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The triggering factors of the Móafellshyrna debris slide in northern Iceland: intense precipitation, earthquake activity and thawing of mountain permafrost.

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Abstract

On the 20th September 2012, a large debris slide occurred in the Móafellshyrna Mountain in the Tröllaskagi peninsula, central north Iceland. Our work describes and discusses the relative importance of the three factors that may have contributed to the failure of the slope: intense precipitation, earthquake activity and thawing of ground ice. We use data from weather stations, seismometers, witness reports and field observations to examine these factors. The slide initiated after an unusually warm and dry summer followed by a month of heavy precipitation. Furthermore, the slide occurred after three seismic episodes, whose epicentres were located ~60 km NNE of Móafellshyrna Mountain. The main source of material for the slide was ice-rich colluvium perched on a topographic bench. Blocks of ice-cemented colluvium slid and then broke off the frontal part of the talus slope, and the landslide also involved a component of debris slide, which mobilised around 312 000 - 480 000 m³ (as estimated from field data and aerial images of erosional morphologies). From our analysis we infer that intense precipitation and seismic activity prior to the slide are the main preparatory factors for the slide. The presence of ice-cemented blocks in the slide's deposits leads us to infer that deep thawing of ground ice was likely the final triggering factor. Ice-cemented blocks of debris have been observed in the deposits of two other recent landslides in northern Iceland, in the Torfufell Mountain and the Árnesfjall Mountain. This suggests that discontinuous mountain permafrost is degrading in Iceland, consistent with the decadal trend of increasing atmospheric temperature in Iceland. This study highlights a newly identified hazard in Iceland: landslides as a result of ground ice thaw. Knowledge of the detailed distribution of mountain permafrost in colluvium on the island is poorly constrained and should be a priority for future research in order to identify zones at risk from this hazard.

