  
 968 (SHD) as a calibrated-age dating technique to the specific characteristics of small  
 969 rock-slope failures (SRSFs). SHD has provided an effective and low-cost method for  
 970 constructing a regional Holocene chronology of SRSFs (12 to 2520 m<sup>3</sup>) in the alpine  
 971 zone of Jotunheimen.

972  
 973 (2) Focusing on a large sample of SRSFs enables the detection of temporal variations  
 974 in the frequency and magnitude of events through the Holocene. Modes in a weighted  
 975 age-frequency distribution at ~8.9, 7.3, 5.9 and 4.5 ka were substantiated by  
 976 probability density function analysis, which produced individual Gaussian age  
 977 distributions of  $9.00 \pm 1.13$  ka,  $7.38 \pm 0.99$  ka,  $6.40 \pm 0.77$  ka and  $4.50 \pm 1.42$  ka.  
 978 Based on this analysis, SRSF activity was relatively low following deglaciation in the  
 979 early Holocene and attained a maximum towards the end of the mid Holocene (~4.5  
 980 ka). Peak SRSF activity lagged behind the Holocene Thermal Maximum by at least  
 981 ~2.2 ka and declined thereafter with a very low frequency of events during the last  
 982 millennium.

983  
 984 (3) Using change detection and discreet Meyer wavelet analysis in combination with  
 985 proxy temperature indicators and an existing permafrost depth model, we propose that  
 986 enhanced SRSF activity was primarily controlled by permafrost degradation. As a  
 987 result, the Holocene permafrost depth record is subdivided into five distinct periods  
 988 and related to the SRSF chronology as follows:

- 989  
990 • 10 - 8.1 ka – ‘stable phase’ – low SRSF activity; maximum Holocene  
991 permafrost depth.
- 992 • 8.1 - 4.8 ka – ‘transition phase’ – increasing susceptibility to SRSF activity;  
993 decreasing permafrost depth.
- 994 • 4.8 - 2.6 ka – ‘peak phase’ – maximum SRSF activity; minimum Holocene  
995 permafrost depth.
- 996 • 2.6 - 0.6 ka – ‘exhaustion phase’ – decreasing SRSF activity; little change in  
997 shallow permafrost depth.

- 0.6 - 0.1 ka – ‘stabilization phase’ – minimum SRSF activity; increasing permafrost depth.

(4) Long-term relative change in permafrost depth provides a compelling explanation for the high-magnitude departures from the SRSF background rate. In particular, the gradual change in permafrost depth during the ‘transition phase’, as opposed to a stochastic response to climate warming, accounts for the significant lag (~2.2 ka) between the Holocene Thermal Maximum and the SRSF frequency peak. Moreover, persistent shallow permafrost during the ‘peak phase’ may be the key driver behind SRSF occurrence by (a) actively destabilizing bedrock cliffs and causing slope failure and/or (b) weakening bedrock cliffs and making them more susceptible to other trigger factors.

(5) Conversely, declining SRSF frequency during the ‘exhaustion phase’ appears to reflect the diminished supply of potentially failable cliffs, even under a shallow permafrost depth scenario. Finally, low frequency of SRSF occurrence during the ‘stabilization phase’ likely reflects an increase in permafrost depth (permafrost aggradation) after ~0.6 ka; a change which would have been sufficient to stabilize slopes and decrease the susceptibility of bedrock cliffs to direct or indirect failure.

(6) This interpretation is supported by geomorphological evidence, given the consistent location of SRSF sites in relation to the local aspect-dependent lower altitudinal limit of permafrost in cliff faces. This new paraperiglacial model attributes enhanced SRSF activity to progressive and intermittent permafrost degradation during Holocene warm periods, including the possibility of renewed aggradation of permafrost during short-term cold periods and renewed degradation during the ensuing warm periods

(7) Our new thermally-driven, permafrost-degradation model of SRSF events in Jotunheimen bears little similarity to existing models of Holocene RSF activity. However, while aspects of this new model require further testing by other methods and in other regions, the results of this study have important implications for climate-change forcing of RSF activity. Projected mean annual global warming is predicted to decrease the area of mountain permafrost and raise lower altitudinal permafrost limits. This in turn will likely destabilize higher bedrock slopes and increase SRSF frequency there. The delayed response of peak SRSF frequency to warming climate, as modulated by permafrost depth, may therefore result in a long-lasting impact of current climate trends on mountain environments.

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For Review Only

**FIGURE CAPTIONS**

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*Fig. 1.* Location map: numbers and open circles identify the studied SRSFs; sites of control points are shown by crosses.

*Fig. 2.* Photographs of selected small rock-slope failures (SRSFs): (A) No. 23, Gravidalen; (B) Nos 7 and 8, Leirdalen; (C) Nos 34-36, Bjørndalen; (D) No. 7, Sognefjell; (E) and (F) No. 22, Gravidalen (also the site of a young control point).

*Fig. 3.* Schematic of the fan-scar-cliff comparison tests with expected differences in mean R-values between fan boulders, scar bedrock surfaces, unfailed cliffs, and rock surfaces used as younger and older control-point surfaces. Expectations apply to single-event SRSF events without the possible complications discussed in the text.

*Fig. 4.* Frequency distributions of four SRSF characteristics: (A) fan volume; (B) altitude; (C) aspect; (D) mean R-value. Eight sites in gabbroic gneiss (Sognefjell) are differentiated by solid black shading from 84 sites in pyroxene-granulite gneiss.

*Fig. 5.* Calibration curves and calibration equations for (A) pyroxene-granulite gneiss and (B) gabbroic gneiss. Note that both calibration curves are based on two control points of known age (25 years and 9700 years) using data presented in Table 3.

*Fig. 6.* Holocene SHD chronologies of SRSF activity for Jotunheimen: (A) individual SHD dates with their 95% confidence intervals in the different subregions; (B) age-frequency distributions of SRSF events at the regional level using 2000-yr, 1000-yr, 500-yr and 200-yr time intervals; (C) weighted age-frequency distribution with age-frequency curve defined by binomial smoothing; (D) variation in the magnitude of SRSF events based on rock volume using 200-year time intervals. Vertical bands (numbered) are the 4 modes in the weighted age-frequency distribution suggesting phases of enhanced regional SRSF activity.

*Fig. 7.* Probability density function analysis of SRSF activity for Jotunheimen: (A) histogram and KS density PDE; (B) individual Gaussian age distributions (n = 5), the sum of which integrates to the cumulative PDE with a model fit that is graphically indistinguishable from from the PDE model. The number of ages listed for each Gaussian age distribution (#) exceeds the total number of SRSF events identified in Jotunheimen as some ages contribute to >1 Gaussian distribution; (C) peak Gaussian numerical ages and 1 $\sigma$  uncertainties for the five individual Gaussian age distributions plotted against the peak probability density (PPD). The PPD scales with the number and spatial clustering of individual ages. Reported RSF volumes are based on the sum of individual SRSF volumes (m<sup>3</sup>) which comprise each Gaussian age distribution; (D) distribution of SRSF ages, sorted by oldest to youngest. The 42 SRSF events which account for the dominant mode at 4.50  $\pm$  1.42 ka (within 1 $\sigma$ ) are highlighted.

*Fig. 8.* Change detection and related analyses: (A) Cumulative sum change detection graph showing positive (blue) and negative (orange) changes and statistically significant departures (> 2 $\sigma$ ) from the background SRSF frequency; (B) modelled permafrost depth in Fennoscandia (5% porosity) from Kukkonen & Šafanda (2001), subdivided into five distinct phases; (C) results of discreet Meyer wavelet analysis, showing the lowest frequency decomposed signal (d<sub>6</sub>).

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4 1797 | *Fig. 9.* Models for different patterns and causes of Holocene variations in SRSF  
5 1798 frequency and/or magnitude: (1) continuity-of-activity; (2) intermittent-earthquakes;  
6 1799 (3) deglaciation-close-tracking; (4) deglaciation-lagging; (5) cool/wet-climate-  
7 1800 response; and (6) the new thermally-driven permafrost-degradation model proposed in  
8 1801 this study for SRSFs in Jotunheimen. The subdivisions of the Holocene shown are  
9 1802 those proposed by Walker et al. (2012).  
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11 1804 | *Fig. 10.* Relationships between SRSF frequency in Jotunheimen and proxy climatic  
12 1805 records: (A) temporal variations in SRSF frequency from Fig. 6C; (B) pollen-based  
13 1806 reconstruction of annual air temperature for Northern Europe expressed as deviations  
14 1807 from the mean (Seppä et al., 2009); (C) mean summer air temperature deviations from  
15 1808 present in the Scandes Mountains based on pine tree-limit variations (Dahl and Nesje,  
16 1809 1996); (D) pollen-based July air temperature variations at Øvre Heimdalsvatnet,  
17 1810 eastern Jotunheimen (Velle et al., 2010); (E) periods of above average air temperature  
18 1811 (shaded) based on the GISP 2 Greenland ice core  $\delta^{18}\text{O}$  record (Alley, 2004; Wanner et  
19 1812 al., 2011); (F) periods of above average sea-surface temperatures in the North Atlantic  
20 1813 Ocean (shaded) based on standardized stacked ice-rafted debris (IRD) records (Bond  
21 1814 et al., 2001; Wanner, et al., 2011); (G) periods when glaciers in the Smørstabbtindan  
22 1815 massif, Jotunheimen, were smaller than today (shaded) based on glaciolacustrine and  
23 1816 glaciofluvial stratigraphy (Matthews and Dresser, 2008). Vertical bands indicate  
24 1817 phases of enhanced regional SRSF frequency (as in Fig. 6).  
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### **SUPPLEMENTARY MATERIAL**

*(Caption to be included with figure)*

30 1823  
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33 1826 | Supplementary Fig. 1: Full results of discreet Meyer wavelet analysis, showing all six  
34 1827 decomposed signals (green), ranging from high ( $d_1$ ) to low frequency ( $d_6$ ), of which  
35 1828 the latter represents the only single event structure of Holocene SRSF activity. The  
36 1829 blue curves ( $a_1 - a_5$ ) represent the cumulative aggregation of the decomposed signals  
37 1830 ( $d_1 - d_6$ ) where  $a_6$  represents the mean background rate of SRSF occurrence ( $0.92 \pm$   
38 1831  $0.20$ ), which is identical to the Holocene mathematical mean. The sum of all  
39 1832 decomposed signals results in a model ( $S_m$ ) that is identical to the 100 yr bin  
40 1833 histogram data ( $S_d$ ).  
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1846 *Table 1.* Data on the 92 small rock-slope failures (SRSFs) located in Jotunheimen:  
 1847 Leirdalen (Nos 1-29); Bjørndalen (30-40); Gravidalen (41-68); Høgvaglura (69-72);  
 1848 Visdalen (73-80); Veodalen (81-84); Sognefjell (85-92).  
 1849 Abbreviations:  $L$  = fan length;  $W$  = fan width;  $V$  = fan volume; SD = standard  
 1850 deviation of R-values;  $C_s$  = error associated with the dated surface;  $C_c$  = error  
 1851 associated with the calibration equation; CI = confidence interval for the SHD age  
 1852 based on the total error ( $C_t$ ).  
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No.	$L$ (m)	$W$ (m)	$V$ (m <sup>3</sup> )	Altitude (m a.s.l.)	Aspect	Mean R-value	SD	$C_s$ (yr)	$C_c$ (yr)	SHD age (yr $\pm$ 95% CI)	Sub-region
1	70	25	1050	1420	West	45.0	9.90	1047	513	7018 $\pm$ 1124	Leirdalen
2	80	20	960	1440	West	44.51	8.80	930	414	7277 $\pm$ 1018	
3	15	9	81	1400	West	39.69	9.47	1001	458	9833 $\pm$ 1101	
4	90	40	2160	1160	West	41.53	9.57	1012	445	8857 $\pm$ 1105	
5	15	8	72	1030	West	43.26	10.03	1060	426	7940 $\pm$ 1143	
6	90	20	1080	1160	West	43.62	10.23	1081	423	7749 $\pm$ 1161	
7	8	25	120	1140	West	44.69	9.41	995	412	7182 $\pm$ 1077	
8	30	25	450	1140	West	46.59	10.35	1094	392	6175 $\pm$ 1162	
9	8	8	38	1135	West	47.28	8.63	912	385	5809 $\pm$ 990	
10	15	25	225	1135	West	44.68	8.85	936	412	7187 $\pm$ 1022	
11	30	20	360	1200	North	52.38	10.07	1064	333	3105 $\pm$ 1115	
12	50	25	750	1425	North	46.49	8.63	912	394	6228 $\pm$ 994	
13	15	25	225	960	East	51.50	8.34	882	342	3572 $\pm$ 946	
14	50	25	750	955	East	49.79	10.74	1135	360	4478 $\pm$ 1191	
15	70	60	2520	950	East	49.28	8.73	923	365	4749 $\pm$ 992	
16	50	25	750	1290	West	48.29	9.98	1055	375	5273 $\pm$ 1120	
17	20	40	480	1320	West	54.10	9.90	1047	315	2193 $\pm$ 1093	
18	20	15	180	1320	West	57.53	10.15	1073	280	375 $\pm$ 1109	
19	30	40	720	1320	West	55.95	8.61	910	296	1213 $\pm$ 957	
20	18	14	151	1120	East	48.79	8.43	891	370	5008 $\pm$ 965	
21	16	8	77	1120	East	44.40	8.29	876	415	7336 $\pm$ 970	
22	25	14	210	1130	East	48.93	9.11	963	368	4934 $\pm$ 1031	
23	40	13	312	1170	East	41.30	9.14	966	447	8979 $\pm$ 1065	
24	25	15	225	1180	East	40.82	9.16	968	296	9233 $\pm$ 1012	
25	15	13	117	1180	East	43.37	9.49	1003	426	7882 $\pm$ 1090	
26	20	4	48	1190	East	44.86	8.70	920	410	7092 $\pm$ 1007	
27	12	8	58	1240	East	49.28	10.53	1113	365	4749 $\pm$ 1171	
28	20	4	48	1240	East	45.92	10.98	1160	399	6530 $\pm$ 1227	
29	22	4	53	1200	East	47.15	8.24	871	387	5878 $\pm$ 953	
30	90	16	864	1370	East	44.27	10.65	1126	416	7405 $\pm$ 1200	Bjørndalen
31	30	15	270	1380	East	44.62	10.10	1068	413	7219 $\pm$ 1145	
32	30	10	180	1380	East	52.60	8.62	911	331	2989 $\pm$ 922	
33	75	30	1350	1360	East	54.91	8.30	877	307	1764 $\pm$ 930	
34	30	15	270	1380	East	49.87	7.53	796	359	4436 $\pm$ 873	
35	30	12	216	1380	East	49.46	7.84	829	363	4653 $\pm$ 905	
36	20	30	360	1380	East	50.19	8.61	910	355	4266 $\pm$ 977	
37	80	35	1680	1330	South	50.23	9.57	960	355	4245 $\pm$ 1024	
38	25	15	225	1300	North	54.07	6.73	711	316	2209 $\pm$ 778	
39	50	30	900	1305	North	55.37	7.95	840	302	1520 $\pm$ 893	
40	25	25	375	1300	North	53.30	8.20	867	323	2617 $\pm$ 925	Gravidalen
41	55	20	660	1480	West	49.43	8.11	857	363	4669 $\pm$ 931	
42	15	35	315	1480	West	55.49	6.69	707	301	1456 $\pm$ 769	
43	65	15	585	1480	West	51.11	8.40	888	346	3778 $\pm$ 953	
44	60	15	540	1470	West	50.84	7.05	745	349	3922 $\pm$ 823	
45	65	25	975	1470	West	50.01	8.85	936	357	4362 $\pm$ 1001	
46	30	15	270	1460	West	52.57	7.97	843	331	3004 $\pm$ 905	
47	75	20	900	1460	West	53.03	6.27	663	326	2761 $\pm$ 739	
48	25	30	450	1430	South	50.01	7.00	740	357	4362 $\pm$ 822	
49	17	8	82	1440	South	49.10	8.45	893	367	4844 $\pm$ 964	
50	40	15	360	1440	South	49.71	7.72	816	360	4521 $\pm$ 892	
51	15	10	90	1440	South	50.38	7.78	822	356	4165 $\pm$ 896	
52	15	6	54	1400	South	56.21	7.38	780	293	1075 $\pm$ 834	
53	10	5	30	1400	South	57.99	6.22	658	275	131 $\pm$ 713	
54	7	8	34	1360	South	47.32	8.00	846	385	5788 $\pm$ 929	
55	10	6	36	1280	South	40.31	10.14	1072	457	9504 $\pm$ 1165	
56	12	5	36	1440	South	48.82	8.12	858	370	4992 $\pm$ 935	
57	6	5	18	1440	South	47.43	7.72	816	384	5729 $\pm$ 902	
58	8	8	38	1440	South	51.63	7.70	814	341	3503 $\pm$ 882	
59	4	5	12	1440	South	51.12	6.62	700	346	3773 $\pm$ 781	
60	7	4	17	1480	South	48.02	7.43	785	378	5416 $\pm$ 872	

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61	20	5	60	1480	South	52.10	11.98	1266	336	3254 ± 1310	
62	14	8	67	1480	South	46.17	9.02	953	399	6397 ± 1033	
63	6	12	43	1430	South	48.74	8.09	855	370	5035 ± 932	
64	10	5	30	1430	South	46.99	7.65	809	388	5963 ± 897	
65	14	3	25	1460	South	49.91	8.38	886	358	4415 ± 956	
66	15	4	36	1520	South	51.92	8.34	882	338	3349 ± 944	
67	6	4	14	1540	South	49.95	9.74	1030	358	4393 ± 1090	
68	10	5	30	1540	South	49.37	7.08	748	364	4701 ± 832	
69	20	15	180	1550	East	50.13	7.74	818	356	4298 ± 892	Hogvaglura
70	50	12	360	1550	East	45.16	10.05	1062	407	6933 ± 1138	
71	20	10	120	1540	East	46.35	8.94	945	395	6302 ± 1024	
72	20	10	120	1540	East	42.10	11.92	1260	439	8555 ± 1334	
73	15	4	36	1420	East	47.03	11.08	1171	388	5941 ± 1234	Visdalen
74	15	9	81	1420	East	50.70	10.47	1107	350	3996 ± 1161	
75	10	4	24	1420	East	54.42	9.47	1001	312	2024 ± 1049	
76	25	10	150	1260	West	49.96	10.20	1078	358	4388 ± 1136	
77	40	20	480	1200	East	51.37	10.30	1089	343	3641 ± 1142	
78	70	30	1260	1200	East	52.98	8.86	937	327	2787 ± 992	
79	35	20	420	1200	East	51.57	7.93	838	341	3535 ± 905	
80	60	8	288	1190	East	50.31	10.75	1136	354	4202 ± 1190	
81	55	40	1320	1350	South	53.33	8.72	922	323	2601 ± 977	Veodalen
82	45	12	324	1340	South	54.33	9.34	987	313	2071 ± 1036	
83	50	25	750	1330	South	51.56	10.15	1073	341	3540 ± 1126	
84	45	40	1080	1330	South	49.46	10.60	1121	363	4653 ± 1178	
85	6	10	36	1375	East	37.17	11.29	900	239	9412 ± 931	Sognefjell
86	7	5	21	1425	South	38.53	8.82	703	244	8868 ± 744	
87	6	6	22	1425	South	39.42	10.43	831	247	8513 ± 868	
88	10	5	30	1430	South	41.38	9.86	786	255	7729 ± 826	
89	6	5	18	1430	South	40.73	9.47	755	253	7989 ± 796	
90	16	10	96	1450	South	38.26	8.83	704	243	8976 ± 745	
91	9	5	27	1435	West	42.33	10.03	800	259	7349 ± 840	
92	10	7	42	1370	East	36.95	9.14	729	238	9500 ± 766	

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1887 *Table 2.* Radiocarbon age control for deglaciation in the study area.