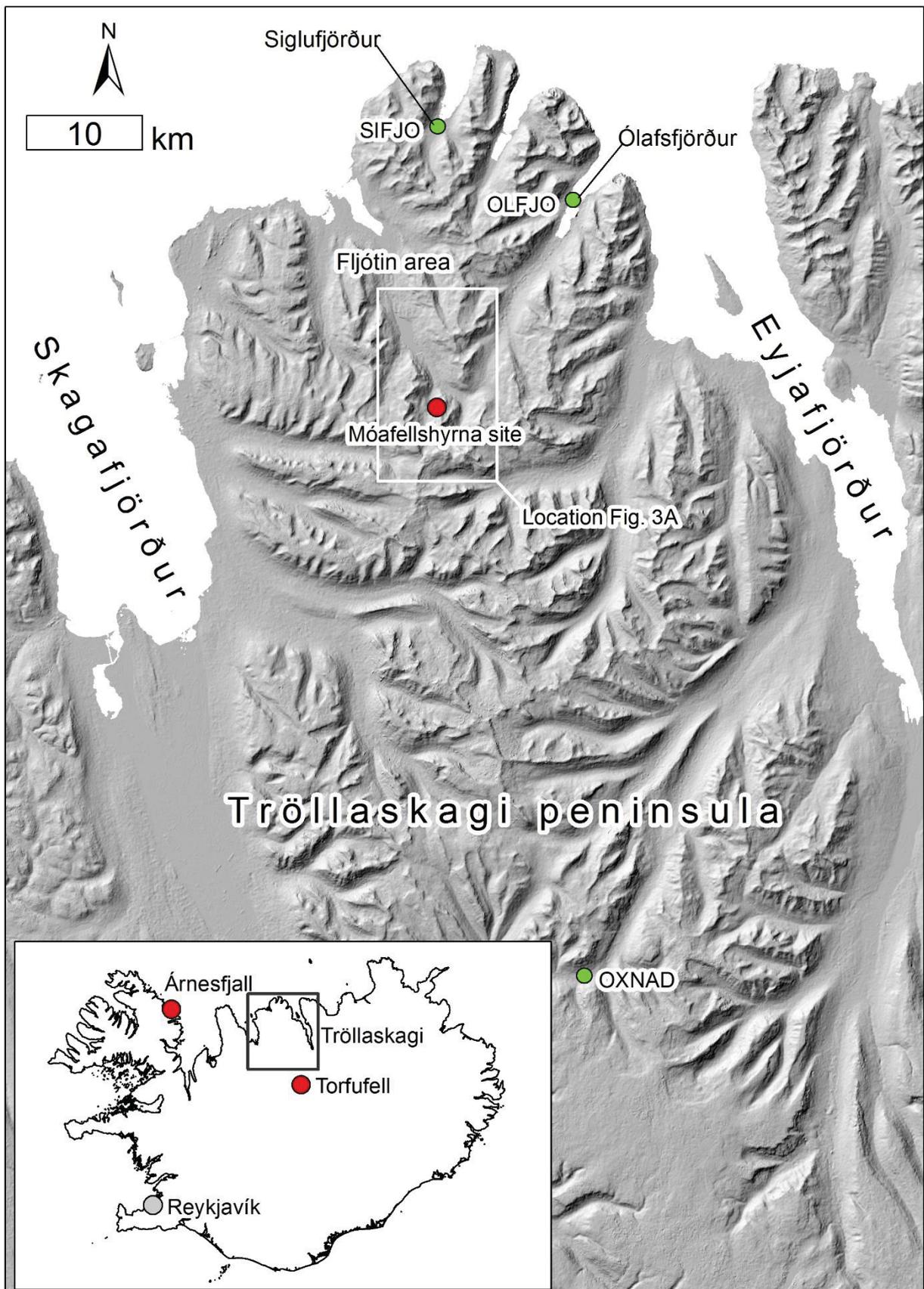
37 **1. Introduction**

38 On the 20th September 2012, the local residents of the Þrasastaðir farm in the Fljótin
39 area (Tröllaskagi peninsula, central north Iceland; Figs. 1,3), heard a loud rumbling noise
40 coming from the Móafellshyrna Mountain, as a large debris slide descended the north-
41 western side of the mountain (Fig. 3). Nine days after the failure, we observed large
42 blocks of ice-cemented deposits within the debris slide deposits.

43 Rapid mass movements, including rock falls, rock avalanches, debris flows and debris
44 slides, are common geomorphological processes in Iceland and present a significant and
45 direct threat to many towns, villages and farmhouses (Decaulne 2005; 2007).

46 Precipitation, snow melt, temperature variations and earthquake activity are the most
47 common triggering factors for landslides in Iceland (Sæmundsson *et al.*, 2003;
48 Sæmundsson and Decaulne, 2007, Decaulne & Sæmundsson 2007). However, during
49 the last decade, three, somewhat unusual, rapid mass movements have occurred in
50 northern Iceland: on the Torfufell Mountain (Eyjafjörður valley, central north Iceland) on
51 14th October 2011, on the Móafellshyrna Mountain on 20th September 2012 (case study
52 of this paper), and on the Árnesfjall Mountain (Westfjords) on 10th July 2014 (Sæmunds-
53 son *et al.*, 2013; Sæmundsson *et al.*, 2014 a/b). In all these landslides ice-cemented
54 debris was found within the deposits, a phenomenon that has never been reported
55 previously in Iceland. The source areas of these slides are all located on steep (~45-60°)
56 NW to NE facing-slopes, where discontinuous permafrost might be expected (e.g.
57 Gorbunov, 1988; King, 1986). The source areas at Torfufell and Móafellshyrna are
58 located at elevations of 750-870 m a.s.l., within the zone of discontinuous permafrost in
59 Iceland as calculated by Etzelmüller *et al.* (2007a), whereas at Árnesfjall the source area
60 is located at about 400 m a.s.l., which is much lower altitude than ever observed for
61 mountain permafrost in Iceland (Etzelmüller *et al.*, 2007a). Intense precipitation was
62 recorded prior to all of these slides (Sæmundsson *et al.*, 2013; Sæmundsson *et al.*, 2014
63 a/b). In this paper, we present the case study of the Móafellshyrna debris slide, where we
64 examine three factors could have contributed to this failure: intense precipitation,
65 earthquake activity and ground-ice degradation (via increased annual temperatures). We
66 emphasise the field evidence of ground ice-thaw, because permafrost degradation has
67 never previously been considered as a triggering factor for gravitational mass
68 movements in Iceland, although permafrost degradation is considered to be an
69 preparatory factor for paraglacial slope failure (McColl 2012).