Location	Altitude (m. a.s.l.)	<sup>14</sup> C age ± 1σ (years BP)	Calibrated age* (cal. years BP)	Reference
<i>Leirdalen/Bjørndalen</i>				
Lower Leirdalen	920	9089 ± 61	10426 – 10170 (94.8%)	Barnett et al. 2000
Bøverkinnhalsen	1020	8570 ± 60	9677 – 9475 (95.4%)	Nesje & Dahl 2001
Bjørndalen	1250	8760 ± 100	10066 – 9547 (77.1%)	Matthews et al. 2005
<i>Sognefjell</i>				
Nedre Hervavatnet	1287	8695 ± 75	9921 – 9530 (94.6%)	Hormes et al. 2009
Gjuvvatnet	1248	8885 ± 140	10247 – 9557 (95.4%)	Karlén & Matthews 1992

\* Most probable range with probability in brackets

1913 *Table 3.* Control point data: **values** used for calibration equations are **indicated** in  
 1914 bold. *Abbreviations:* Gneiss = pyroxene-granulite gneiss; Gabbro = gabbroic gneiss;  
 1915 Combined = data combined from two replicate sites; SD = standard deviation; CI =  
 1916 confidence interval; n = sample size.

Control point	Geology	Type	Age (years)	Mean R-value	SD	95% CI	n
Gravdalen	Gneiss	SRSF	25	58.22	6.29	0.72	300
Gravdalen	Gneiss	Road cut	25	58.15	6.56	0.75	300
<b>Gravdalen</b>	<b>Gneiss</b>	<b>Combined</b>	<b>25</b>	<b>58.19</b>	<b>6.42</b>	<b>0.52</b>	<b>600</b>
<b>Sognefjell</b>	<b>Gabbro</b>	<b>Road cut</b>	<b>25</b>	<b>60.65</b>	<b>7.26</b>	<b>0.83</b>	<b>300</b>
Gravdalen	Gneiss	Bedrock	9700	39.71	4.80	1.25	60
Leirdalen	Gneiss	Bedrock	9700	40.19	4.69	1.22	60
<b>SE Smørstabbtindan</b>	<b>Gneiss</b>	<b>Combined</b>	<b>9700</b>	<b>39.94</b>	<b>4.79</b>	<b>0.87</b>	<b>120</b>
Leirbreen	Gabbro	Bedrock	9700	35.78	2.84	0.74	60
Bøverbreen	Gabbro	Bedrock	9700	37.12	3.53	0.92	60
<b>W Smørstabbtindan</b>	<b>Gabbro</b>	<b>Combined</b>	<b>9700</b>	<b>36.45</b>	<b>3.25</b>	<b>0.59</b>	<b>120</b>

1927 *Table 4.* Comparative R-values from fans, scars and unfailed cliffs associated with  
 1928 selected SRSFs. **Further information on these six SRSFs are provided in Table 1.**

No.	Fan	Scar	Unfailed cliff
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		Mean			SD			95% CI		
		Mean	SD	95% CI	Mean	SD	95% CI	Mean	SD	95% CI
1931										
1932										
1933										
1934										
1935	5	43.26	10.03	2.00	41.34	7.75	1.55	42.20	7.86	1.57
1936	46	51.63	7.70	1.54	51.32#	8.10	1.62	41.63†	9.20	1.83
1937	47	51.12	6.62	1.32	54.05#	8.05	1.69	43.26†	10.19	2.03
1938	51	37.17*	11.29	2.25	42.89	9.73	1.94	38.54	10.37	2.07
1939	58	36.95*	9.14	1.82	43.99	10.44	2.08	40.68	12.30	2.45
1940	81	49.96*	10.20	2.03	54.47#	8.07	1.60	43.38†	10.78	2.15

\* Fan significantly different from scar (p<0.05)

# Scar significantly different from unfailed cliff (p<0.05)

† Unfailed cliff significantly different from fan (p<0.05)

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For Review Only

1 Small rock-slope failures conditioned by Holocene permafrost  
 2 degradation: a new approach and conceptual model based on Schmidt-  
 3 hammer exposure-age dating, Jotunheimen, southern Norway

4  
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23  
 24 Rock-slope failures (RSFs) constitute significant natural hazards but the geophysical  
 25 processes which control their timing are poorly understood. However, robust  
 26 chronologies can provide valuable information on the environmental controls on RSF  
 27 occurrence: information which can inform models of RSF activity in response to  
 28 climatic forcing. This paper uses Schmidt-hammer exposure-age dating (SHD) of  
 29 boulder deposits to construct a detailed regional Holocene chronology of the  
 30 frequency and magnitude of small rock-slope failures (SRSFs) in Jotunheimen,  
 31 Norway. By focusing on the depositional fans of SRSFs ( $\leq 10^3 \text{ m}^3$ ), rather than on the  
 32 corresponding features of massive RSFs ( $\sim 10^8 \text{ m}^3$ ), 92 single-event RSFs are targeted  
 33 for chronology building. A weighted SHD age-frequency distribution and probability  
 34 density function analysis indicate four centennial- to millennial-scale periods of  
 35 enhanced SRSF frequency, with a dominant mode at  $\sim 4.5 \text{ ka}$ . Using change detection  
 36 and discreet Meyer wavelet analysis, in combination with existing permafrost depth  
 37 models, we propose that enhanced SRSF activity was primarily controlled by  
 38 permafrost degradation. Long-term relative change in permafrost depth provides a  
 39 compelling explanation for the high-magnitude departures from the SRSF background  
 40 rate and accounts for (1) the timing of peak SRSF frequency, (2) the significant lag  
 41 ( $\sim 2.2 \text{ ka}$ ) between the Holocene Thermal Maximum and the SRSF frequency peak  
 42 and (3) the marked decline in frequency in the late-Holocene. This interpretation is  
 43 supported by geomorphological evidence, as the spatial distribution of SRSFs is  
 44 strongly correlated with the aspect-dependent lower altitudinal limit of mountain  
 45 permafrost in cliff faces. Results are indicative of a causal relationship between  
 46 episodes of relatively warm climate, permafrost degradation and the transition to a  
 47 seasonal-freezing climatic regime. This study highlights permafrost degradation as a  
 48 conditioning factor for cliff collapse, and hence the importance of paraperiglacial  
 49 processes; a result with implications for slope instability in glacial and periglacial  
 50 environments under global warming scenarios.

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4 52  
5 53 *Key words:* small rock-slope failures (SRSFs), Schmidt-hammer exposure-age dating  
6 54 (SHD), permafrost degradation, Holocene Thermal Maximum, climate-change  
7 55 impacts, paraperiglacial processes, southern Norway  
8 56  
9 57

10 58 Rock-slope failures (RSFs) are indicative of instability in the landscape and reflect  
11 59 several geophysical processes and potential trigger factors related to rock mechanics,  
12 60 geomorphology, hydrology and environmental change. Moreover, RSFs constitute  
13 61 significant natural hazards. As a result, understanding the environmental controls on  
14 62 RSF occurrence provides crucial information which can inform modelling of future  
15 63 RSF activity in response to climate forcing (Rapp 1960a, 1960b; Brunsdén & Prior  
16 64 1984; Evans et al. 2006; Clague & Stead 2012; Davies 2015).  
17 65

18 66 Numerous RSFs have been investigated in regions of high relief and, in some  
19 67 cases, RSF deposits have been dated (e.g. Korup et al. 2007; Ballantyne et al. 2014a,  
20 68 2014b). However, previous research has primarily focused on modern examples,  
21 69 spectacular cases or small numbers of massive rock-slope failures (MRSFs;  $\sim 10^8 \text{ m}^3$ )  
22 70 which, in combination with uncertainty associated with current geochronological  
23 71 approaches, limits our understanding of the fundamental geophysical processes and  
24 72 environmental controls that determine RSF occurrence. Particular studies of RSFs  
25 73 have used a variety of techniques and, on some occasions, a combination of  
26 74 geochronological methods (Lang et al. 1999; Hermanns et al. 2000; Crosta & Clague  
27 75 2009; Deline & Kirkbride 2009; Prager et al. 2009; Pánek 2014; Böhme et al. 2015;  
28 76 Moreiras et al. 2015; Mercier et al. 2017), but the opportunities for accurate dating are  
29 77 relatively rare.  
30 78

31 79 The primary method for numerical-age dating of RSF deposits is terrestrial  
32 80 cosmogenic nuclide dating (TCND;  $^{10}\text{Be}$ ,  $^{26}\text{Al}$ ,  $^{36}\text{Cl}$ ) as this technique permits direct  
33 81 sampling and age determination of the exposed rock surfaces associated with RSFs  
34 82 (Hermanns et al. 2001, 2004, 2017; Cossart et al. 2008; Dortch et al. 2009; Ivy-Ochs  
35 83 et al. 2009; Penna et al. 2011; Ballantyne & Stone 2013; Ballantyne et al. 2013,  
36 84 2014a, 2014b; Böhme et al. 2015; Schleier et al. 2015, 2017). However, the high  
37 85 financial cost of this technique limits its routine application which, in turn, often  
38 86 prevents statistically robust identification and rejection of erroneous results (Tomkins  
39 87 et al. 2018b). Consequently, there are still few reliable chronologies of RSFs which  
40 88 limits our understanding of the environmental factors determining their spatial and  
41 89 temporal occurrence.  
42 90

43 91 In this paper we develop a methodology for the investigation and dating of  
44 92 RSFs, with targeted study of 'small rock-slope failures' (SRSFs;  $< 10^3 \text{ m}^3$ ). This focus  
45 93 has the advantage over MRSFs of permitting the dating and study of a relatively large  
46 94 sample of simple, likely single-event RSFs within a specified region. The  
47 95 methodology has been developed in conjunction with the relatively new calibrated-  
48 96 age dating technique of Schmidt-hammer exposure-age dating (SHD) (Shakesby et al.  
49 97 2006, 2011; Matthews & Owen 2011; Matthews et al. 2015; Matthews & Wilson  
50 98 2015; Winkler et al. 2010, 2016; Wilson et al. 2017). SHD has the potential to  
51 99 estimate the numerical age of rock-surface exposure at low cost with comparable  
52 100 accuracy and precision, and greater representativeness, than TCND over the Late  
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3 101 Glacial and Holocene (cf. Winkler 2009; Winkler & Matthews 2010; Matthews &  
4 102 Winkler 2011; Matthews et al. 2013; Wilson & Matthews 2016; Tomkins et al. 2016,  
5 103 2018a, 2018b, 2018c).

6 104  
7 105 Specific objectives of this paper are three-fold:

- 8 106  
9 107
  - To establish a Holocene chronology of SRSF events in the alpine zone of
  - 10 108 Jotunheimen, southern Norway and identify any phases of instability;
  - 11 109 • To explore relationships between the timing of Holocene SRSF events and
  - 12 110 regional environmental changes, including climatic changes; and
  - 13 111 • To develop further the potential of SHD as a calibrated-age dating technique
  - 14 112 in the context of RSFs.

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16 114

### 17 114 18 115 Study area and environmental context

19 116  
20 117 SRSFs were investigated in a broad area of northern Jotunheimen, the highest  
21 118 mountain massif in southern Norway, which culminates in Galdhøpiggen (2469 m  
22 119 above sea level; a.s.l.). The study area extends from Sognefjell in the west to  
23 120 Veodalen in the east (Fig. 1). Most SRSFs were found in Leirdalen, Bjørndalen (a  
24 121 western tributary valley to upper Leirdalen) and Gravidalen. The SRSFs occurred over  
25 122 an altitudinal range of 600 m (950-1550 m a.s.l.), mainly above the tree line, which  
26 123 lies at ~1000-1100 m a.s.l., in the alpine zone, and mainly in the low- and mid-alpine  
27 124 belts (Moen 1999). Examples of SRSFs from the study area are shown in Fig. 2.

28 125  
29 126 Climatic data from the Sognefjell meteorological station (1413 m a.s.l.)  
30 127 indicate a mean annual air temperature of +3.1 °C (mean July temperature +13.4 °C;  
31 128 mean January temperature -10.7 °C), and a mean annual precipitation of 860 mm,  
32 129 much of which occurs as snow (climatic normals AD 1961-1990; Aune 1993; Førland  
33 130 1993). These data are consistent with a lower altitudinal limit of discontinuous  
34 131 permafrost at ~ 1450 m a.s.l. in the Galdhøpiggen massif (Ødgård et al. 1992; Isaksen  
35 132 et al. 2002; Farbrot et al. 2009; Lilleøren et al. 2012) with permafrost limits rising  
36 133 eastwards as continentality increases (Etzelmüller et al. 2003; Ginås et al. 2017).  
37 134 However, Hipp et al. (2014) have demonstrated a large difference of several hundred  
38 135 metres in the lower limits of permafrost between north- and south-facing rock walls.  
39 136 In the Galdhøpiggen massif, the lower altitudinal limit of rock-wall permafrost is  
40 137 located at 1500-1700 m a.s.l. in south-facing rock walls but only 1200-1300 m a.s.l. in  
41 138 shaded, north-facing rock walls (Hipp et al. 2014). Small valley glaciers, cirque  
42 139 glaciers and ice caps are common at and above these altitudes on the surrounding  
43 140 mountain peaks and plateaux (Andreassen & Winsvold 2012).

44 141  
45 142 The metamorphic geology of the region consists primarily of pyroxene-  
46 143 granulite gneiss with peridotite intrusions and quartzitic veins (Battey & McRitchie  
47 144 1973, 1975; Lutro & Tveten 1996), and gabbroic gneiss in the area investigated on  
48 145 Sognefjell (Gibbs & Banham 1979). Only boulders and bedrock of pyroxene-  
49 146 granulite gneiss and gabbroic gneiss were used in this study, as described below.  
50 147 Although these broad lithological categories include quite variable mineralogy, any  
51 148 differences in surface R-values due to lithology will likely be significantly smaller  
52 149 than the effect of variable exposure age given the relatively long Holocene  
53 150 timescales of exposure and limited climatic variability within the study region.

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3 151 Topographically, most of the valley-side slopes have experienced a considerable  
4 152 degree of glacial erosion, although elements of ancient palaeic surfaces are  
5 153 preserved in the landscape (Ahlmann 1922; Gjessing 1967; Lidmar-Bergström et al.  
6 154 2000) due, at least in part, to non-erosive, cold-based conditions during glaciations.  
7  
8 155

9 156 Jotunheimen was located near the position of the main ice-divide and ice-  
10 157 accumulation area of the Scandinavian ice-sheet at the maximum of the Last  
11 158 (Weichselian) Glaciation. Deglaciation of the main valleys is likely to have occurred  
12 159 by ~9.7 ka, following the Erdalen Event, late in the Preboreal chronozone (Dahl et  
13 160 al. 2002; Matthews & Dresser 2008; Velle et al. 2010). Most glaciers appear to have  
14 161 melted away during the Holocene Thermal Maximum (Nesje 2009) when permafrost  
15 162 limits were also higher than today (Lilleøren et al. 2012), but regenerated during  
16 163 neoglaciation, certainly by 5.5 ka and possibly as early as 7.6 ka (Ødgård et al.  
17 164 2017). Both neoglaciation and lowering of permafrost limits occurred as a result of  
18 165 climatic deterioration (cooler and wetter) in the late Holocene, culminating in the  
19 166 Little Ice Age glacier maximum of the eighteenth century (Matthews 1991, 2005;  
20 167 Matthews & Dresser 2008). Future predicted mean annual warming of 0.3-0.4 °C  
21 168 per decade in Scandinavia (Benestad 2005) is likely to lead to unprecedented glacier  
22 169 retreat (Nesje et al. 2008) and a continuing rise in permafrost limits (Lilleøren et al.  
23 170 2012).  
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26 172

27 173 Methodology

28 174

29 175 *Definitions and criteria for recognition of SRSFs*

30 176

31 177 The term ‘rock-slope failure’ (RSF) refers to both (1) a mass-movement process  
32 178 involving the deformation and loss of integrity of a volume of intact bedrock followed  
33 179 by its *en masse* collapse and downslope movement under gravity and (2) the resulting  
34 180 landform. This definition is used here to distinguish RSF from ‘rockfall’ – the  
35 181 smaller-scale process involving the piecemeal detachment and free fall of individual  
36 182 rock particles – even though the term rockfall is commonly used at all scales,  
37 183 including the largest landslides and rock avalanches (MRSFs), which are often  
38 184 complex and multiphase (cf. Bates & Jackson 1987; Cruden & Varnes 1996; Braathen  
39 185 et al. 2004; Luckman 2004; Evans et al. 2006; Hermanns et al. 2006; Jarman 2006;  
40 186 Frattini et al. 2012; Hermanns & Longva 2012; Shakesby 2014; Brideau & Roberts  
41 187 2015).  
42 188

43 188

44 189 Fundamental to this study was the selection of SRSF landforms that  
45 190 represented, as far as it was possible to ascertain, the product of single events. Criteria  
46 191 for recognition of such SRSFs were as follows:  
47 192

48 192

- 49 193
- 50 194 • a compact and coherent depositional fan of predominantly angular boulders  
51 195 located close to a bedrock cliff.
  - 52 196 • a simple erosional scar in the cliff, immediately upslope of the fan, which is  
53 197 comparable in scale to the fan and therefore represents the likely source of the  
54 198 failed rock material;
  - 55 199 • an absence of alternative sources of boulders up-slope of the scar.

56 199

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3 200 Although no upper limit was placed on the size of the SRSFs recognized in this study,  
4 201 these criteria become less easily satisfied as RSFs increase in size. The lower size  
5 202 limit was the practical one of sufficient boulders for reliable Schmidt hammer  
6 203 measurement. Thus, the size range included in the study was determined by the RSFs  
7 204 in the region. Furthermore, the 92 investigated cases represent the whole population  
8 205 of SRSFs that satisfied the above criteria in the study area.

#### 9 206 10 207 *Measurement of SRSF characteristics*

11 208  
12 209 Estimates were made in the field of the length and average width of the depositional  
13 210 fan of each SRSF. Aspect and the altitude of the fan apex were estimated from  
14 211 topographic maps at a scale of 1:50,000 with a contour interval of 20 m,  
15 212 supplemented by altimeter and GPS measurements in the field. Fan volume was  
16 213 calculated from the length and average width measurements, assuming an average fan  
17 214 thickness of 1 m and a voids fraction (volume of voids/total fan volume) of 40%.  
18 215 Although some of the largest fans are thicker than 1 m in places, all are thinly spread  
19 216 across and down slope and rarely involve piles of debris. Lower voids fractions have  
20 217 generally been used for MRSFs, rock avalanches, talus and other mass movement  
21 218 types involving mixed particle sizes, fine matrix and/or compacted material (Owen et  
22 219 al. 2010; Sass & Wollny 2001; Hungr & Evans 2004; Wilson 2009; Stock &  
23 220 Uhrhammer 2010; Sandøy et al. 2017). The value of 40% is justified given the  
24 221 absence of fine matrix (Fig. 2) and lack of compaction, and its compatibility with  
25 222 similar values for clean, open-graded, angular aggregate material used as backfill in  
26 223 foundation engineering (StormTech 2012; cf. Dann et al. 2009).

#### 27 224 28 225 *Measurement of Schmidt-hammer R-values*

29 226  
30 227 N-type mechanical Schmidt hammers (Proceq 2004; Winkler & Matthews 2014) were  
31 228 used to measure rebound (R-) values from 100 boulders in each depositional fan. R-  
32 229 values reflect lithologically-determined rock hardness and the compressive strength of  
33 230 the rock surface: hence, R-values decline following exposure of a rock surface to  
34 231 subaerial weathering. For boulder surfaces of the same lithology but differing age, R-  
35 232 values therefore reflect the exposure age (time elapsed since exposure) of the rock  
36 233 surface. Use of one impact per boulder from a large sample of boulders ensures that  
37 234 the R-value frequency distribution can be used to approximate the boulder-age  
38 235 distribution (Matthews et al. 2014, 2015).

39 236  
40 237 Precautions taken to eliminate or reduce possible sources of uncertainties and  
41 238 errors in Schmidt-hammer measurement included avoiding unstable or small boulders,  
42 239 boulder or bedrock edges, joints or cracks, unusual lithologies and lichen-covered or  
43 240 wet surfaces (cf. Shakesby et al. 2006; Matthews & Owen 2010; Viles et al. 2011).  
44 241 Rock surfaces were not cleaned or artificially abraded prior to impact with the  
45 242 Schmidt hammer (cf. the carborundum treatment of Viles et al. 2011) because such  
46 243 treatment would likely remove age-related weathering effects. However, there is  
47 244 continued debate as to whether rock surfaces should be abraded prior to testing  
48 245 (Moses et al. 2014) although a consistent sampling approach may enable age-related  
49 246 information to be retained (c.f. Tomkins et al. 2018b). Where possible, horizontal  
50 247 boulder surfaces were impacted but only vertical rock faces were available on cliffs.  
51 248 The two hammers used had been recently re-calibrated at a recognised service centre  
52 249 and were tested frequently on the manufacturer's test anvil throughout the study to

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3 250 ensure there had been no deterioration in instrument performance following large  
4 251 numbers of impacts (cf. McCarroll 1987, 1994; Winkler & Matthews 2016).  
5 252 Measurements at 84 sites were restricted to rock surfaces of pyroxene-granulite  
6 253 gneiss. At the 8 sites on Sognefjell, gneissic rocks with gabbroic textures were used,  
7 254 which necessitated a separate calibration equation (see below).  
8 255

#### 9 256 *Testing the validity of the approach*

10 257

11 258 In order to test the validity of our approach, and especially whether the boulders  
12 259 comprising the depositional fans actually represent single rock-failure events and  
13 260 whether the local source of the boulders had been correctly identified, R-value  
14 261 distributions associated with six fans and their corresponding scars were investigated.  
15 262 Two separate tests of validity were conducted.  
16 263

17 264 First, in the *fan-scar comparison test*, a comparable sample of R-values ( $n =$   
18 265 100) from the surface of the corresponding scar was compared with the R-value  
19 266 distribution of the fan to identify whether or not the scar was the likely source of the  
20 267 boulders in the fan. If the scar was indeed the source of the boulders, the expectation  
21 268 would be no significant difference in the R-values derived from the scar and its  
22 269 corresponding fan because both would have experienced exposure over the same  
23 270 period of time.  
24 271

25 272 Second, the *unfailed-cliff test* required a comparable sample of R-values ( $n =$   
26 273 100) from the adjacent intact (unfailed) bedrock cliff and also aimed to establish that  
27 274 the cliff was the bedrock source for the fan boulders. If this was the case, it would be  
28 275 expected that R-values from the unfailed cliff would be similar to or lower than the R-  
29 276 values of both the scar and the fan. Any departure from these expectations would  
30 277 indicate possible flaws in our approach.  
31 278

32 279 The principles behind the fan-scar comparison test and the unfailed-cliff test  
33 280 are illustrated in Fig. 3, which also shows the expected relationships between R-  
34 281 values from the fans and R-values from the rock surfaces used as control points in the  
35 282 calibration equations.  
36 283

#### 37 284 *Calibrated-age dating using SHD*

38 285

39 286 Although there was earlier use of the Schmidt hammer for dating purposes (e.g.  
40 287 Matthews & Shakesby 1984; Nesje et al. 1994; Aa & Sjøstad 2000; Aa et al. 2007),  
41 288 SHD has been developed more recently as a calibrated-age dating technique (Colman  
42 289 et al., 1987) incorporating measures of uncertainty based on statistical confidence  
43 290 intervals (cf. Shakesby et al. 2006; Matthews & Owen 2011; Matthews & Winkler  
44 291 2011; Matthews & McEwen 2013). Critically, this involves the derivation of a  
45 292 calibration equation and confidence limits for age.  
46 293

47 294 The calibration equation is based on linear regression of surface age ( $Y$ ) on  
48 295 mean R-value ( $X$ ):  
49 296

$$50 297 Y = a + bX \quad (1)$$

51 298



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3 299 A linear relationship can be justified on both theoretical and empirical grounds.  
4 300 Although chemical weathering rates are likely to decline over longer timescales  
5 301 (Colman 1981; Colman & Dethier 1986; Stahl et al. 2013; Tomkins et al. 2018a,  
6 302 2018b), near-linear rates can be expected over the Holocene timescale, especially  
7 303 where relatively resistant lithologies are subject to relatively slow rates of chemical  
8 304 weathering in a periglacial environment (André 1996, 2002; Nicholson 2008, 2009;  
9 305 Matthews & Owen 2011; Matthews et al. 2016). Although physical (freeze-thaw)  
10 306 weathering is well known in periglacial environments, it is highly dependent on  
11 307 moisture availability for ice-lens growth (Hallet et al. 1991; Hall et al. 2002; Murton  
12 308 et al. 2006; Matsuoka & Murton 2008) and there is no evidence that it has affected the  
13 309 well-drained surfaces used in this study (neither boulders in the dated depositional  
14 310 fans nor bedrock control surfaces).  
15 311

16 312 Furthermore, Shakesby et al. (2011) specifically tested the linearity  
17 313 assumption in relation to granite boulders on independently-dated staircases of raised  
18 314 beaches deposited since 10.4 ka in northern Sweden, with the conclusion that the  
19 315 relationship between mean R-value and age was best described by a linear function.  
20 316 The same conclusion can be reached from age-calibration curves in the British Isles  
21 317 (Tomkins et al. 2018a) and the Pyrenees (Tomkins et al. 2018b), which are based on  
22 318 54 and 52 <sup>10</sup>Be TCND-dated granitic surfaces respectively, all associated with glacial  
23 319 depositional or erosional landforms (moraine boulders or ice-sculpted bedrock).  
24 320 While the Pyrenean age-calibration curve is clearly non-linear over the full age range  
25 321 of ~50 ka, both age-calibration curves evidence linearity over the last ~20 ka. Other  
26 322 studies that have suggested non-linear relationships have involved long timescales  
27 323 and/or have had insufficient control points to test the linearity assumption rigorously  
28 324 over the Holocene timescale (e.g. Betts & Latta 2000; Sánchez et al. 2009; Černá &  
29 325 Engel 2011; Stahl et al. 2013).  
30 326

31 327 Based on two control points, the *b* coefficient can be defined as:  
32 328

$$33 329 \quad b = (y_1 - y_2) / (x_1 - x_2) \quad (2)$$

34 330  
35 331 where  $x_1$  and  $x_2$  are the mean R-values of the older and younger control points,  
36 332 respectively, and  $y_1$  and  $y_2$  are their respective ages. Once the *b* coefficient is known,  
37 333 the *a* coefficient is found by substitution in equation (1). Only two control points of  
38 334 widely differing age are available from Jotunheimen (see below). Provided they are of  
39 335 good quality, however, two control points are sufficient for accurate R-value  
40 336 calibration provided the underlying relationship between R-value and age is  
41 337 approximately linear.  
42 338

43 339 For a landform produced by a single event, the SHD age resulting from this  
44 340 calibration is the average age of the surface boulders and hence the landform age  
45 341 (Matthews et al. 2015). Confidence intervals for the SHD age (95%) are calculated as  
46 342 the total error ( $C_t$ ) by combining the error associated with the calibration equation ( $C_c$ )  
47 343 with the sampling error associated with the surface to be dated ( $C_s$ ):  
48 344

$$49 345 \quad C_t = \sqrt{(C_c^2 + C_s^2)} \quad (3)$$

$$50 346 \quad C_c = C_o - [(C_o - C_y) (R_s - R_o) / (R_y - R_o)] \quad (4)$$

51 348

$$C_s = b[ts / \sqrt{(n-1)}] \quad (5)$$

where  $C_o$  and  $C_y$  are the 95% confidence intervals of the older and younger control points (in years); and  $R_o$ ,  $R_y$  and  $R_s$  are the mean R-values of the older control point, the younger control point and the surface to be dated, respectively.  $C_s$  depends on the number of R-value impacts on the surface to be dated (sample size,  $n$ ), the standard deviation of those impacts ( $s$ ), and Student's  $t$  statistic. Thus, the confidence interval ( $C_t$ ) associated with any SHD age depends not only on the sample sizes used to establish the calibration equation and characterize the surface to be dated but also the natural variability exhibited by all the rock surfaces involved.

#### *Control points for calibration equations*

For this study, we constructed separate calibration equations for rock surfaces composed of pyroxene-granulite gneiss and gabbroic gneiss (each equation based on two control points). Data for the older control points, which relate to glacially-scoured bedrock surfaces, were taken from Matthews & Owen (2010). Their data from four sites in Leirdalen and Gravdalen (S and E Smørstabbtindan) were used for the pyroxene-granulite gneiss calibration equation: four sites near Leirbreen and Bøverbreen, close to Sognefjell (W Smørstabbtindan) supplied the data for the gabbroic gneiss calibration equation (Fig. 1).

Evidence for deglaciation of these sites is provided by basal  $^{14}\text{C}$  dates from peat bogs and lakes in Leirdalen, Bjørndalen, and on Sognefjell (Table 2). These  $^{14}\text{C}$  dates were recalibrated to calendar age ranges with the OxCal online program (v.4.3) using the IntCal13 calibration dataset (Reimer et al. 2013). Although one of the calibrated age ranges is significantly older, 9.7 ka is the only date for deglaciation that is compatible with the other four  $^{14}\text{C}$  dates. Use of 9.7 ka as the age of the old control points for SHD calibration can be justified on the further grounds that it is the expected date for termination of the Erdalen Event in neighbouring regions (Dahl et al. 2002) and is consistent with empirical evidence for and large-scale modeling of deglaciation in southern Norway (Dahl et al. 2002; Goehring et al. 2008; Nesje 2009; Mangerud et al. 2011; Hughes et al. 2016; Stroeven et al. 2016). Thus, the potential errors in the old control points appear to be small in relation to the calibration errors ( $C_c$  and  $C_s$ ) that are taken fully into account in this study.

Calibration equations given in Matthews & Owen (2010) for these rock types could not be used because their younger control points were derived from glacially-abraded surfaces from glacier forelands. Such smooth surfaces are not appropriate as a source of young control points for dating the exposure-age of boulders originating from SRSFs, which are rougher in texture yielding lower R-values than abraded surfaces of the same age (Shakesby et al. 2006; Matthews & McEwen 2013; Matthews et al. 2015). In contrast, after prolonged weathering, originally smooth surfaces are expected to yield similar R-values, and hence SHD ages, to initially rough surfaces.

Young control points with similar roughness properties to fresh boulder surfaces derived from SRSFs were therefore sought. These included: (1) boulders and bedrock surfaces produced by a recent rock-slope failure in Gravdalen and (2) bedrock exposed recently in road cuts in Gravdalen and on Sognefjell (Fig. 1). Both types of surfaces have been shown in previous studies to yield R-values that are

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3 399 statistically indistinguishable from each other provided sufficient care is taken to  
4 400 impact only truly fresh rock surfaces (Matthews & Wilson 2015; Matthews et al.  
5 401 2016). Furthermore, both types of recent rock surfaces used as young control points in  
6 402 this study were lichen-free and hence were assigned a maximum exposure age of 25  
7 403 years based on various estimates of the time required for the establishment (ecesis) of  
8 404 crustose lichens on bedrock surfaces in this environment (Matthews 2005; Matthews  
9 405 & Owen 2008; Matthews & Vater 2015). Errors in the age of the young control point  
10 406 are therefore considered to be negligible in the context of this study.  
11 407

#### 12 408 *Chronology construction and analysis*

13 409  
14 410 Holocene chronologies of SRSF events were constructed from the SHD ages of the 92  
15 411 SRSF fans using a number of statistical approaches. First, graphical analysis of age-  
16 412 frequency distributions used 2000-yr, 1000-yr, 500-yr and 200-yr time intervals to  
17 413 define major clusters of SHD ages and hence possible multi-centennial to millennial  
18 414 phases of enhanced SRSF frequency (Matthews et al. 2009; Matthews & Seppälä  
19 415 2015). Based on the same events weighted according to their rock volume, a second  
20 416 chronology was constructed showing the changing magnitude of SRSF events through  
21 417 the Holocene.  
22 418

23 419 In order to take account of dating uncertainty, a weighted age-frequency  
24 420 distribution was constructed in which each SHD age was plotted over five 200-yr age  
25 421 classes: a weight of 4 was used for the central class; the second and fourth classes  
26 422 were weighted 2. Thus, the SHD age was plotted over a range of 1000 yr, consistent  
27 423 with the average 95% confidence interval of  $\pm 991$  yr calculated for the 92 SRSF fans  
28 424 (see below). One-sample  $\chi^2$  tests were used to test the hypothesis that the dated events  
29 425 were sampled from an underlying population of events with an even distribution  
30 426 through time.  
31 427

32 428 To support weighted age-frequency analysis, the distribution of calculated  
33 429 SRSF ages was analysed using probability density function analysis. Probability  
34 430 density estimates (PDEs) were produced and modelled to separate out individual  
35 431 Gaussian distributions using the KS density kernel in MATLAB (2015) and a  
36 432 dynamic smoothing window based on age uncertainty (cf. Dortch et al. 2013). The  
37 433 sum of individual Gaussian distributions integrates to the cumulative PDE at 1000  
38 434 iterations to obtain a good model fit. The goodness of fit between the re-integrated  
39 435 PDE, which is derived from individual Gaussian distributions, and the cumulative  
40 436 PDE, which is derived from the full age dataset, is indicated graphically. PDE  
41 437 analysis was repeated using a number of individual Gaussian distributions ( $n = 1-10$ ).  
42 438 To avoid over-interpretation of SRSF modes, the PDE model with the minimum  
43 439 number of individual Gaussian distributions, which also achieved a good model fit,  
44 440 was selected. This analytical method has primarily been employed in studies using  
45 441  $^{10}\text{Be}$  (cf. Dortch et al. 2013; Murari et al. 2014) or SHD (Barr et al. 2017; Tomkins et  
46 442 al. 2018a; 2018b; 2018c) to account for negative or positive skew of moraine boulder  
47 443 datasets and to identify and reject ages that are compromised by moraine degradation  
48 444 (Briner et al. 2005; Heyman et al. 2011) or nuclide inheritance (Hallet & Putknonen  
49 445 1996). In these applications, PDE analysis and interpretation of individual Gaussian  
50 446 distributions (cf. Fig. 3 in Dortch et al. 2013) is based on the assumption that analysed  
51 447 ages relate to a single event e.g. moraine deposition. This assumption is clearly not  
52 448 applicable to the analysis of SRSF ages, as each numerical age relates to a distinct  
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3 449 event and an individual landform. As a result, individual Gaussian distributions are  
4 450 interpreted as reflecting the temporal clustering of events. The characteristics of  
5 451 individual Gaussian distributions, i.e. the peak probability density, width of PDE tails,  
6 452  $1\sigma$  uncertainties and the number of contributing ages (Fig. 7), were used to assess the  
7 453 significance and temporal clustering of SRSF events in Jotunheimen over the last ~10  
8 454 ka.  
9 455

10 456 The individual distributions resulting from the PDE analysis indicated that  
11 457 further analysis was necessary. Thus, a change detection analysis approach was  
12 458 undertaken in MATLAB (2015) to identify statistically unique events. Change  
13 459 detection analysis utilizes the cumulative sum algorithm (cusum), which is commonly  
14 460 used to detect abrupt change in time series data in fields ranging from seismology  
15 461 (Dera & Shumwayb 1999), remote sensed imagery (Lu et al. 2016), and GPS  
16 462 monitoring (Goudarzi et al. 2013). Parameters were set by using the average  
17 463 frequency and occurrence (~1 occurrence per 100 years) of SRSFs throughout the  
18 464 Holocene to filter out 'background' SRSF occurrence. The alarm limit was set at  $\geq 2$   
19 465 standard errors above background. To further explore the temporal pattern of SRSFs,  
20 466 discrete Meyer wavelet analysis was undertaken in MATLAB (2015) to decompose  
21 467 SRSF occurrence through time. Wavelets are discrete oscillations in both time and  
22 468 amplitude and, as such, are useful for identifying discrete events. Wavelet analysis  
23 469 has been used to identify climate signals from various records including  $\delta O^{18}$  (Lau &  
24 470 Weng 1995), and sea surface temperature (Torrence & Compo 1998). The 100 yr  
25 471 binned SRSF age data was passed through the discrete Meyer wavelet with six levels  
26 472 of deconvolution.  
27 473