70 In the following sections we describe a) the state of knowledge of permafrost in Iceland,
71 b) seismic conditions in Iceland and their role in previous mass wasting events, c)
72 general meteorological conditions in Iceland. We then report our results, reconstructing
73 the conditions that favoured the occurrence of the landslide. We then discuss our results,
74 dividing the factors that induced the landslide into, i) preparatory and ii) triggering (as per
75 McColl, 2012). We finally posit that Móafellshyrna landslide was preconditioned by
76 combination of intense precipitation in the weeks prior to the slide and the seismic activity
77 on 18th and 19th September and that the degradation of ground-ice was the final trigger.



Station name	Station ID	Altitude	Years analysed	Precipitation	Temperature
Siglufjörður	SIFJO	6.0 m a.s.l.	2012	X	X
Öxnadalshéiði	OXNAD	540.0 m a.s.l.	2005 - 2012	-	X
Ólafsfjörður	OLFJO	5.0 m a.s.l.	2000 - 2012	X	X

79 Figure 1 – The Móafellshyrna site, located in the Tröllaskagi peninsula, northern Iceland (see Fig. 3 for
80 detailed location), and the location of the weather stations used for this study. The hillshaded digital
81 elevation model used as a basis of this map is from the Digital Elevation Model over Europe (EU-DEM)
82 from the Global Monitoring for Environment and Security service for geospatial reference data access
83 project (GMES RDA). The table provides details on the Icelandic Meteorological Office weather stations,
84 whose datasets have been used for this study. Symbols “X” and “-“ mean that data are or are not available
85 at the stations, respectively.

86 **1.1 Permafrost distribution in Iceland**

87 During the last decade our knowledge of the regional distribution of permafrost in Iceland
88 has increased considerably (e.g. Etzelmüller *et al.*, 2007a/b; Farbrot *et al.*, 2007a/b;
89 Kellerer-Pirklbauer, 2007; Lilleören *et al.*, 2013; Kneisel *et al.*, 2007; Kneisel, 2010).
90 There are several published works on permafrost in the central highlands of Iceland
91 dating back to the 1950s, and these studies focused on geomorphological features, such
92 as palsas and patterned ground (e.g., Thorarinsson, 1951; 1964; Friedman *et al.*, 1971;
93 Schunke, 1974; Stingl and Herrmann, 1976; Priesnitz and Schunke 1978; Hirakawa,
94 1986; Thorhallsdottir, 1994; 1996; 1997; Sæmundsson *et al.*, 2012). These works both
95 mapped permafrost conditions and a recent study at the Orravatnsrústir palsa site, NE of
96 the Hofsjökull ice cap, showed clear indications of declining permafrost conditions from
97 2000-2010 in the highlands (Sæmundsson *et al.*, 2012). Studies have also sought to
98 better define the spatial distribution of mountain permafrost in Iceland. Such studies used
99 inventories of rock glaciers and stable ice-corded moraines combined with meteorological
100 data analyses (e.g., Etzelmüller *et al.*, 2007a/b; Farbrot *et al.*, 2007a/b; Kellerer-
101 Pirklbauer, 2007; Lilleören *et al.*, 2013) and Electrical Resistivity Tomography (e.g.,
102 Kneisel *et al.*, 2007; Kneisel, 2010; Sæmundsson *et al.*, 2012). These works did not
103 comment on the dynamics of the permafrost.