28 474 Major and minor changes in SRSF activity were then compared with changes  
29 475 in regional Holocene climatic and other geo-environmental indicators to infer possible  
30 476 causes. Specific analyses were performed to investigate relationships between the  
31 477 occurrence of SRSF events and the lower altitudinal limits of discontinuous  
32 478 permafrost using aspect-dependent limits determined for rock walls in the  
33 479 Galdhøpiggen massif by Hipp et al. (2014). The current (AD 2010-2013) lower limits  
34 480 that were used for rock walls facing north, east, south and west were 1250 m, 1450 m,  
35 481 1600 m and 1450 m, respectively.  
36 482

## 37 483 Results

### 38 484 *Data on the SRSFs*

39 485  
40 486 Data on the size and environmental characteristics of the SRSFs are summarized in  
41 487 Table 1 and Fig. 4. The volume of the fans (Fig. 4A) ranges from 12 to 2520 m<sup>3</sup>, with  
42 488 90% <1000 m<sup>3</sup>, 40% <100 m<sup>3</sup> and a median size of only 180 m<sup>3</sup>. The altitudinal range  
43 489 is 960 to 1550 m a.s.l. (Fig. 4B), with a mean altitude of 1340 m a.s.l. There is a  
44 490 preferred aspect with 43% facing east, 34% facing south and 17% facing west, but  
45 491 only 5% facing north (Fig. 4C).  
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49 495 Schmidt-hammer R-values vary widely between SRSFs (Table 1) and the  
50 496 frequency distribution of mean R-values reveals several important features (Fig. 4D).  
51 497 Mean R-values exhibit a very wide range of >20 units from 37.0 to 57.5. The overall  
52 498 mean R-value across the 92 SRSFs is 48.2 but those R-values associated with  
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3 499 gabbroic gneiss (overall mean R-value 39.4, n = 8) are appreciably lower than the  
4 500 remainder involving pyroxene-granulite gneiss (overall mean R-value 49.1, n = 84).  
5 501 The latter value corresponds closely with the 49-50 modal class for the distribution.  
6 502

#### 7 503 *Control-point data and calibration equations*

8 504  
9 505 Data from the control points (Table 3) indicate widely different mean R-values  
10 506 (differing by at least 20 units) for surfaces that differ in age by ~9700 years. It should  
11 507 also be noted that the overlapping 95% confidence intervals associated with each pair  
12 508 of replicates for particular control points indicate that their mean R-values do not  
13 509 differ significantly from each other. Control surfaces of the same age on different  
14 510 lithologies are, however, characterized by non-overlapping confidence intervals, and  
15 511 thus show significantly different mean R-values and justify the use of separate  
16 512 calibration equations for SRSFs developed in pyroxene-granulite gneiss and gabbroic  
17 513 gneiss. The calibration equations derived from these data for the two lithologies are  
18 514 shown in Figure 5 alongside the linear relationships they represent.  
19 515

#### 20 516 *Fan-scar-cliff comparison tests*

21 517  
22 518 Mean R-values for three of the six fans tested did not differ significantly from the  
23 519 mean R-values of the corresponding scars, in accordance with expectation (Fig. 3 and  
24 520 Table 4). However, three fans (Nos 51, 58 and 81) are characterized by mean R-  
25 521 values that are significantly lower than the mean R-values from their scars. This  
26 522 suggests one or more of four possible explanations: (1) rock surfaces of some  
27 523 boulders in these fans are more weathered because they include the products of older  
28 524 rock failures than those that produced the measured bedrock faces of the scars; (2)  
29 525 some of the measured R-values from boulders in the fans reflect the incorporation of  
30 526 bedrock surfaces that were pre-weathered on the cliff face before the failures  
31 527 occurred; (3) some of the R-values from boulders in the fans reflect the incorporation  
32 528 of inherited structures (e.g. joint planes) that were pre-weathered at depth before the  
33 529 failures occurred; and (4) at least part of the cliff bedrock is more resistant to  
34 530 weathering than the boulder surfaces measured in the fans. Interestingly, no fan  
35 531 exhibits a mean R-value that is significantly greater than that of its corresponding  
36 532 scar. This shows that even where more than one phase of activity seems possible, any  
37 533 blocks that were later removed from the scars were insufficient in number to affect  
38 534 appreciably the mean R-values of the fans.  
39 535

40 536 Comparisons between scars and unfailed cliffs or between fans and unfailed  
41 537 cliffs are entirely in agreement with expectation. In three cases (fan Nos 5, 51 and 58)  
42 538 neither the mean R-values for scars and unfailed cliffs nor the mean R-values for fans  
43 539 and unfailed cliffs differ significantly, suggesting that all the exposed surfaces are of  
44 540 the same age (and relatively old). In the other three cases (fan Nos 46, 47 and 81) the  
45 541 mean R-values of the scars and the fans are both significantly higher than the mean R-  
46 542 values of the unfailed cliffs, confirming the SRSFs are younger than the exposure age  
47 543 of the unfailed cliffs.  
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49 545 Comparison of the mean R-values from unfailed cliffs with the values from  
50 546 the older control points given in Table 3 indicates that unfailed cliff surfaces were  
51 547 exposed during or immediately after deglaciation at ~9700 cal. BP. As all surfaces  
52 548 yielded mean R-values lower than those characteristic of the younger control points  
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3 549 (Table 4), it appears that fan deposition and scar exposure occurred throughout the  
4 550 Holocene and, in some cases, thousands of years after regional deglaciation. As a  
5 551 result, the temporal distribution of fan mean-R-values likely reflects the timing of  
6 552 single-event SRSF activity.  
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#### 8 554 *Temporal variations in SRSF activity*

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10 556 The age of each SRSF event, including its 95% confidence interval, is summarized  
11 557 graphically in Fig. 6A. Although there is some evidence of differences in the age  
12 558 distributions between the different valleys, there is no statistically significant  
13 559 correlation between SRSF age and altitude and no significant difference in age  
14 560 between aspects. The overall mean age of all 92 SRSF events is 5124 years, which  
15 561 equates with an average regional frequency of 1 in 105 years.  
16 562

17 563 Simple age-frequency distributions of the SRSF events within the region as a  
18 564 whole are shown in Fig. 6B. Although these events occurred without any prolonged  
19 565 break in activity, their frequency varied considerably over the last ~10,000 years. The  
20 566 distribution based on 2000-yr time intervals has a single mode indicating an increase  
21 567 in the frequency of events through the early Holocene, a distinct peak in activity in the  
22 568 6.0-4.0 ka time interval, and a consistent decline in activity thereafter. The use of  
23 569 1000-year time intervals reveals two modes – at 8.0-7.0 and 5.0-4.0 ka, respectively.  
24 570 At least three modes can be recognized when 500-yr time intervals are used (at 9.0-  
25 571 8.5, 7.5-7.0 and 4.5-4.0 ka) and many more can possibly be discerned in the  
26 572 distribution based on 200-year time intervals. However, analysis of SRSF modes based on  
27 573 200-year time intervals is not advisable, as this time interval (0.2 ka) is significantly smaller  
28 574 than the typical uncertainty of SRSF ages (~1 ka). Despite this, the hypothesis of an even  
29 575 distribution of SRSF events through time can be rejected at  $p < 0.01$  irrespective of the  
30 576 age classes used.  
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32 578 The weighted age-frequency distribution (Fig. 6C) has four modes (at ~ 8.9,  
33 579 7.3, 5.9 and 4.5 ka), which suggests that only four minor phases of enhanced SRSF  
34 580 frequency are meaningful. Furthermore, according to the weighted distribution, the  
35 581 frequency of events declines steadily after ~4.5 ka with no marked fluctuations.  
36 582

37 583 The temporal pattern in the magnitude of the SRSFs (rock volume), as shown  
38 584 in in Fig. 6D, is substantially the same as the frequency distribution (compare with  
39 585 use of a 200-yr interval in Fig. 6B). In particular, the age-volume distribution has a  
40 586 similar major peak between 4.8 and 4.2 ka, and relatively little activity before 9.0 ka  
41 587 or after 1.0 ka.  
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43 589 Probability density function analysis indicates that the spread of SRSF ages  
44 590 does not conform to a normal distribution (Fig. 7A) and, instead, is best explained by  
45 591 5 individual Gaussian age distributions (Fig. 7B). The sum of individual Gaussian  
46 592 distributions produces a re-integrated PDE which achieves a good model fit with the  
47 593 cumulative PDE. PDE analysis using  $< 5$  individual Gaussian age distributions returns  
48 594 a poor ( $n \leq 3$ ) or sub-optimal ( $n = 4$ ) model fit. PDE analysis using  $> 5$  individual  
49 595 Gaussian age distributions does not therefore significantly improve the model fit and  
50 596 instead risks over-interpretation of the number of SRSF modes. PDE analysis returns  
51 597 peak Gaussian ages (Fig. 7C) of  $9.00 \pm 1.13$  ka ( $n = 14$ ),  $7.38 \pm 0.99$  ka ( $n = 17$ ),  $6.40$   
52 598  $\pm 0.77$  ka ( $n = 14$ ),  $4.50 \pm 1.42$  ka ( $n = 42$ ) and  $1.90 \pm 1.42$  ka ( $n = 18$ ). Although  
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3 599 these modes overlap with adjacent modes within  $1\sigma$ , statistically significant  
4 600 differences between sequential Gaussian age distributions are revealed by two-sample  
5 601 Students t-tests ( $p < 0.01$ ).  
6 602

7 603 These Gaussian age distributions closely match the four modes identified in  
8 604 weighted age-frequency analysis, with a dominant mode at  $\sim 4.5$  ka (Fig. 7B). This  
9 605 mode is the highest probability Gaussian distribution, comprises a significant number  
10 606 of SRSF events ( $n = 42$ ; Fig. 7D) and accounts for a large proportion of total SRSF  
11 607 volume over the last  $\sim 10$  ka ( $18,744 \text{ m}^3$ ). In contrast to weighted age-frequency  
12 608 analysis, PDE analysis returns an additional Gaussian age distribution during the late  
13 609 Holocene at  $\sim 1.9$  ka. However, this is unlikely to reflect a period of enhanced SRSF  
14 610 activity as there is no clear clustering of SRSF ages (Fig. 7A), as evidenced by  
15 611 weighted age-frequency analysis. Instead, late Holocene ages likely reflect declining  
16 612 SRSF activity after the mid-Holocene peak.  
17 613

18 614 The combined results of the age-frequency analyses and the Gaussian  
19 615 separation achieved for PDEs demonstrate that SRSF occurrence through time is non-  
20 616 uniform and multi-modal. Most notable is the high level of occurrence during the mid  
21 617 Holocene, the clear statistical significance of which is confirmed by the results of  
22 618 change detection analysis. The cumulative sum change detection graph (Fig. 8A)  
23 619 shows a clear peak in the rate of SRSF intensity between 4.8 and 2.6 ka, significantly  
24 620 exceeding the  $2\sigma$  threshold, with the largest departure from background occurring at  
25 621 4.3 ka. Conversely, SRSF intensity is significantly reduced beyond the negative  $2\sigma$   
26 622 threshold during the late Holocene at 0.6–0.1 ka. These peaks are a significant  
27 623 departure from the normal rate of occurrence during the Holocene. The three other  
28 624 modes identified above as statistically significant must be regarded as relatively small  
29 625 departures from background SRSF periodicity.  
30 626

31 627 Meyer wavelet analysis was used to explore the two statistically significant  
32 628 departures ( $> 2\sigma$ ) from the background SRSF rate, as identified by change detection  
33 629 analysis. The lowest frequency decomposed signal ( $d_6$ ) is shown in Fig. 8C. The full  
34 630 analysis record is provided in Supplementary Fig. 1.  
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## 37 633 Discussion

### 38 634 *Previous models of the timing of RSFs*

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41 637 Widely different conceptual models can be proposed to describe and explain the  
42 638 temporal distribution of Late Pleistocene and Holocene RSFs. A schematic  
43 639 representation of several models, each of which links a distinctive pattern of change in  
44 640 the frequency and/or magnitude of RSFs to one or more specific causes or triggers, is  
45 641 shown in Fig. 9. Although they have been based mainly on MRSFs, these models are  
46 642 introduced here as a basis for discussion of our Holocene SRSFs. It should be  
47 643 emphasised, moreover, that RSFs may be multicausal and that most if not all of the  
48 644 models have yet to be rigorously tested against data sets with a large number of  
49 645 consistently dated RSFs.  
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51 647 *Model 1.* – The ‘continuity-of-activity model’ proposes that there are no significant  
52 648 temporal variations in the frequency and/or magnitude of RSFs throughout the  
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3 649 Holocene. Despite the small number of dated RSFs available in most studies, few  
4 650 authors have advocated this model. However, the model does appear to be consistent  
5 651 with the temporal distribution of about 60 RSFs located in an extensive area of the  
6 652 Alps centred on the Austrian Tyrol (Prager et al. 2008), which exhibits only limited  
7 653 evidence of temporal clustering at ~10.5-9.4 ka and 4.2-3.0 ka. Prager et al. (2008)  
8 654 attributed the continuity of activity to complex interactions between the processes  
9 655 characterizing models 2-5 together with rock-strength degrading processes such as  
10 656 time-dependent progressive fracture propagation that can both prepare and trigger  
11 657 slope instabilities.  
12 658

13 659 *Model 2.* – The ‘intermittent-earthquakes model’ is applicable to tectonically active  
14 660 regions and assumes that RSFs are triggered directly by large-magnitude earthquakes  
15 661 generated by tectonically-driven uplift or other crustal stresses. Such earthquakes are  
16 662 essentially randomly distributed in time and therefore bear little or no relationship to  
17 663 deglaciation, climate or any of the other potential causative factors in models 3-5 that  
18 664 are effective in tectonically stable regions (see, for example, Fjeldskaar et al. 2000;  
19 665 Hermanns et al. 2001; Keefer 2002, 2015; Hewitt et al. 2008; Antinao & Gosse 2009;  
20 666 Stock & Uhrhammer 2010; Penna et al. 2011; McPhillips et al. 2014; Marc et al.  
21 667 2015; Murphy 2015).  
22 668

23 669 *Model 3.* – The ‘deglaciation-close-tracking model’ is characterised by a dominant  
24 670 peak in RSF activity immediately (i.e. within the first millennium) following regional  
25 671 deglaciation, with subsequent asymptotic decline in activity. The temporal pattern of  
26 672 activity is therefore a typical paraglacial response (cf. Ballantyne 2002). Causal  
27 673 factors that may account for such a pattern include glacial unloading, glacial  
28 674 debuitressing, stress-release fracturing, enhanced groundwater pressure in rock joints  
29 675 and permafrost degradation, all closely associated in time with deglaciation (Fischer  
30 676 et al. 2006; Cossart et al. 2008; McColl 2012; McColl & Davies 2012; Ballantyne et  
31 677 al. 2014a, 2014b; Böhme et al. 2015; Deline et al. 2015; Mercier et al. 2017).  
32 678 Hermanns et al. (2017) found nearly half of 22 dated rock avalanches in southwest  
33 679 Norway occurred within the first millennium following local deglaciation. Although  
34 680 the majority of RSF events occur shortly after deglaciation, some occur much later,  
35 681 due to time-dependent fracture propagation and progressive failure (e.g. Eberhardt et  
36 682 al. 2004; Krautblatter et al. 2013; Phillips et al. 2017). The occurrence of recent RSFs  
37 683 on glacier forelands following the retreat of mountain glaciers from their Little Ice  
38 684 Age maximum limits provides some support for this model (Evans & Clague 1994;  
39 685 Holm et al. 2004; Matthews & Shakesby 2004; Arsenault & Meigs 2005; Allen et al.  
40 686 2010; Stoffel & Huggel 2012).  
41 687

42 688 *Model 4.* – The ‘deglaciation-lagging model’ features a significantly delayed response  
43 689 to deglaciation. Peak RSF activity typically occurs within a few millennia of  
44 690 deglaciation and corresponds with maximum glacio-isostatic rebound (Hicks et al.  
45 691 2000; Ballantyne & Stone 2013; Ballantyne et al. 2013, 2014a, 2014b; Cossart et al.  
46 692 2014; Decaulne et al. 2016). The cause of RSF events is seen as fault reactivation and  
47 693 fracture propagation triggered by earthquakes, the frequency of earthquakes and RSFs  
48 694 generally diminishing through the Holocene as the rate of glacio-isostatic uplift  
49 695 declines.  
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51 697 *Model 5.* – The ‘cool/wet-climate-response model’ applies particularly to the  
52 698 Holocene, reflecting several possible effects of climatic variations on RSF activity.  
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699 Field monitoring, historical documentation and palaeo-studies indicate that  
 700 precipitation variations can be a dominant trigger factor in the timing of RSFs but  
 701 both cooler conditions and indirect effects such as variations in cleft water pressure,  
 702 frost shattering and permafrost degradation have also been implicated in rock-slope  
 703 instability (Eisbacher & Clague 1984; Matthews et al. 1997; Trauth et al. 2000, 2003;  
 704 Dapples et al. 2003; Soldati et al. 2004; Prager et al. 2008; Crozier 2010; Borgatti &  
 705 Soldati 2010; Blikra & Christiansen 2014; Zerathe et al. 2014; Johnson et al. 2017).  
 706 Furthermore, Evans & Clague (1994), Huggel et al. (2010, 2012) and Stoffel &  
 707 Huggel (2012) highlighted the possible effects of recent climate warming on RSFs,  
 708 and direct solar heating of rock faces has also been examined as a possible trigger (cf.  
 709 Allen & Huggel 2013; Collins & Stock 2016). In Fig. 7, model 5 assumes cool/wet  
 710 conditions produce an increase in RSF activity, resulting in a strong rising trend  
 711 through the late Holocene with fluctuations culminating in a Little Ice Age maximum  
 712 of RSF activity.

#### 714 *A new model of Holocene SRSF activity in Jotunheimen*

716 Based on analysis of Holocene SRSF activity in Jotunheimen and comparison with  
 717 regional climatic and geo-environmental indicators, a new thermally-driven,  
 718 permafrost-degradation model is proposed (Fig. 7, model 6). This model is  
 719 characterized by several key elements:

- 721 • minimal activity following deglaciation in the early Holocene;
- 722 • maximum activity late in the mid Holocene on the multi-millennial timescale;
- 723 • declining activity through the late Holocene with a second minimum close to  
 724 the present;
- 725 • secondary fluctuations on multi-centennial to millennial timescales throughout  
 726 the Holocene;

728 This pattern of change bears little relationship to any of the previous models,  
 729 which are clearly inappropriate in the context of these data. Model 1 can be rejected  
 730 for Jotunheimen on the basis of the  $\chi^2$  tests in Table 5. Although there is an element of  
 731 randomness in our data, and earthquakes do occasionally occur in this part of southern  
 732 Norway, their magnitudes tend to be too low to be effective in triggering SRSFs  
 733 inland from the seismically more active coastal and off-shore areas (cf. Bungum et al.  
 734 2000; Fjeldskaar et al. 2000; Hicks et al. 2000; Olesen et al. 2000; Blikra et al. 2006).  
 735 Moreover, there is no sign of a dominant early-Holocene activity peak in our  
 736 histogram or change detection analysis, which is the characteristic feature of the two  
 737 deglaciation-related models (3 and 4). Absence of an early peak may well be  
 738 accounted for by considerable thinning of the Late Weichselian Ice Sheet prior to final  
 739 deglaciation in Jotunheimen (Goehring et al. 2008; Mangerud et al. 2011; Hughes et  
 740 al. 2016; Stroeven et al. 2016), which is likely to have reduced the scale of any  
 741 paraglacial effects on RSFs after  $\sim 10.0$  ka. For example, over half (56%) of the  
 742 estimated glacio-isostatic rebound of 160 m that has taken place in Jotunheimen since  
 743 12.0 ka was completed prior to 10.0 ka and a further quarter (26%) by 6.0 ka (Lyså et  
 744 al. 2008). Finally, the temporal pattern of SRSF activity in Jotunheimen is negatively  
 745 correlated with model 5, which indicates that cool/wet conditions should be rejected  
 746 as the major cause of enhanced SRSF activity. Instead, this inverse pattern points to  
 747 the counterintuitive conclusion that enhanced activity is linked to relatively warm  
 748 climatic conditions.

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3 7494 750 *Association of SRSF activity with the thermal climate record*

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6 752 The possible associations between enhanced Holocene SRSF activity and relatively  
7 753 warm climatic conditions can be explored with reference to proxy temperature records  
8 754 and reconstructions of temperature-sensitive geo-environmental indicators (Fig. 10A-  
9 755 G).

10 756

11 757 The long-term annual air temperature trend for Northern Europe shown in Fig.

12 758 10B is a stacked pollen-based reconstruction expressed as deviations from the mean

13 759 (Seppä et al. 2009). The Holocene Thermal Maximum (HTM) is clearly expressed in

14 760 this figure from ~8.0 to 4.0 ka by mean annual temperatures consistently &gt;0.5 °C

15 761 higher than today. Alkenone-based temperature reconstruction similarly documents

16 762 warmest sea-surface temperatures in the North Atlantic at this time (Eldevik et al.

17 763 2014; see also Jansen et al. 2008; Renssen et al. 2012). However, other

18 764 reconstructions based on chironomids (Velle et al. 2010), aquatic macrofossils

19 765 (Väiliranta et al. 2015) and megafossils (Dahl &amp; Nesje 1996; Paus &amp; Haugland 2017),

20 766 which are not dependent on tree-pollen production or ocean temperatures, indicate

21 767 that the highest temperatures probably occurred at 10.0–8.0 ka. Mean summer

22 768 temperatures estimated from pine-tree limits in the Scandes Mountains (Dahl &amp; Nesje

23 769 1996), for example, peak at ~1.5 °C above present temperatures around 9.0 ka (Fig.

24 770 10C). An early temperature maximum at ~9.0 ka is also shown in the pollen-based

25 771 reconstruction of July air temperature from Øvre Heimdalsvatnet in the low-alpine

26 772 belt of eastern Jotunheimen (Fig. 10D, Velle et al. 2010). At this location, a

27 773 temperature of at least 3.5 °C higher than present was attained by 9.0 ka, falling to the

28 774 long-term Holocene average by 4.0 ka. Comparison with these reconstructions

29 775 indicates that (1) SRSF frequency increased during the HTM and (2) maximum

30 776 activity was not reached until late in the HTM.

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32 778 Three other palaeorecords can be used to focus on shorter-term warm intervals

33 779 comparable in scale with our minor phases of enhanced SRSF frequency (Fig. 10E-

34 780 G). The first of these (Fig. 10E), based on a standardized temperature reconstruction

35 781 derived from the record of  $\delta^{18}\text{O}$  in the GISP 2 Greenland ice core (Alley 2004;

36 782 Wanner et al. 2011, their Fig. 1a), shows periods of above average air temperature.

37 783 Fig. 10F, based on the North Atlantic standardized stacked ocean ice-rafted debris

38 784 (IRD) record (Bond et al. 2001; Wanner et al. 2011, their Fig. 3a), shows periods

39 785 between IRD events, when sea-surface temperatures are likely to have been above the

40 786 long-term average. Both sets of warm periods demonstrate only moderate agreement

41 787 between themselves and with our minor phases of enhanced SRSF frequency. There is

42 788 poorer agreement (particularly in the late Holocene after ~3.0 ka) with the final

43 789 record, which relates to variations in the size of mountain glaciers in the study area

44 790 (Fig. 10G). Glacier variations are widely accepted as climate indicators that reflect, in

45 791 part, temporal variations in summer temperature, especially in the case of glaciers in

46 792 continental locations where winter precipitation variations tend to be less effective

47 793 than in maritime regions (Oerlemans 2005; Bakke et al. 2008; Nesje et al. 2008;

48 794 Winkler et al. 2010). Local glacier variations in the Smørstabbtindan massif,

49 795 Jotunheimen, which is centrally located in relation to the sites of our SRSF events in a

50 796 relatively continental region of southern Norway, exhibit at least nine Holocene time

51 797 intervals when the glaciers were smaller than they are today, including a prolonged

52 798 period from ~7.8 to 4.8 ka, which includes most of the HTM (Fig. 10G; Matthews &amp;

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3 799 Dresser 2008).

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5 801 Thus, overall, a strong case can be made for linking millennial-scale variations  
6 802 in SRSF activity to the thermal environment. However, causal mechanisms are  
7 803 required to answer the following questions: (1) why was maximum SRSF activity  
8 804 attained late in the mid-Holocene, rather than earlier in the HTM when temperatures  
9 805 were at a maximum; and (2) why was there not a closer relationship between the  
10 806 minor phases of enhanced SRSF activity and shorter-term warm periods, such as the  
11 807 Mediaeval, Roman and Bronze Age warm periods, in particular during the late-  
12 808 Holocene? We propose that permafrost degradation, and climate-dependent variation  
13 809 in permafrost depth, can explain the temporal pattern of SRSF activity and, in  
14 810 particular, the departure of the temporal pattern of SRSF activity from a simple  
15 811 'warm-climate' model.

16 812

17 813 *Conditionality of SRSF activity on permafrost degradation*

18 814

19 815 To interpret the results of both the change detection analysis and Meyer wavelet  
20 816 analysis, a modelled permafrost record for Fennoscandia (Kukkonen & Šafanda 2001)  
21 817 is used (Fig. 8B). This provides a basis for attributing SRSF activity in Jotunheimen  
22 818 to permafrost degradation by focusing on relative changes to permafrost depth in  
23 819 bedrock over the last ~10 ka. The 5% porosity model was selected for comparison as  
24 820 this is more representative than the 0% porosity model given the numerous fractures  
25 821 that lead to slope instability and SRSFs. The permafrost model shows a significant  
26 822 decrease in depth beginning at ~8 ka and reaching a steady 'shallow' equilibrium by  
27 823 ~5 ka. Permafrost is relatively stable from 5 ka until ~0.6 ka when permafrost depth  
28 824 increases. This permafrost model is subdivided into five distinct periods and is related  
29 825 to the SRSF record as follows:

30 826

31 827 *Phase 1: 10.0–8.1 ka ('stable phase').* – SRSF frequency is in equilibrium with  
32 828 permafrost with no alarms detected in the change detection analysis and no low-order  
33 829 oscillations in the Meyer wavelet record. Bedrock permafrost is stable throughout this  
34 830 period and is used to define background Holocene depth. In this phase, persistent  
35 831 bedrock permafrost acts to stabilize slopes and limit major SRSF activity.

36 832

37 833 *Phase 2: 8.1–4.8 ka ('transition phase').* – Progressive warming throughout the mid-  
38 834 Holocene, as recorded in palaeo-climate reconstructions, acts to decrease permafrost  
39 835 depth. In response, there is a minor progressive decrease in negative change detection  
40 836 rates and increase in positive change detection within  $2\sigma$ . This trend is matched by  
41 837 Meyer wavelet analysis, with a progressive increase in SRSF frequency above the  
42 838 Holocene background rate. In this phase, a gradual (~3 ka) but clear transition from  
43 839 'deeper' to 'shallower' permafrost (~28% depth change) is matched by a minor  
44 840 increase in SRSF frequency and may explain the minor phases of enhanced SRSF  
45 841 activity identified during this period. Moreover, this gradual change in permafrost  
46 842 depth, as opposed to a stochastic response to climate warming, provides a compelling  
47 843 explanation for the significant lag between SRSF activity and the HTM.

48 844

49 845 *Phase 3: 4.8–2.6 ka ('peak phase').* – Permafrost depth is more-or-less stable and  
50 846 remains close to its minimum Holocene depth for ~2 ka. This period is matched by  
51 847 SRSF activity, as change detection analysis records a significant, sustained and  
52 848 positive rate of change ( $> 2\sigma$ ) for ~2.2 ka, with a maximum attained at ~4.3 ka and

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3 849 with SRSF frequency significantly exceeding the average frequency until  $\sim 3.3$  ka ( $>$   
4 850  $6\sigma$ ). This change is matched by the Meyer wavelet record, with a peak at  $\sim 4.6$  ka and  
5 851 a gradual decline to the Holocene background rate at  $\sim 2.5$  ka. In this phase, persistent  
6 852 shallow permafrost may directly influence SRSF occurrence by (1) actively  
7 853 destabilizing bedrock cliffs and causing slope failure and/or (2) weakening bedrock  
8 854 cliffs and making them more susceptible to other trigger factors.

9 855  
10 856 *Phase 4: 2.6–0.6 ka ('exhaustion phase').* – Permafrost depth remains relatively stable  
11 857 and shallow for  $\sim 2$  ka, with no significant deviation from modelled depths during the  
12 858 'peak phase'. However, there is a clear decrease in SRSF frequency after the mid-  
13 859 Holocene peak with a return to the Holocene background rate, as revealed by both  
14 860 change detection and Meyer wavelet analysis. In this phase, we propose that bedrock  
15 861 cliffs have reached a new equilibrium with permafrost, as the majority of slopes that  
16 862 can fail under these permafrost conditions have failed by this time; that is, the supply  
17 863 of 'potentially failable' cliffs is exhausted. As a result, SRSF occurrence returns to an  
18 864 average frequency comparable with the 'stable phase' of the early Holocene.

19 865  
20 866 *Phase 5: 0.6 - 0.1 ka ('stabilization phase').* – Contrary to the dominant Holocene  
21 867 trend, this short-term late-Holocene phase shows a clear increase in permafrost depth  
22 868 after  $\sim 0.6$  ka. This transition is coeval with a statistically significant decrease in SRSF  
23 869 frequency ( $> 2\sigma$ ) while Meyer wavelet analysis records the continued decrease in  
24 870 frequency below the Holocene background level. These data suggest that an increase  
25 871 in bedrock permafrost depth directly controls SRSF activity by stabilizing slopes and  
26 872 decreasing the susceptibility of bedrock cliffs to direct or indirect failure.

27 873  
28 874 The correlation between SRSF frequency and permafrost depth in bedrock as  
29 875 modeled by Kukkonen & Šafanda (2001) provides a compelling explanation for the  
30 876 low-frequency variations in SRSF activity during the Holocene and, in particular, for:

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32 878
- 33 879 • the significant departure from mean Holocene SRSF frequency at the end of
  - 34 880 the mid Holocene;
  - 35 881 • the lag between the HTM and the SRSF frequency peak;
  - 36 882 • the low SRSF frequency in the early Holocene; and
  - 37 883 • the marked decline in SRSF frequency near the end of the late Holocene (after
  - 38 884  $\sim 0.6$  ka).

39 885 These explanations are supported by change detection analysis and ( $d_6$ ) Meyer  
40 886 wavelet analysis. A causal link between SRSF frequency and regional permafrost  
41 887 degradation is also supported by the close match between the altitudinal distribution  
42 888 of the 92 SRSFs and the current aspect-dependent lower altitudinal limit of permafrost  
43 889 in rock faces in the Galdhøpiggen massif (Hipp et al. 2014). Approximately 87% ( $n =$   
44 890 80) of SRSFs occur within  $\pm 300$  m of the limit and  $\sim 62\%$  ( $n = 57$ ) are  $\leq 200$  m below  
45 891 this limit. A small number of SRSFs are found above the permafrost limit ( $\sim 16\%$ ;  $n =$   
46 892 15) but the majority are restricted to within  $\leq 50$  m above this limit. These data imply  
47 893 a causal relationship between SRSF occurrence and the time-dependent degradation  
48 894 and aggradation of bedrock permafrost during the Holocene, as driven by climate and  
49 895 locally controlled by aspect. Based on an altitudinal lapse rate of  $0.6$  °C per 100 m in  
50 896 mean annual air temperatures (MAAT), this implies that all SRSF sites would have  
51 897 been in the permafrost zone when temperatures were  $3.0$  °C lower than today. It is  
52 898 likely, therefore, that much of the permafrost that had survived or developed in SRSF

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3 899 cliffs following deglaciation would have degraded during the HTM when MAAT is  
4 900 likely to have reached 2.0–3.0 °C warmer than at present and when permafrost limits  
5 901 would have been correspondingly higher (Lilleøren et al. 2012).  
6 902

7 903 Higher-frequency changes in SRSF activity as reflected by weighted age-  
8 904 frequency (Fig. 6C) and (d<sub>1</sub>-d<sub>5</sub>) wavelet analysis (Supplementary Fig. 1) can be  
9 905 interpreted as represent Holocene background SRSF frequency after removal of the  
10 906 mid-Holocene positive peak and the late-Holocene negative peak of the change  
11 907 detection analysis (Fig. 8A). These higher frequency changes are more challenging to  
12 908 interpret, given the limited availability of palaeo-environmental records (e.g. seasonal  
13 909 paleo-precipitation data, storm-event chronologies, palaeoseismic and groundwater  
14 910 flux records) and the inherent SHD age uncertainties. The conceptual models related  
15 911 to deglaciation and characterized by early-Holocene peak activity (Fig. 9) can be  
16 912 discounted as these bear limited resemblance to the chronology of SRSF events.  
17 913

18 914 Changes in permafrost depth might be expected to play a role in explaining the  
19 915 higher-frequency changes. However, we cannot preclude a contribution to higher-  
20 916 frequency variability from the continuity, earthquake, and cool/wet climate conceptual  
21 917 models (Fig.9). Thawing permafrost may be a direct trigger factor for SRSF events  
22 918 due, for example, to loss of strength or elevated hydrostatic pressure, or it may render  
23 919 the rock slope susceptible to other triggers involving meltwater from spring snow melt  
24 920 or extreme rainfall events in summer (Gruber et al. 2004; Gruber & Haeberli 2007;  
25 921 Krautblatter et al. 2013; Blikra & Christiansen 2014; Draebing et al. 2014;  
26 922 Krautblatter & Leith 2015; Messenzehl & Dikau 2017). Extreme summer rainfall  
27 923 events, which are likely to have been more frequent during warm periods, have been  
28 924 implicated in triggering debris-flow events in Leirdalen (Matthews et al. 2009) and  
29 925 might have triggered some SRSFs.  
30 926

### 31 927 *Further conceptual and methodological implications* 32 928

33 929 Thus, the timing of SRSFs in this study, with fluctuating SRSF activity rising to a  
34 930 sustained peak at the transition from the mid- to late-Holocene, suggests the  
35 931 importance of progressive but intermittent permafrost degradation lagging behind the  
36 932 highest temperatures of the Holocene. Subsequent declining SRSF frequencies, in  
37 933 contrast, appear to signal exhaustion of the supply of failable cliffs and/or renewed  
38 934 aggradation of permafrost.  
39 935

40 936 These fundamental findings recognize that Holocene SRSF activity in  
41 937 Jotunheimen essentially reflects paraperiglacial processes: that is, it is a conditional  
42 938 response to the transition from a permafrost to a seasonal-freezing climatic regime as  
43 939 permafrost depth decreases (cf. Mercier 2008; Scarpozza 2016; Matthews et al. 2017).  
44 940 While this model is primarily applicable to the SRSFs sampled in this study, it could  
45 941 be tested in comparable mountain regions. In particular, links between permafrost  
46 942 degradation and enhanced slope failure may explain SRSF frequency in regions with  
47 943 comparable seismotectonics, glaciation and deglaciation histories or climatic trends.  
48 944 Robust SRSF chronologies would need to be constructed to test the model, either  
49 945 using radiometric methods (e.g. <sup>10</sup>Be) or calibrated-age dating techniques (e.g. SHD).  
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51 947 Our new SRSF chronology indicates, moreover, that SHD can be used to  
52 948 generate reliable SRSF chronologies, although further work is necessary to verify this  
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3 949 technique by directly comparing age estimates for individual landforms derived from  
4 950 both SHD and radiometric methods.

5 951  
6 952 Finally, the recognition of a causal link between climate, permafrost  
7 953 degradation and enhanced slope instability may have important implications for  
8 954 glacial and periglacial environments under global warming scenarios. In particular,  
9 955 while widespread retreat of mountain ice caps and valley glaciers may trigger initial  
10 956 slope instability, our data suggest that the geomorphological impact of current  
11 957 climatic and deglacial trends and, in particular, the slow transition from glacial to  
12 958 periglacial, and to seasonal-freezing climatic regimes, may have a long-lasting impact  
13 959 on mountain environments.

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15 961

16 962 Conclusions

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18 964 (1) We have developed an approach to the exposure-age dating of a large sample of  
19 965 rock-slope failures, which involves adapting Schmidt-hammer exposure-age dating  
20 966 (SHD) as a calibrated-age dating technique to the specific characteristics of small  
21 967 rock-slope failures (SRSFs). SHD has provided an effective and low-cost method for  
22 968 constructing a regional Holocene chronology of SRSFs (12 to 2520 m<sup>3</sup>) in the alpine  
23 969 zone of Jotunheimen.

24 970

25 971 (2) Focusing on a large sample of SRSFs enables the detection of temporal variations  
26 972 in the frequency and magnitude of events through the Holocene. Modes in a weighted  
27 973 age-frequency distribution at ~8.9, 7.3, 5.9 and 4.5 ka were substantiated by  
28 974 probability density function analysis, which produced individual Gaussian age  
29 975 distributions of  $9.00 \pm 1.13$  ka,  $7.38 \pm 0.99$  ka,  $6.40 \pm 0.77$  ka and  $4.50 \pm 1.42$  ka.  
30 976 Based on this analysis, SRSF activity was relatively low following deglaciation in the  
31 977 early Holocene and attained a maximum towards the end of the mid Holocene (~4.5  
32 978 ka). Peak SRSF activity lagged behind the Holocene Thermal Maximum by at least  
33 979 ~2.2 ka and declined thereafter with a very low frequency of events during the last  
34 980 millennium.