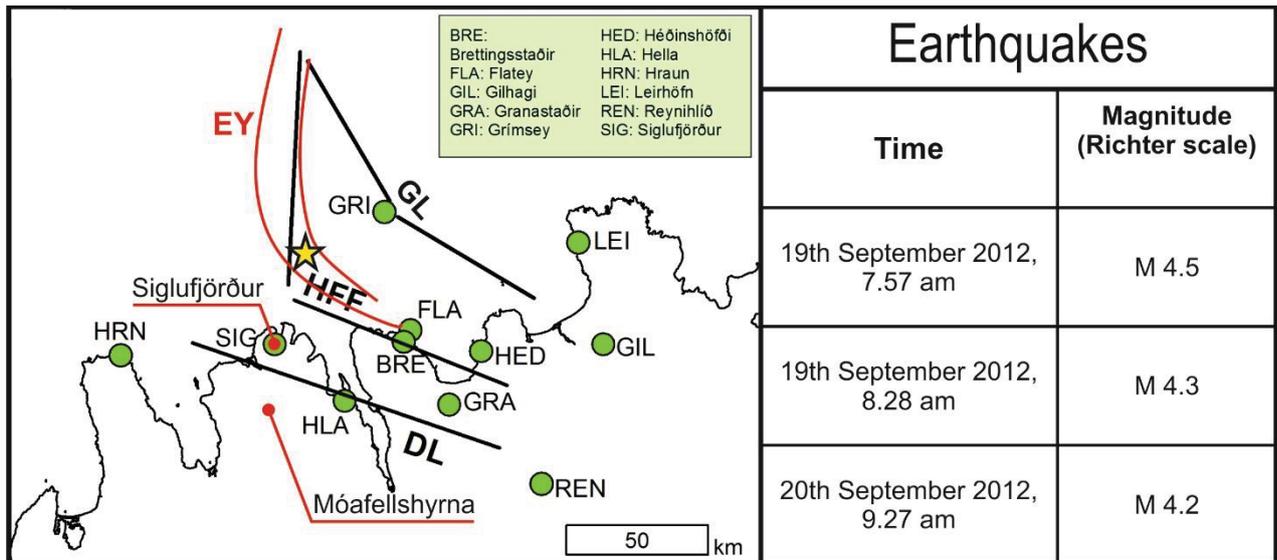
104 Etzelmüller *et al.* (2007a) published the first regional distribution map of mountain perma-
105 frost in Iceland. This was the first attempt to understand the overall extent of mountain
106 permafrost on the island. The map was based on a model which used meteorological
107 data calibrated with ground surface temperature data and validated using ground
108 temperature data from four shallow boreholes and an inventory of rock glaciers.
109 According to Etzelmüller *et al.* (2007a), mountain permafrost is widespread in the
110 northern and eastern Iceland above 800-900 m, covering 8% of the island, or around
111 8.000 km².

112 A warming trend in atmospheric temperature has been observed in Iceland over the last
113 two centuries. In Stykkishólmur in western Iceland records show an increase of about
114 0.7°C per century (Nawri & Björnsson, 2010; Jónsson, 2008). From 1975 to 2008 the
115 warming rate was much higher or about 0.35°C per decade (Ministry for the Environment
116 of Iceland, 2010). However, so far relatively little attention has been paid to the
117 consequences of the recent climate change on the possible degradation of mountain
118 permafrost in Iceland.

119 **1.2 Seismic activity in northern Iceland**

120 The seismic activity in Iceland is related to its position on the Mid-Atlantic plate boundary,
121 which crosses the island and its location over the Icelandic Hotspot (e.g. Tryggvason *et*
122 *al.*, 1983; Wolfe *et al.*, 1997; Allen *et al.*, 2002; Bjarnason, 2008; Einarsson, 2008;
123 Thordarson & Höskuldsson 2002). The seismic activity in the northernmost region of the
124 island is related to the Tjörnes Fracture Zone (Einarsson and Björnsson, 1979;
125 Sæmundsson, 1974), which is defined by three seismically active lineaments - the
126 Grímsey Oblique Rift, the Húsavík Flatey Fault and the Dalvík lineament; (e.g.,
127 Sæmundsson, 1974; Gudmundsson, 2000; 2006) - and includes the Eyjafjarðaráll N-S
128 extensional graben, located north offshore the Tröllaskagi peninsula (Fig. 2). It was
129 activity on this fault system which caused the earthquakes prior to the Móafellshyrna
130 slide.