35 981

36 982 (3) Using change detection and discreet Meyer wavelet analysis in combination with  
37 983 proxy temperature indicators and an existing permafrost depth model, we propose that  
38 984 enhanced SRSF activity was primarily controlled by permafrost degradation. As a  
39 985 result, the Holocene permafrost depth record is subdivided into five distinct periods  
40 986 and related to the SRSF chronology as follows:

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- 43 989 • 10 - 8.1 ka – ‘stable phase’ – low SRSF activity; maximum Holocene  
44 990 permafrost depth.
  - 45 991 • 8.1 - 4.8 ka – ‘transition phase’ – increasing susceptibility to SRSF activity;  
46 992 decreasing permafrost depth.
  - 47 993 • 4.8 - 2.6 ka – ‘peak phase’ – maximum SRSF activity; minimum Holocene  
48 994 permafrost depth.
  - 49 995 • 2.6 - 0.6 ka – ‘exhaustion phase’ – decreasing SRSF activity; little change in  
50 996 shallow permafrost depth.
  - 51 997 • 0.6 - 0.1 ka – ‘stabilization phase’ – minimum SRSF activity; increasing  
52 998 permafrost depth.
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3 999 (4) Long-term relative change in permafrost depth provides a compelling explanation  
4 1000 for the high-magnitude departures from the SRSF background rate. In particular, the  
5 1001 gradual change in permafrost depth during the ‘transition phase’, as opposed to a  
6 1002 stochastic response to climate warming, accounts for the significant lag (~2.2 ka)  
7 1003 between the Holocene Thermal Maximum and the SRSF frequency peak. Moreover,  
8 1004 persistent shallow permafrost during the ‘peak phase’ may be the key driver behind  
9 1005 SRSF occurrence by (a) actively destabilizing bedrock cliffs and causing slope failure  
10 1006 and/or (b) weakening bedrock cliffs and making them more susceptible to other  
11 1007 trigger factors.

12 1008  
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14 1009 (5) Conversely, declining SRSF frequency during the ‘exhaustion phase’ appears to  
15 1010 reflect the diminished supply of potentially failable cliffs, even under a shallow  
16 1011 permafrost depth scenario. Finally, low frequency of SRSF occurrence during the  
17 1012 ‘stabilization phase’ likely reflects an increase in permafrost depth (permafrost  
18 1013 aggradation) after ~0.6 ka; a change which would have been sufficient to stabilize  
19 1014 slopes and decrease the susceptibility of bedrock cliffs to direct or indirect failure.  
20 1015

21 1016 (6) This interpretation is supported by geomorphological evidence, given the  
22 1017 consistent location of SRSF sites in relation to the local aspect-dependent lower  
23 1018 altitudinal limit of permafrost in cliff faces. This new paraperiglacial model attributes  
24 1019 enhanced SRSF activity to progressive and intermittent permafrost degradation during  
25 1020 Holocene warm periods, including the possibility of renewed aggradation of  
26 1021 permafrost during short-term cold periods and renewed degradation during the  
27 1022 ensuing warm periods  
28 1023

29 1024 (7) Our new thermally-driven, permafrost-degradation model of SRSF events in  
30 1025 Jotunheimen bears little similarity to existing models of Holocene RSF activity.  
31 1026 However, while aspects of this new model require further testing by other methods  
32 1027 and in other regions, the results of this study have important implications for climate-  
33 1028 change forcing of RSF activity. Projected mean annual global warming is predicted to  
34 1029 decrease the area of mountain permafrost and raise lower altitudinal permafrost limits.  
35 1030 This in turn will likely destabilize higher bedrock slopes and increase SRSF frequency  
36 1031 there. The delayed response of peak SRSF frequency to warming climate, as  
37 1032 modulated by permafrost depth, may therefore result in a long-lasting impact of  
38 1033 current climate trends on mountain environments.  
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41 1036  
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49 1043 No. 205 (see <http://jotunheimenresearch.wixsite.com/home>).  
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## FIGURE CAPTIONS

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5 1747 *Fig. 1.* Location map: numbers and open circles identify the studied SRSFs; sites of  
6 1748 control points are shown by crosses.  
7 1749
- 8 1750 *Fig. 2.* Photographs of selected small rock-slope failures (SRSFs): (A) No. 23,  
9 1751 Gravdalen; (B) Nos 7 and 8, Leirdalen; (C) Nos 34-36, Bjørndalen; (D) No. 7,  
10 1752 Sognefjell; (E) and (F) No. 22, Gravdalen (also the site of a young control point).  
11 1753
- 12 1754 *Fig. 3.* Schematic of the fan-scar-cliff comparison tests with expected differences in  
13 1755 mean R-values between fan boulders, scar bedrock surfaces, unfailed cliffs, and rock  
14 1756 surfaces used as younger and older control-point surfaces. Expectations apply to  
15 1757 single-event SRSF events without the possible complications discussed in the text.  
16 1758
- 17 1759 *Fig. 4.* Frequency distributions of four SRSF characteristics: (A) fan volume; (B)  
18 1760 altitude; (C) aspect; (D) mean R-value. Eight sites in gabbroic gneiss (Sognefjell) are  
19 1761 differentiated by solid black shading from 84 sites in pyroxene-granulite gneiss.  
20 1762
- 21 1763 *Fig. 5.* Calibration curves and calibration equations for (A) pyroxene-granulite gneiss  
22 1764 and (B) gabbroic gneiss. Note that both calibration curves are based on two control  
23 1765 points of known age (25 years and 9700 years) using data presented in Table 3.  
24 1766
- 25 1767 *Fig. 6.* Holocene SHD chronologies of SRSF activity for Jotunheimen: (A) individual  
26 1768 SHD dates with their 95% confidence intervals in the different subregions; (B) age-  
27 1769 frequency distributions of SRSF events at the regional level using 2000-yr, 1000-yr,  
28 1770 500-yr and 200-yr time intervals; (C) weighted age-frequency distribution with age-  
29 1771 frequency curve defined by binomial smoothing; (D) variation in the magnitude of  
30 1772 SRSF events based on rock volume using 200-year time intervals. Vertical bands  
31 1773 (numbered) are the 4 modes in the weighted age-frequency distribution suggesting  
32 1774 phases of enhanced regional SRSF activity.  
33 1775
- 34 1776 *Fig. 7.* Probability density function analysis of SRSF activity for Jotunheimen: (A)  
35 1777 histogram and KS density PDE; (B) individual Gaussian age distributions ( $n = 5$ ), the  
36 1778 sum of which integrates to the cumulative PDE with a model fit that is graphically  
37 1779 indistinguishable from from the PDE model. The number of ages listed for each  
38 1780 Gaussian age distribution (#) exceeds the total number of SRSF events identified in  
39 1781 Jotunheimen as some ages contribute to  $>1$  Gaussian distribution; (C) peak Gaussian  
40 1782 numerical ages and  $1\sigma$  uncertainties for the five individual Gaussian age distributions  
41 1783 plotted against the peak probability density (PPD). The PPD scales with the number  
42 1784 and spatial clustering of individual ages. Reported RSF volumes are based on the sum  
43 1785 of individual SRSF volumes ( $m^3$ ) which comprise each Gaussian age distribution; (D)  
44 1786 distribution of SRSF ages, sorted by oldest to youngest. The 42 SRSF events which  
45 1787 account for the dominant mode at  $4.50 \pm 1.42$  ka (within  $1\sigma$ ) are highlighted.  
46 1788
- 47 1789 *Fig. 8.* Change detection and related analyses: (A) cumulative sum change detection  
48 1790 graph showing positive (blue) and negative (orange) changes and statistically  
49 1791 significant departures ( $> 2\sigma$ ) from the background SRSF frequency; (B) modelled  
50 1792 permafrost depth in Fennoscandia (5% porosity) from Kukkonen & Šafanda (2001),  
51 1793 subdivided into five distinct phases; (C) results of discreet Meyer wavelet analysis,  
52 1794 showing the lowest frequency decomposed signal ( $d_6$ ).  
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4 1796 *Fig. 9.* Models for different patterns and causes of Holocene variations in RSF  
5 1797 frequency and/or magnitude: (1) continuity-of-activity; (2) intermittent-earthquakes;  
6 1798 (3) deglaciation-close-tracking; (4) deglaciation-lagging; (5) cool/wet-climate-  
7 1799 response; and (6) the new thermally-driven permafrost-degradation model proposed in  
8 1800 this study for SRSFs in Jotunheimen. The subdivisions of the Holocene shown are  
9 1801 those proposed by Walker et al. (2012).

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11 1803 *Fig. 10.* Relationships between SRSF frequency in Jotunheimen and proxy climatic  
12 1804 records: (A) temporal variations in SRSF frequency from Fig. 6C; (B) pollen-based  
13 1805 reconstruction of annual air temperature for Northern Europe expressed as deviations  
14 1806 from the mean (Seppä et al., 2009); (C) mean summer air temperature deviations from  
15 1807 present in the Scandes Mountains based on pine tree-limit variations (Dahl and Nesje,  
16 1808 1996); (D) pollen-based July air temperature variations at Øvre Heimdalsvatnet,  
17 1809 eastern Jotunheimen (Velle et al., 2010); (E) periods of above average air temperature  
18 1810 (shaded) based on the GISP 2 Greenland ice core  $\delta^{18}\text{O}$  record (Alley, 2004; Wanner et  
19 1811 al., 2011); (F) periods of above average sea-surface temperatures in the North Atlantic  
20 1812 Ocean (shaded) based on standardized stacked ice-rafted debris (IRD) records (Bond  
21 1813 et al., 2001; Wanner, et al., 2011); (G) periods when glaciers in the Smørstabbtindan  
22 1814 massif, Jotunheimen, were smaller than today (shaded) based on glaciolacustrine and  
23 1815 glaciofluvial stratigraphy (Matthews and Dresser, 2008). Vertical bands indicate  
24 1816 phases of enhanced regional SRSF frequency (as in Fig. 6).  
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### SUPPLEMENTARY MATERIAL

(Caption to be included with figure)

34 1824  
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36 1826 Supplementary Fig. 1: Full results of discreet Meyer wavelet analysis, showing all six  
37 1827 decomposed signals (green), ranging from high ( $d_1$ ) to low frequency ( $d_6$ ), of which  
38 1828 the latter represents the only single event structure of Holocene SRSF activity. The  
39 1829 blue curves ( $a_1 - a_5$ ) represent the cumulative aggregation of the decomposed signals  
40 1830 ( $d_1 - d_6$ ) where  $a_6$  represents the mean background rate of SRSF occurrence ( $0.92 \pm$   
41 1831  $0.20$ ), which is identical to the Holocene mathematical mean. The sum of all  
42 1832 decomposed signals results in a model ( $S_m$ ) that is identical to the 100 yr bin  
43 1833 histogram data ( $S_d$ ).  
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3 1845 *Table 1.* Data on the 92 small rock-slope failures (SRSFs) located in Jotunheimen:  
4 1846 Leirdalen (Nos 1-29); Bjørndalen (30-40); Gravidalen (41-68); Høgvaglura (69-72);  
5 1847 Visdalen (73-80); Veodalen (81-84); Sognefjell (85-92).  
6 1848 Abbreviations:  $L$  = fan length;  $W$  = fan width;  $V$  = fan volume; SD = standard  
7 1849 deviation of R-values;  $C_s$  = error associated with the dated surface;  $C_c$  = error  
8 1850 associated with the calibration equation; CI = confidence interval for the SHD age  
9 1851 based on the total error ( $C_t$ ).  
10 1852

No.	$L$ (m)	$W$ (m)	$V$ (m <sup>3</sup> )	Altitude (m a.s.l.)	Aspect	Mean R-value	SD	$C_s$ (yr)	$C_c$ (yr)	SHD age (yr $\pm$ 95% CI)	Sub-region
1	70	25	1050	1420	West	45.0	9.90	1047	513	7018 $\pm$ 1124	Leirdalen
2	80	20	960	1440	West	44.51	8.80	930	414	7277 $\pm$ 1018	
3	15	9	81	1400	West	39.69	9.47	1001	458	9833 $\pm$ 1101	
4	90	40	2160	1160	West	41.53	9.57	1012	445	8857 $\pm$ 1105	
5	15	8	72	1030	West	43.26	10.03	1060	426	7940 $\pm$ 1143	
6	90	20	1080	1160	West	43.62	10.23	1081	423	7749 $\pm$ 1161	
7	8	25	120	1140	West	44.69	9.41	995	412	7182 $\pm$ 1077	
8	30	25	450	1140	West	46.59	10.35	1094	392	6175 $\pm$ 1162	
9	8	8	38	1135	West	47.28	8.63	912	385	5809 $\pm$ 990	
10	15	25	225	1135	West	44.68	8.85	936	412	7187 $\pm$ 1022	
11	30	20	360	1200	North	52.38	10.07	1064	333	3105 $\pm$ 1115	
12	50	25	750	1425	North	46.49	8.63	912	394	6228 $\pm$ 994	
13	15	25	225	960	East	51.50	8.34	882	342	3572 $\pm$ 946	
14	50	25	750	955	East	49.79	10.74	1135	360	4478 $\pm$ 1191	
15	70	60	2520	950	East	49.28	8.73	923	365	4749 $\pm$ 992	
16	50	25	750	1290	West	48.29	9.98	1055	375	5273 $\pm$ 1120	
17	20	40	480	1320	West	54.10	9.90	1047	315	2193 $\pm$ 1093	
18	20	15	180	1320	West	57.53	10.15	1073	280	375 $\pm$ 1109	
19	30	40	720	1320	West	55.95	8.61	910	296	1213 $\pm$ 957	
20	18	14	151	1120	East	48.79	8.43	891	370	5008 $\pm$ 965	
21	16	8	77	1120	East	44.40	8.29	876	415	7336 $\pm$ 970	
22	25	14	210	1130	East	48.93	9.11	963	368	4934 $\pm$ 1031	
23	40	13	312	1170	East	41.30	9.14	966	447	8979 $\pm$ 1065	
24	25	15	225	1180	East	40.82	9.16	968	296	9233 $\pm$ 1012	
25	15	13	117	1180	East	43.37	9.49	1003	426	7882 $\pm$ 1090	
26	20	4	48	1190	East	44.86	8.70	920	410	7092 $\pm$ 1007	
27	12	8	58	1240	East	49.28	10.53	1113	365	4749 $\pm$ 1171	
28	20	4	48	1240	East	45.92	10.98	1160	399	6530 $\pm$ 1227	
29	22	4	53	1200	East	47.15	8.24	871	387	5878 $\pm$ 953	
30	90	16	864	1370	East	44.27	10.65	1126	416	7405 $\pm$ 1200	Bjørndalen
31	30	15	270	1380	East	44.62	10.10	1068	413	7219 $\pm$ 1145	
32	30	10	180	1380	East	52.60	8.62	911	331	2989 $\pm$ 922	
33	75	30	1350	1360	East	54.91	8.30	877	307	1764 $\pm$ 930	
34	30	15	270	1380	East	49.87	7.53	796	359	4436 $\pm$ 873	
35	30	12	216	1380	East	49.46	7.84	829	363	4653 $\pm$ 905	
36	20	30	360	1380	East	50.19	8.61	910	355	4266 $\pm$ 977	
37	80	35	1680	1330	South	50.23	9.57	960	355	4245 $\pm$ 1024	
38	25	15	225	1300	North	54.07	6.73	711	316	2209 $\pm$ 778	
39	50	30	900	1305	North	55.37	7.95	840	302	1520 $\pm$ 893	
40	25	25	375	1300	North	53.30	8.20	867	323	2617 $\pm$ 925	Gravidalen
41	55	20	660	1480	West	49.43	8.11	857	363	4669 $\pm$ 931	
42	15	35	315	1480	West	55.49	6.69	707	301	1456 $\pm$ 769	
43	65	15	585	1480	West	51.11	8.40	888	346	3778 $\pm$ 953	
44	60	15	540	1470	West	50.84	7.05	745	349	3922 $\pm$ 823	
45	65	25	975	1470	West	50.01	8.85	936	357	4362 $\pm$ 1001	
46	30	15	270	1460	West	52.57	7.97	843	331	3004 $\pm$ 905	
47	75	20	900	1460	West	53.03	6.27	663	326	2761 $\pm$ 739	
48	25	30	450	1430	South	50.01	7.00	740	357	4362 $\pm$ 822	
49	17	8	82	1440	South	49.10	8.45	893	367	4844 $\pm$ 964	
50	40	15	360	1440	South	49.71	7.72	816	360	4521 $\pm$ 892	
51	15	10	90	1440	South	50.38	7.78	822	356	4165 $\pm$ 896	
52	15	6	54	1400	South	56.21	7.38	780	293	1075 $\pm$ 834	
53	10	5	30	1400	South	57.99	6.22	658	275	131 $\pm$ 713	
54	7	8	34	1360	South	47.32	8.00	846	385	5788 $\pm$ 929	
55	10	6	36	1280	South	40.31	10.14	1072	457	9504 $\pm$ 1165	
56	12	5	36	1440	South	48.82	8.12	858	370	4992 $\pm$ 935	
57	6	5	18	1440	South	47.43	7.72	816	384	5729 $\pm$ 902	
58	8	8	38	1440	South	51.63	7.70	814	341	3503 $\pm$ 882	
59	4	5	12	1440	South	51.12	6.62	700	346	3773 $\pm$ 781	
60	7	4	17	1480	South	48.02	7.43	785	378	5416 $\pm$ 872	



61	20	5	60	1480	South	52.10	11.98	1266	336	3254 ± 1310	
62	14	8	67	1480	South	46.17	9.02	953	399	6397 ± 1033	
63	6	12	43	1430	South	48.74	8.09	855	370	5035 ± 932	
64	10	5	30	1430	South	46.99	7.65	809	388	5963 ± 897	
65	14	3	25	1460	South	49.91	8.38	886	358	4415 ± 956	
66	15	4	36	1520	South	51.92	8.34	882	338	3349 ± 944	
67	6	4	14	1540	South	49.95	9.74	1030	358	4393 ± 1090	
68	10	5	30	1540	South	49.37	7.08	748	364	4701 ± 832	
69	20	15	180	1550	East	50.13	7.74	818	356	4298 ± 892	Hogvaglura
70	50	12	360	1550	East	45.16	10.05	1062	407	6933 ± 1138	
71	20	10	120	1540	East	46.35	8.94	945	395	6302 ± 1024	
72	20	10	120	1540	East	42.10	11.92	1260	439	8555 ± 1334	
73	15	4	36	1420	East	47.03	11.08	1171	388	5941 ± 1234	Visdalen
74	15	9	81	1420	East	50.70	10.47	1107	350	3996 ± 1161	
75	10	4	24	1420	East	54.42	9.47	1001	312	2024 ± 1049	
76	25	10	150	1260	West	49.96	10.20	1078	358	4388 ± 1136	
77	40	20	480	1200	East	51.37	10.30	1089	343	3641 ± 1142	
78	70	30	1260	1200	East	52.98	8.86	937	327	2787 ± 992	
79	35	20	420	1200	East	51.57	7.93	838	341	3535 ± 905	
80	60	8	288	1190	East	50.31	10.75	1136	354	4202 ± 1190	
81	55	40	1320	1350	South	53.33	8.72	922	323	2601 ± 977	Veodalen
82	45	12	324	1340	South	54.33	9.34	987	313	2071 ± 1036	
83	50	25	750	1330	South	51.56	10.15	1073	341	3540 ± 1126	
84	45	40	1080	1330	South	49.46	10.60	1121	363	4653 ± 1178	
85	6	10	36	1375	East	37.17	11.29	900	239	9412 ± 931	Sognefjell
86	7	5	21	1425	South	38.53	8.82	703	244	8868 ± 744	
87	6	6	22	1425	South	39.42	10.43	831	247	8513 ± 868	
88	10	5	30	1430	South	41.38	9.86	786	255	7729 ± 826	
89	6	5	18	1430	South	40.73	9.47	755	253	7989 ± 796	
90	16	10	96	1450	South	38.26	8.83	704	243	8976 ± 745	
91	9	5	27	1435	West	42.33	10.03	800	259	7349 ± 840	
92	10	7	42	1370	East	36.95	9.14	729	238	9500 ± 766	

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*Table 2.* Radiocarbon age control for deglaciation in the study area.

Location	Altitude (m. a.s.l.)	<sup>14</sup> C age ± 1σ (years BP)	Calibrated age* (cal. years BP)	Reference
<i>Leirdalen/Bjørndalen</i>				
Lower Leirdalen	920	9089 ± 61	10426 – 10170 (94.8%)	Barnett et al. 2000
Bøverkinnhalsen	1020	8570 ± 60	9677 – 9475 (95.4%)	Nesje & Dahl 2001
Bjørndalen	1250	8760 ± 100	10066 – 9547 (77.1%)	Matthews et al. 2005
<i>Sognefjell</i>				
Nedre Hervavatnet	1287	8695 ± 75	9921 – 9530 (94.6%)	Hormes et al. 2009
Gjuvvatnet	1248	8885 ± 140	10247 – 9557 (95.4%)	Karlén & Matthews 1992

\* Most probable range with probability in brackets

*Table 3.* Control point data: values used for calibration equations are indicated in bold. *Abbreviations:* Gneiss = pyroxene-granulite gneiss; Gabbro = gabbroic gneiss; Combined = data combined from two replicate sites; SD = standard deviation; CI = confidence interval; n = sample size.

Control point	Geology	Type	Age (years)	Mean R-value	SD	95% CI	n
Gravdalen	Gneiss	SRSF	25	58.22	6.29	0.72	300
Gravdalen	Gneiss	Road cut	25	58.15	6.56	0.75	300
<b>Gravdalen</b>	<b>Gneiss</b>	<b>Combined</b>	<b>25</b>	<b>58.19</b>	<b>6.42</b>	<b>0.52</b>	<b>600</b>
<b>Sognefjell</b>	<b>Gabbro</b>	<b>Road cut</b>	<b>25</b>	<b>60.65</b>	<b>7.26</b>	<b>0.83</b>	<b>300</b>
Gravdalen	Gneiss	Bedrock	9700	39.71	4.80	1.25	60
Leirdalen	Gneiss	Bedrock	9700	40.19	4.69	1.22	60
<b>SE Smørstabbtindan</b>	<b>Gneiss</b>	<b>Combined</b>	<b>9700</b>	<b>39.94</b>	<b>4.79</b>	<b>0.87</b>	<b>120</b>
Leirbreen	Gabbro	Bedrock	9700	35.78	2.84	0.74	60
Bøverbreen	Gabbro	Bedrock	9700	37.12	3.53	0.92	60
<b>W Smørstabbtindan</b>	<b>Gabbro</b>	<b>Combined</b>	<b>9700</b>	<b>36.45</b>	<b>3.25</b>	<b>0.59</b>	<b>120</b>

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*Table 4.* Comparative R-values from fans, scars and unfailed cliffs associated with selected SRSFs. Further information on these six SRSFs are provided in Table 1.

No.	Fan			Scar			Unfailed cliff		
	Mean	SD	95% CI	Mean	SD	95% CI	Mean	SD	95% CI
5	43.26	10.03	2.00	41.34	7.75	1.55	42.20	7.86	1.57
46	51.63	7.70	1.54	51.32#	8.10	1.62	41.63†	9.20	1.83
47	51.12	6.62	1.32	54.05#	8.05	1.69	43.26†	10.19	2.03
51	37.17*	11.29	2.25	42.89	9.73	1.94	38.54	10.37	2.07
58	36.95*	9.14	1.82	43.99	10.44	2.08	40.68	12.30	2.45
81	49.96*	10.20	2.03	54.47#	8.07	1.60	43.38†	10.78	2.15

\* Fan significantly different from scar (p<0.05)

# Scar significantly different from unfailed cliff (p<0.05)

† Unfailed cliff significantly different from fan (p<0.05)

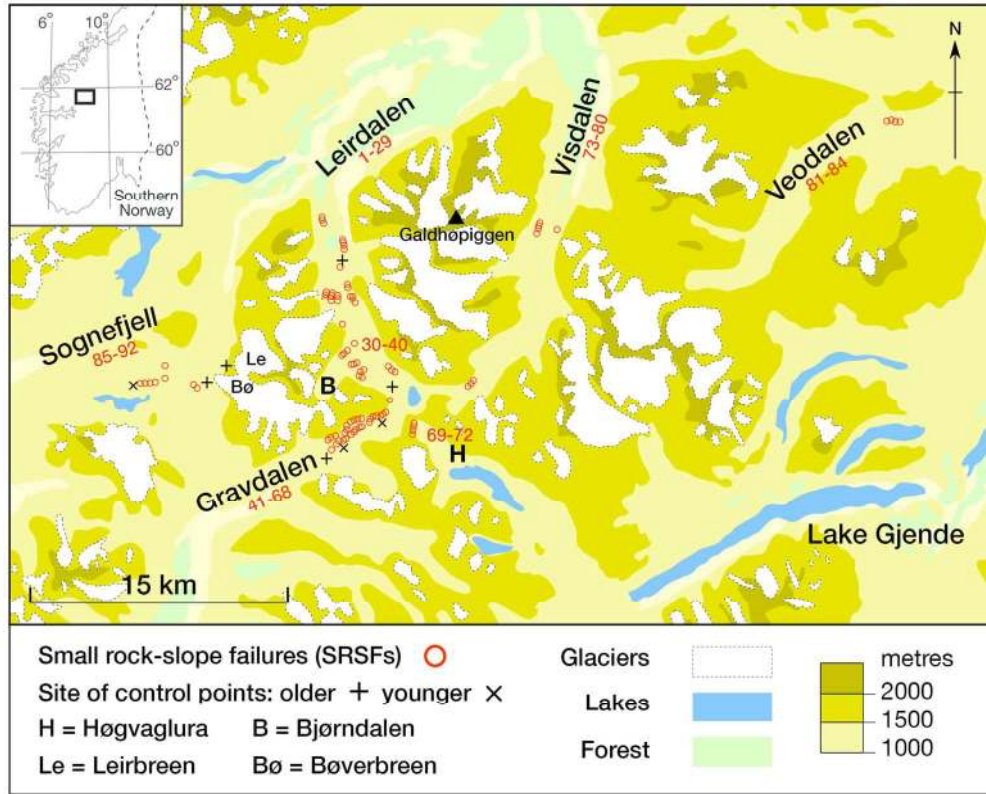


Fig. 1. Location map: numbers and open circles identify the studied SRSFs; sites of control points are shown by crosses.

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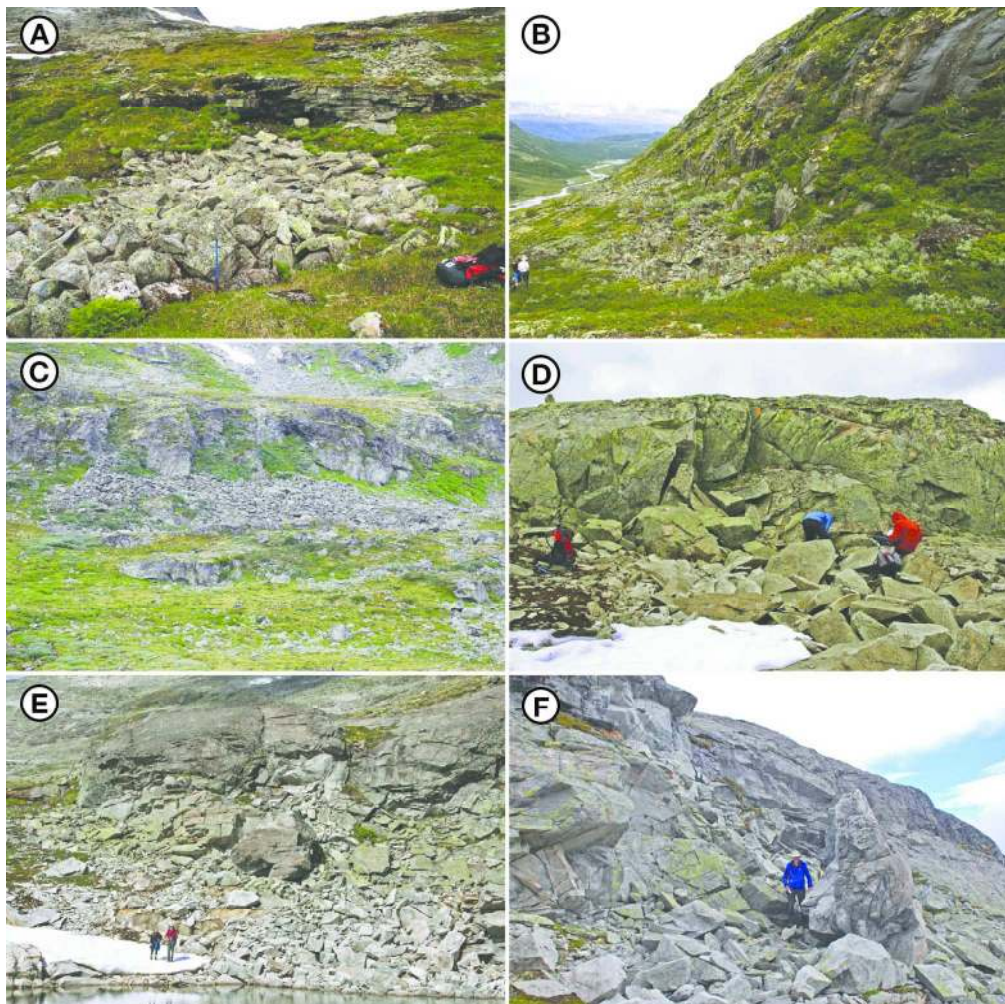
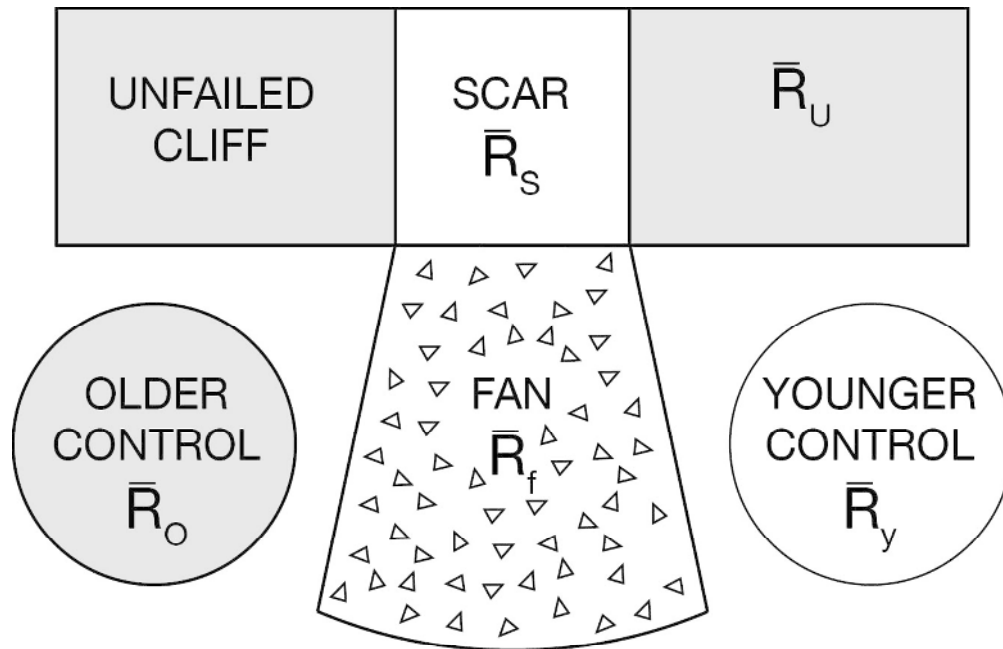


Fig. 2. Photographs of selected small rock-slope failures (SRSFs): (A) No. 23, Gravdalen; (B) Nos 7 and 8, Leirdalen; (C) Nos 34-36, Bjørndalen; (D) No. 7, Sognefjell; (E) and (F) No. 22, Gravdalen (also the site of a young control point).

174x173mm (300 x 300 DPI)



Expected mean R-values:

$$\bar{R}_y \geq \bar{R}_s = \bar{R}_f \geq \bar{R}_u \geq \bar{R}_o$$

Fig. 3. Schematic of the fan-scar-cliff comparison tests with expected differences in mean R-values between fan boulders, scar bedrock surfaces, unfailed cliffs, and rock surfaces used as younger and older control-point surfaces. Expectations apply to single-event SRSF events without the possible complications discussed in the text.

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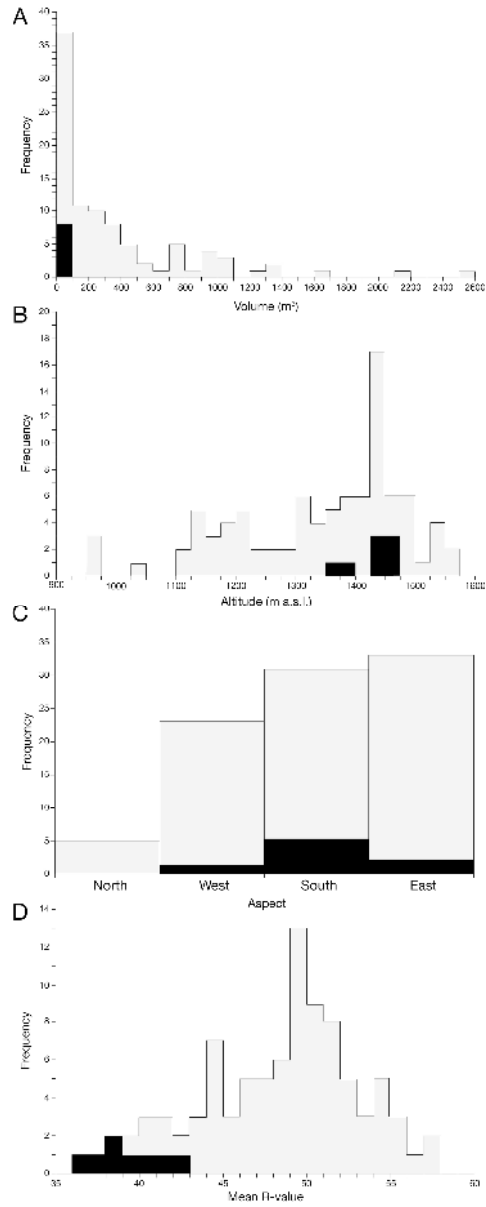


Fig. 4. Frequency distributions of four SRSF characteristics: (A) fan volume; (B) altitude; (C) aspect; (D) mean R-value. Eight sites in gabbroic gneiss (Sognefjell) are differentiated by solid black shading from 84 sites in pyroxene-granulite gneiss.

276x701mm (300 x 300 DPI)

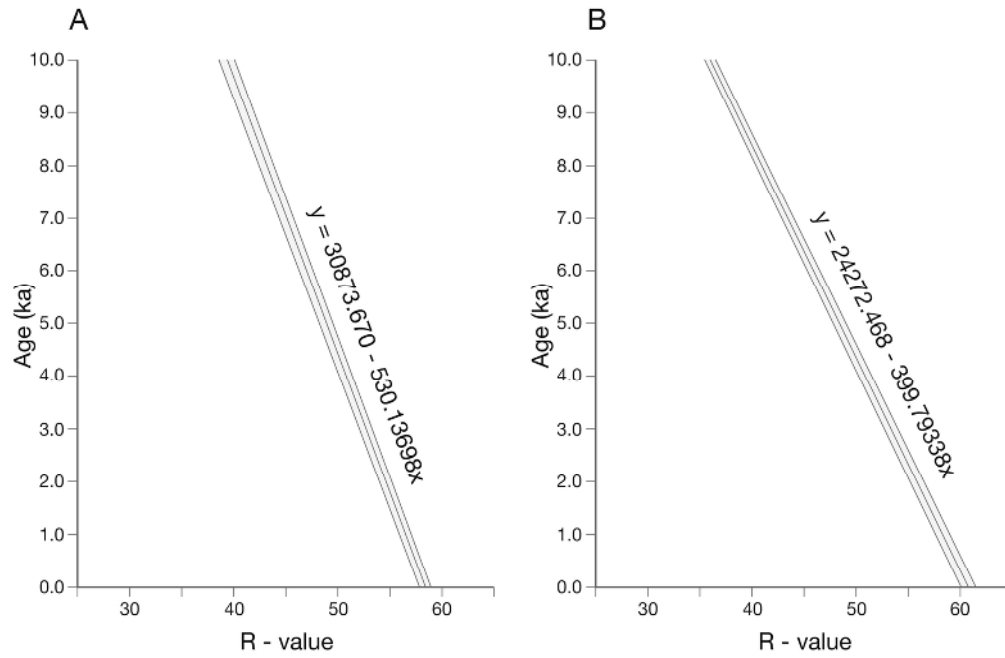


Fig. 5. Calibration curves and calibration equations for (A) pyroxene-granulite gneiss and (B) gabbroic gneiss. Note that both calibration curves are based on two control points of known age (25 years and 9700 years) using data presented in Table 3.