131 Earthquakes are known to have triggered landslide and rockfall activity in Iceland in the
132 past (e.g. Jónsson, 1957; Jónsson and Pétursson, 1992; Thorarinsson, 1937, 1959;
133 Halldórsson, 1984; Jensen, 2000; Sæmundsson *et al.*, 2003; Ágústsson and Pétursson,
134 2013), but no study has explored in detail the influence of earthquake activity as a
135 preparatory and/or triggering factor on rockfall or landslides in Iceland, as has been done
136 elsewhere in the world (e.g., Harp and Jibson, 1996; Yin *et al.*, 2009). The above
137 mentioned Icelandic studies relate mass movements to larger earthquakes than those
138 prior to the slide in Móafellshyrna, e.g. in June 2000, when two earthquakes of magnitude
139 6.4 occurred in Iceland, with the epicentre in the middle of the southern lowlands.
140 Rockfall activity was reported as a result of this event as far as 75 km from the epicentre
141 (Sæmundsson *et al.*, 2003).



142
 143 Figure 2 - The structural elements of the Tjörnes Fracture Zone marked in black (Grímsey lineament
 144 (GOR), Húsavík-Flatey fault (HFF) and Dalvík lineament (DL); from Stefánsson *et al.*, 2008) and the
 145 Eyjafjarðaráll graben marked in red (EY); the position of the epicenter zone for the earthquakes preceding
 146 the Móafellshyrna slide is marked with a yellow star, the Icelandic Meteorological Office (IMO)
 147 seismometers in the area marked with the green dots and the labels refer to their abbreviated names, as
 148 given in full in the key. On the right are reported the timing and magnitude of the earthquake sequence in
 149 the Eyjafjarðaráll graben from 19th to 20th September 2012 (from Gudmundsson *et al.*, 2014).

150 **1.3 General weather conditions in Iceland**

151 Weather patterns in Iceland are highly variable, with frequent and strong variation in
 152 precipitation and temperature; this is mainly because Iceland is located on the main path
 153 taken by North Atlantic low-pressure systems (Einarsson 1984). The mean annual air
 154 temperature for the period 1971–2000 was 4–5°C in the south, 3–4°C in the east and
 155 west and 2–3°C in northern coastal parts of the country (Tveito *et al.*, 2000). Hence
 156 precipitation can fall as both snow and rain. The two main dominant precipitation wind
 157 direction in the Tröllaskagi area are NE and SW (Brynjólfsson and Ólafsson 2008,
 158 Arnalds *et al.*, 2001). The precipitation is heaviest during strong NE winds.
 159 Consequently, mean annual precipitation increases from about 500 – 1000 mm per year
 160 in the central and northern parts of the country to more than 3000 mm/yr in the southeast
 161 (Crochet *et al.*, 2007). During the winter months from October to April the precipitation in
 162 the outer part of the Tröllaskagi peninsula is almost exclusive snow or sleet and the main
 163 part of the snow avalanche activity is associated with strong north-easterly wind. The
 164 northern part of Tröllaskagi peninsula is generally a heavy snow prone area and
 165 Siglufjörður and Fijótín area are generally considered to be one of the heaviest snow
 166 prone areas in Iceland (Arnalds *et al.* 2001).

167 **2. Geographic and Geologic Setting of Móafellshyrna**

168 The Tröllaskagi peninsula is a mountain massif located between the Eyjafjörður fjord in
169 the east and the Skagafjörður fjord in the west (Fig. 1). The peninsula is topped by flat
170 summits reaching up to 1000 – 1500 m a.s.l. and sculptured by glacial erosion with
171 glacially carved fjords, valleys and cirques. Such over-steepened landscapes are
172 recognised as one of the key pre-conditioning factors for failures (McColl 2012). Over
173 150 alpine glaciers, mainly north facing, have been mapped in this massif (Sigurðsson
174 and Williams, 2008).