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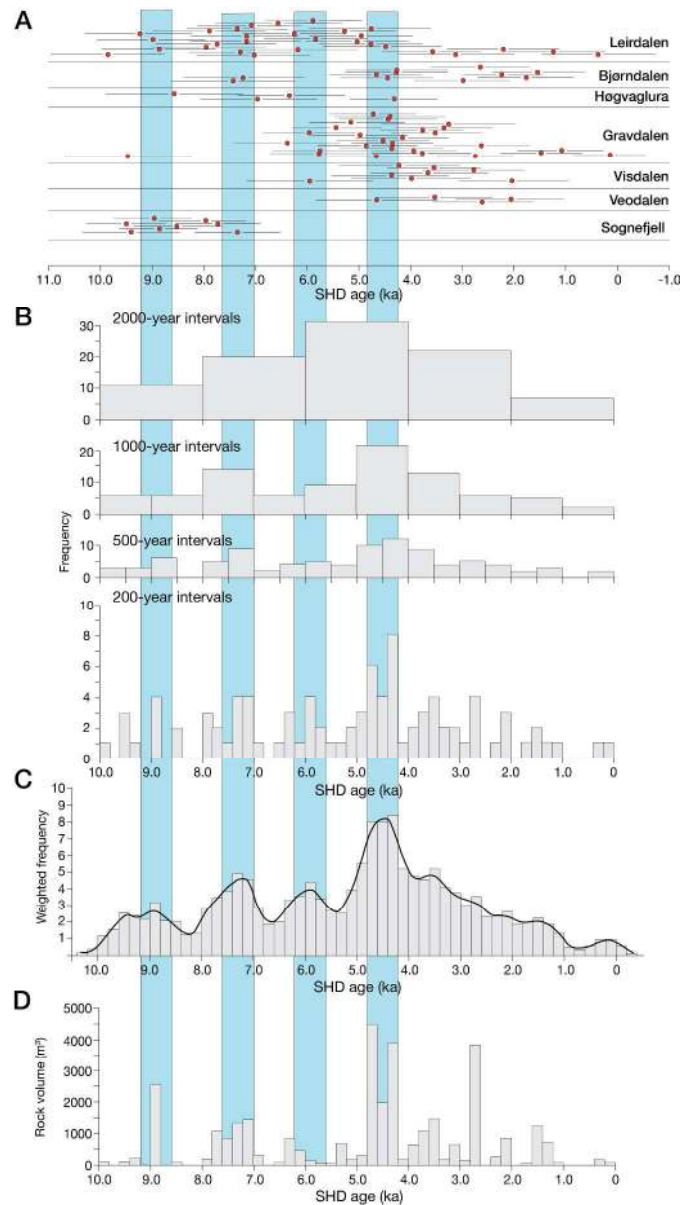


Fig. 6. Holocene SHD chronologies of SRSF activity for Jotunheimen: (A) individual SHD dates with their 95% confidence intervals in the different subregions; (B) age-frequency distributions of SRSF events at the regional level using 2000-yr, 1000-yr, 500-yr and 200-yr time intervals; (C) weighted age-frequency distribution with age-frequency curve defined by binomial smoothing; (D) variation in the magnitude of SRSF events based on rock volume using 200-year time intervals. Vertical bands (numbered) are the 4 modes in the weighted age-frequency distribution suggesting phases of enhanced regional SRSF activity.

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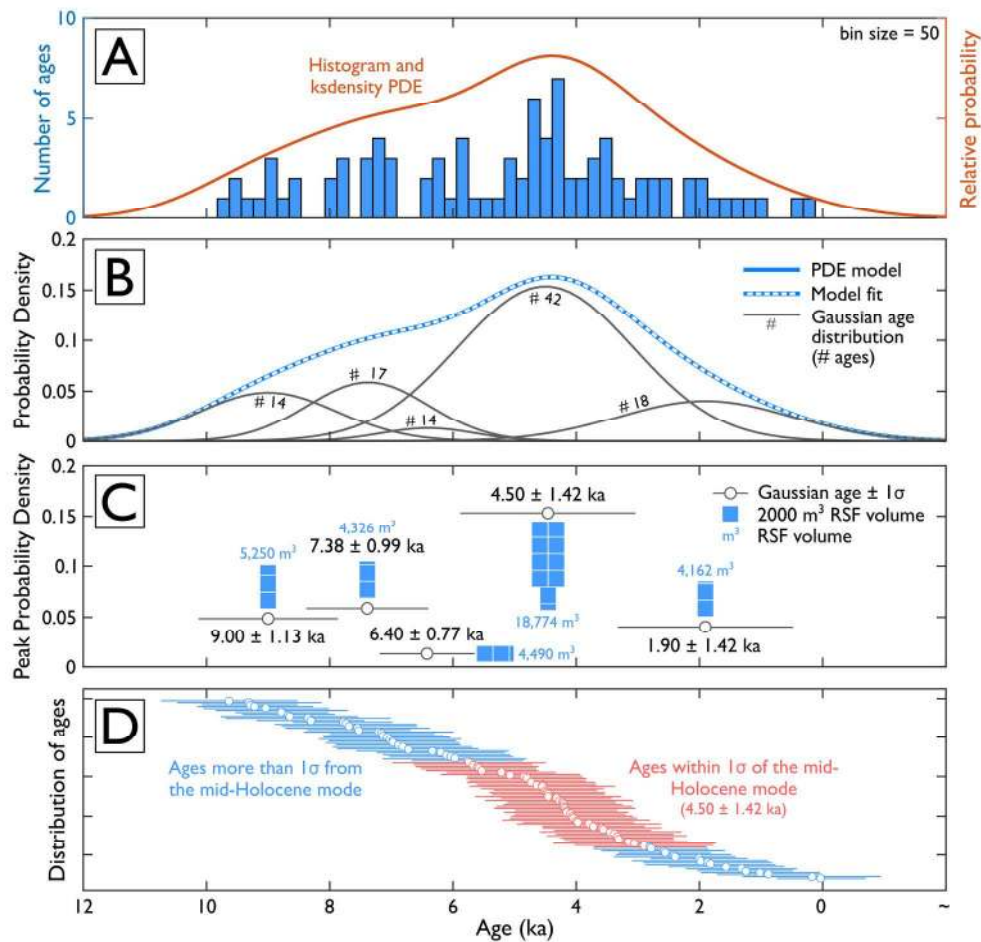


Fig. 7. Probability density function analysis of SRSF activity for Jotunheimen: (A) histogram and KS density PDE; (B) individual Gaussian age distributions ( $n = 5$ ), the sum of which integrates to the cumulative PDE with a model fit that is graphically indistinguishable from the PDE model. The number of ages listed for each Gaussian age distribution (#) exceeds the total number of SRSF events identified in Jotunheimen as some ages contribute to  $>1$  Gaussian distribution; (C) peak Gaussian numerical ages and  $1\sigma$  uncertainties for the five individual Gaussian age distributions plotted against the peak probability density (PPD). The PPD scales with the number and spatial clustering of individual ages. Reported RSF volumes are based on the sum of individual SRSF volumes ( $m^3$ ) which comprise each Gaussian age distribution; (D) distribution of SRSF ages, sorted by oldest to youngest. The 42 SRSF events which account for the dominant mode at  $4.50 \pm 1.42$  ka (within  $1\sigma$ ) are highlighted.

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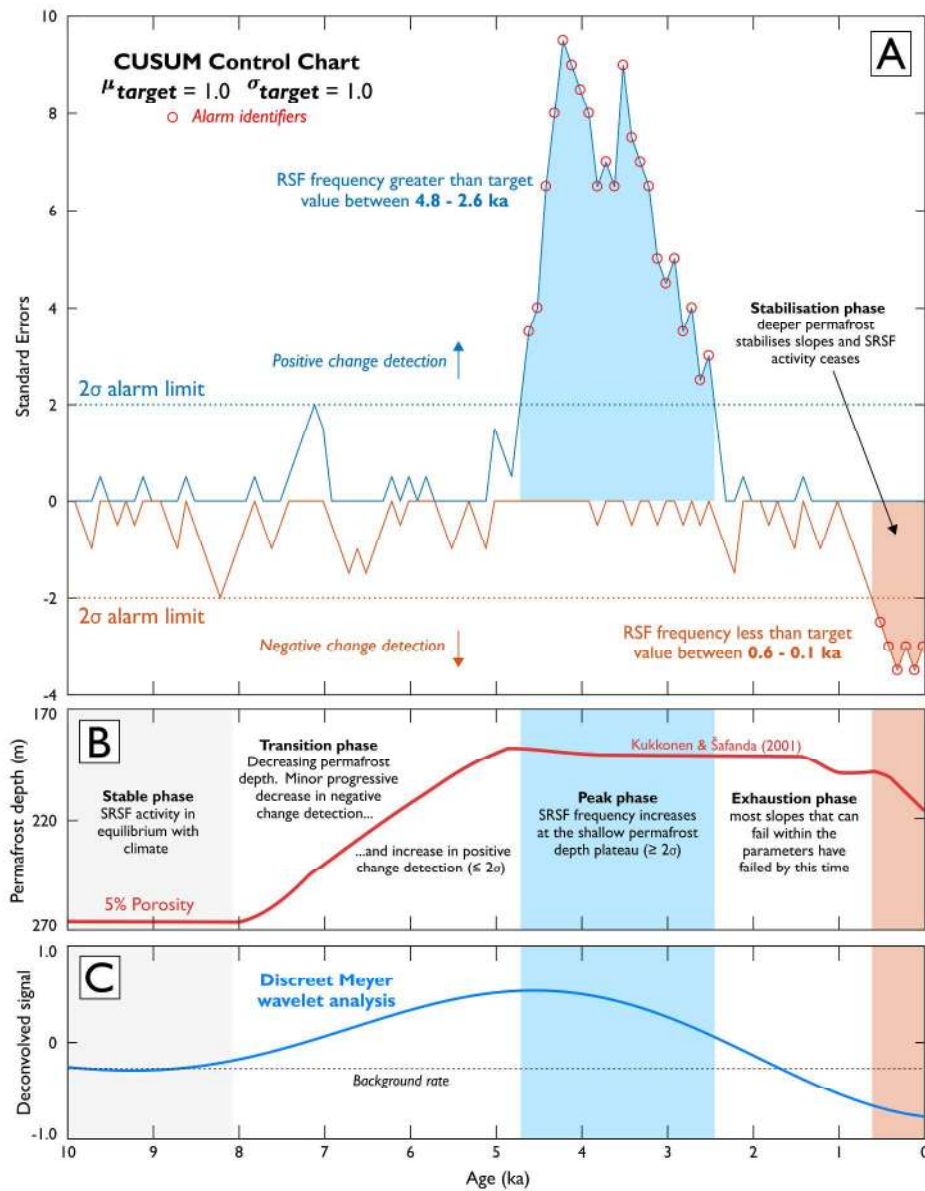


Fig. 8. Change detection and related analyses: (A) cumulative sum change detection graph showing positive (blue) and negative (orange) changes and statistically significant departures ( $> 2\sigma$ ) from the background SRSF frequency; (B) modelled permafrost depth in Fennoscandia (5% porosity) from Kukkonen & Šafanda (2001), subdivided into five distinct phases; (C) results of discrete Meyer wavelet analysis, showing the lowest frequency decomposed signal (d6).

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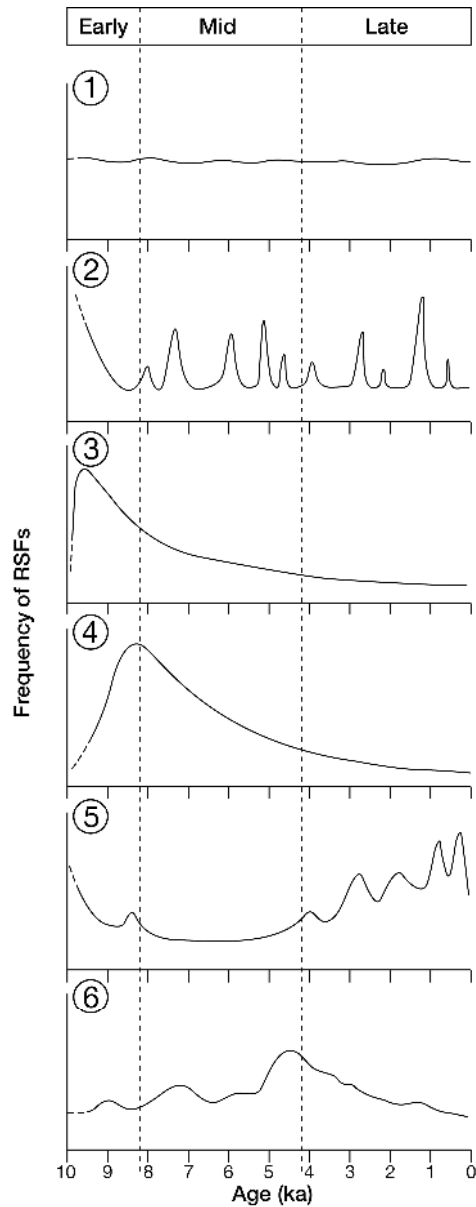


Fig. 9. Models for different patterns and causes of Holocene variations in RSF frequency and/or magnitude: (1) continuity-of-activity; (2) intermittent-earthquakes; (3) deglaciation-close-tracking; (4) deglaciation-lagging; (5) cool/wet-climate-response; and (6) the new thermally-driven permafrost-degradation model proposed in this study for SRSFs in Jotunheimen. The subdivisions of the Holocene shown are those proposed by Walker et al. (2012).

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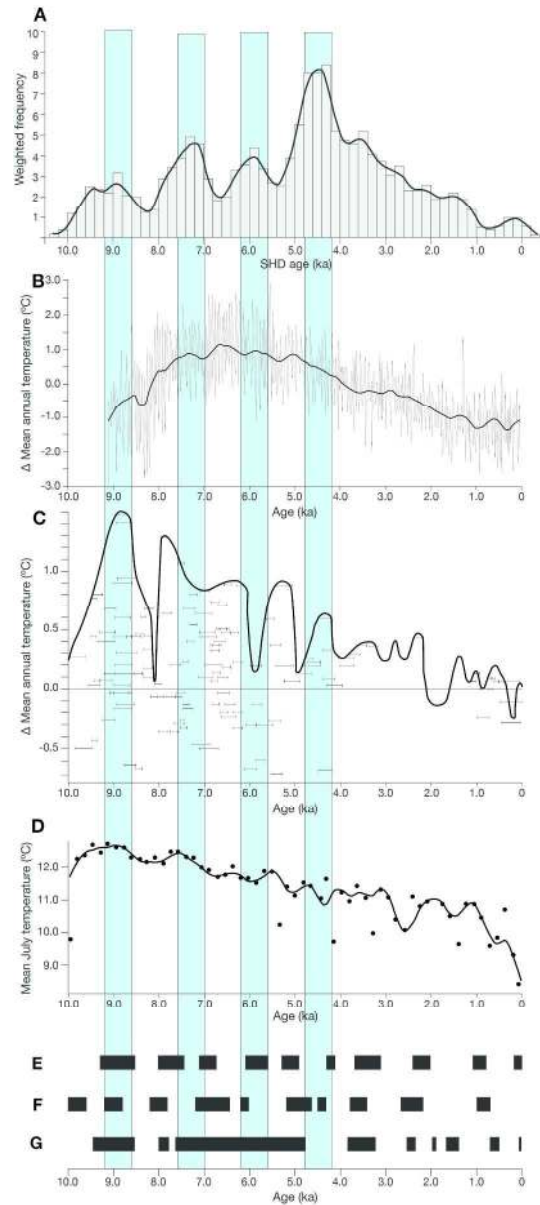
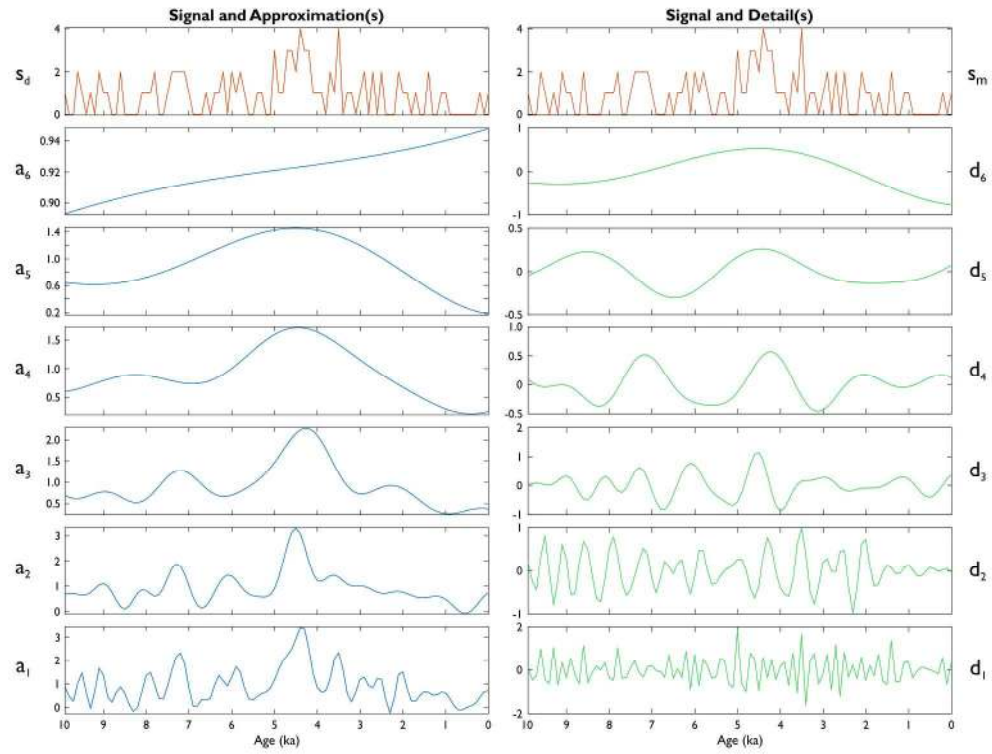


Fig. 10. Relationships between SRSF frequency in Jotunheimen and proxy climatic records: (A) temporal variations in SRSF frequency from Fig. 6C; (B) pollen-based reconstruction of annual air temperature for Northern Europe expressed as deviations from the mean (Seppä et al., 2009); (C) mean summer air temperature deviations from present in the Scandes Mountains based on pine tree-limit variations (Dahl and Nesje, 1996); (D) pollen-based July air temperature variations at Øvre Heimdalsvatnet, eastern Jotunheimen (Velle et al., 2010); (E) periods of above average air temperature (shaded) based on the GISP 2 Greenland ice core  $\delta^{18}\text{O}$  record (Alley, 2004; Wanner et al., 2011); (F) periods of above average sea-surface temperatures in the North Atlantic Ocean (shaded) based on standardized stacked ice-rafted debris (IRD) records (Bond et al., 2001; Wanner, et al., 2011); (G) periods when glaciers in the Smørstabbtindan massif, Jotunheimen, were smaller than today (shaded) based on glaciolacustrine and glaciofluvial stratigraphy (Matthews and Dresser, 2008). Vertical bands indicate phases of enhanced regional SRSF frequency (as in Fig. 6).

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