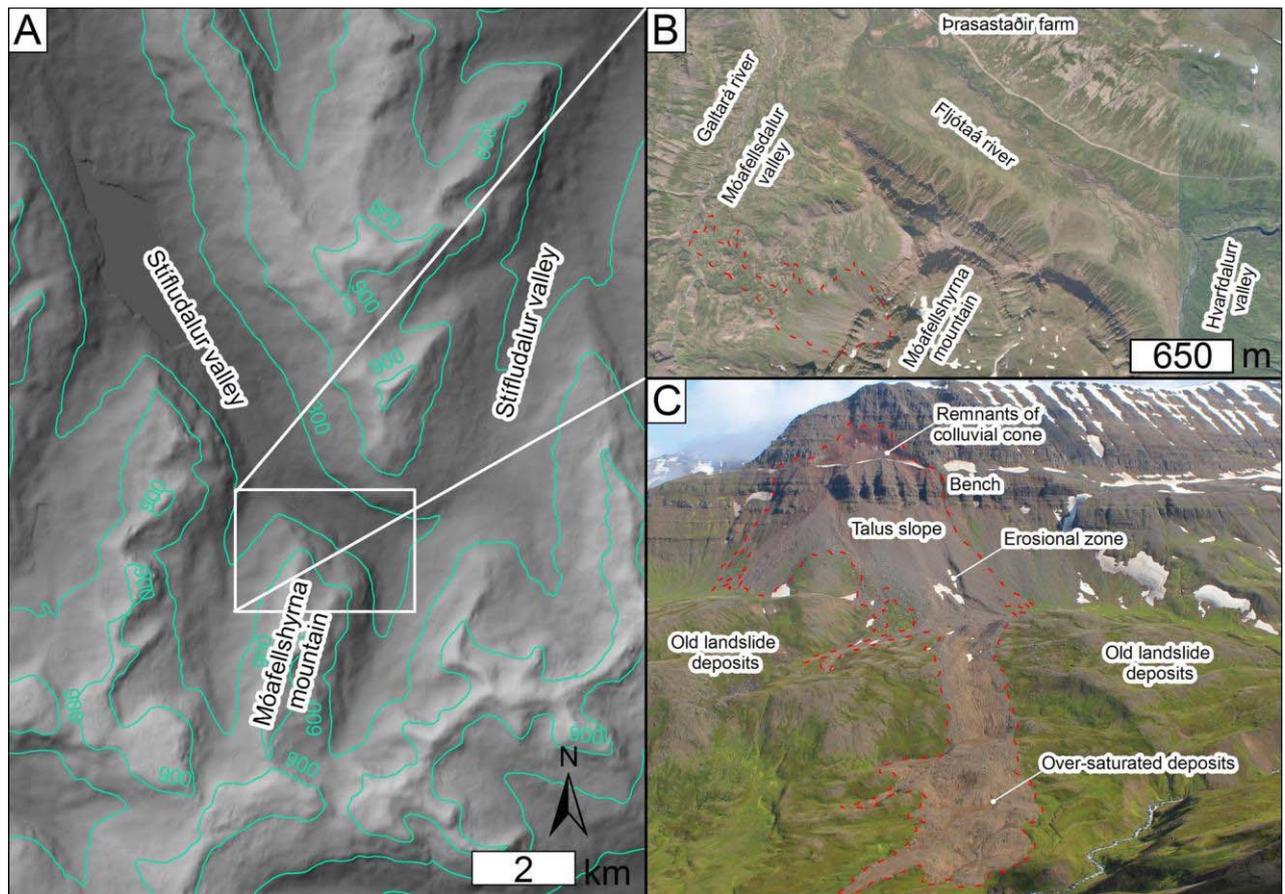
175 The bedrock in the outermost part of the Tröllaskagi peninsula falls within the Tertiary
176 basalt series (16-3.3 M years; e.g., Moorbath *et al.*, 1968; McDougall *et al.*, 1984;
177 Watkins and Walker, 1977). This series is mostly jointed basaltic lava flows, composed of
178 2 to 30 m thick individual flows separated by lithified sedimentary horizons (from few
179 centimetres up to tenths of meters thick) (Sæmundsson 1979; Sæmundsson *et al.*, 1980).
180 The bedrock is heavily jointed and intersected by dikes, and the general dip angle of the
181 lava beds is towards the southwest (Sæmundsson *et al.*, 1980; Johannesson and
182 Sæmundsson, 2009; Hjartarson & Sæmundsson 2014).

183 The Tröllaskagi peninsula is located on the main path of the North Atlantic low pressure
184 system, with a variable and turbulent weather. Winters are long and mild and summers
185 brief and cool (Einarsson, 1984). The Mean Annual Air Temperature (MAAT) for the
186 period 1971-2000 was 2-3°C in the northern coastal areas (Tveito *et al.*, 2000; Crochet *et*
187 *al.*, 2007). The data series for the Tröllaskagi area between 1940-1970 show MAAT of 2-
188 4°C in the coastal areas and -2 to -4°C at the summits (Einarsson, 1964). According to
189 Crochet *et al.* (2007), the annual precipitation from 1971-2000 in the Tröllaskagi area
190 varies from 1000-1500 mm in the coastal lowlands up to 2000-2500 mm on the summits.
191 Localised orographic effects mean that in the Tröllaskagi peninsula precipitation is higher
192 near the coastline when there are northerly winds (Brynjólfsson and Ólafsson 2008).
193 Conversely, the precipitation is likely to be higher in the lowlands in the interior of the
194 Tröllaskagi peninsula area than at the coast during periods with southerly winds. In the
195 mountains of the peninsular the orographic effect would also play a role during southerly
196 winds precipitation may in fact be even higher at a higher elevation. According to Arnalds
197 *et al.* (2001), more than half of the recorded precipitation in Siglufjörður occurs
198 throughout the winter months (Oct-Apr), where the precipitation is almost exclusively
199 snow or sleet. This can also be said about the rest of Tröllaskagi peninsula. Avalanches

200 during winter months in Tröllaskagi peninsula can be associated with strong north-
201 easterly winds with snowfall or snowfall during south-westerly winds associated with
202 lower wind speed. Slush flows are not common in Tröllaskagi area.

203 The Móafellshyrna Mountain is located in the Fljótin area in the outermost part of the
204 Tröllaskagi peninsula in central north Iceland (Figs.1, 3). The mountain is 1044 m high,
205 with narrow alpine-mountain ridge orientated in the north-south direction. It is located
206 between the Móafellsdalur tributary valley in the west and the Hvarfdalur tributary valley
207 in the east (Fig. 3). In the outermost part of the western valley side of Móafellsdalur are
208 old remnants of a landslide that originated from the western mountain side. The age of
209 these old landslide deposits is not known. Jónsson (1957, 1976) stated that the landslide
210 activity in Iceland was most intense shortly after the last deglaciation. His statement has
211 been confirmed by later studies (e.g. Jonson *et al.* 2004, Sigurgeirsson and Hjartarson
212 2011, Mercier *et al.* 2012 and 2013, Coquin *et al.* 2015 and 2016, Decaulne *et al.* 2016).
213 The farm Þrasastaðir, the innermost inhabited farm in the Stífludalur valley, is situated on
214 the northern side of the Stífludalur on the northern side of the Fljótaá River facing the
215 Móafellshyrna mountain. The residents of the farm witnessed and observed the slide
216 (Fig. 3) and their report is given in Section 4.1.

217 The landscape in the Fljótin area is predominantly carved by glacial erosion. The area
218 was heavily glaciated during the maximum extent of the last glaciation, but the
219 deglaciation history of the area is not well documented. Researchers suggest that the
220 area was deglaciated in Early Preboreal times (e.g., Norðdahl and Pétursson, 2005;
221 Norðdahl *et al.*, 2008; Ingólfsson *et al.*, 2010; Coquin *et al.* 2016). The upper part of the
222 valley sides (up to 600-900 m a.s.l.) are often very steep, with fractured and loose
223 bedrock produced by, erosion during deglaciation, unloading by the glacier, frost
224 shattering and freeze-thaw processes, while the lower parts are more gentle and
225 generally covered by glacial deposits or talus.



226
 227 Figure 3 – Geographical setting of Móafellshyrna mountain (see Fig. 1 for location). (A) Hillshaded digital
 228 elevation model and contours (in green, metres above sea level) of the Móafellshyrna region. Elevation
 229 data are from EU-DEM from GMES RDA. (B) Aerial photograph of the Móafellshyrna site taken before the
 230 slide in 2012 from the website Samsyn. (C) Oblique photo of the Móafellshyrna debris slide in July 2015
 231 taken by Costanza Morino. In panels B and C the perimeter of the slide is marked with red dashed line
 232 taken from the trimble data for the deposits and reconstructed from photographs for the upper part, in C
 233 this line has been manually traced onto the oblique image.

234

235 **3. Methods**

236 The triggering factors of the Móafellshyrna debris slide event in September 2012 were
 237 analysed using meteorological data from the Icelandic Meteorological Office (IMO) for the
 238 three months prior to the slide and for the period 2010-2012. Data for the seismic activity
 239 of the north coast were also obtained from the IMO. We also interviewed the inhabitants
 240 of the valley and performed our own field investigations.

241 **3.1 Meteorological data**

242 Only two weather stations in the northern part of the Tröllaskagi peninsula measure
 243 precipitation, one located in the town of Siglufjörður at 6 m a.s.l. (35 km north of the site)

244 and the other one in the town of Ólafsfjörður at 5 m a.s.l. (21 km northeast of the site).
245 Three weather-stations were used for this study that collect temperature records: two
246 operated by IMO - Siglufjörður (WMO (World Meteorological Organization) ID: 4157),
247 Ólafsfjörður (WMO ID: 4155) and one station located at the Öxnadalsheiði highlands
248 pass (WMO ID: 4859) at 540 m a.s.l. operated by the Icelandic Road and Coastal
249 Administration (Fig. 1). A problem in our approach is that the majority of the stations used
250 to establish mean atmospheric temperatures are located near the coast and therefore at
251 low altitudes. This leads to a potential bias when evaluating trends in temperature,
252 because such stations may not be representative of the atmospheric temperatures
253 experienced in the highlands. To overcome this, we applied the environmental lapse rate
254 of 0.649°C per 100 m (Sheridan *et al.* 2010) to the mean temperatures recorded at all
255 three stations as an estimate of the temperature at the source zone of the Móafellshyrna
256 debris slide. We do not attempt to correct the precipitation data collected near the coast
257 for the inland conditions, because this would not only require a temperature correction,
258 but we would need to take into account variations in wind speed, wind direction and
259 pressure in a meteorological model, which is beyond the scope of this paper.

260 **3.2 Direct report from witnesses**

261 The local residents of the farm Prasastaðir witnessed the release of the landslide, and
262 were interviewed on the day of the event regarding the earthquake activity prior to the
263 slide, timing of the slide and the events that occurred during the first few hours of the
264 slide. The slide was photographed by the first author only a few hours after it had
265 occurred. Ongoing rockfall activity on the slope prohibited more detailed observations
266 immediately after the event. It was not considered safe to perform field analysis until 29th
267 September 2012, when direct observations and measurements were made. The site was
268 revisited in summer 2015 to track changes in the slide.

269 **3.3 Field measurements and survey**

270 The boundary of the landslide body was mapped on the 29th June 2012 with Trimble
271 GEOXT from the GeoExplorer CE series with an accuracy of 1-2 m. Thickness
272 measurements of the slide were performed both with direct measurements in the field on
273 29th September 2012 and from aerial photographs and erosional and depositional
274 features mapped. For geomorphological mapping both ground photographs and aerial
275 photographs from the National Land Survey of Iceland (www.lmi.is) and Loftmyndir ehf
276 (www.map.is) and the UK's Natural Environment Research Council Airborne Research

277 and Survey Facility (NERC ARSF) flown in 2015 were used in concert with field
278 observations. Landslide volume estimates were derived from morphometric properties of
279 the deposits measured directly in the field and from aerial photographs.

280 **4. Results**

281

282 **4.1 Witness report**

283 The residents of the Þrasastaðir farm, located at the junction between the Móafellsdalur
284 and the Stífludalur valleys and 1.7 km from the terminal deposits of the landslide (Fig. 3),
285 were interviewed only few hours after the slide. They recounted that on the 20th of
286 September 2012 at around 12:30 they heard a rumbling noise, which originated from the
287 Móafellshyrna mountain. They also recounted that a black tension crack, in the snow
288 covered mountain, progressively formed above the colluvial cone at around 850 m a.s.l.
289 They saw large blocks of debris that broke off the frontal part of the cone and fell onto the
290 talus slope below. This activity was most intense in the first 1-1.5 hours, but they reported
291 that there were intermittent noises and rock fall activity throughout the day.

292 The residents of the Þrasastaðir farm felt all the three earthquakes that occurred on the
293 19th and 20th September, with the last one only three hours before the debris slide event.
294 It was estimated that less than 1 m of snow was on the ground at the time of the slide.

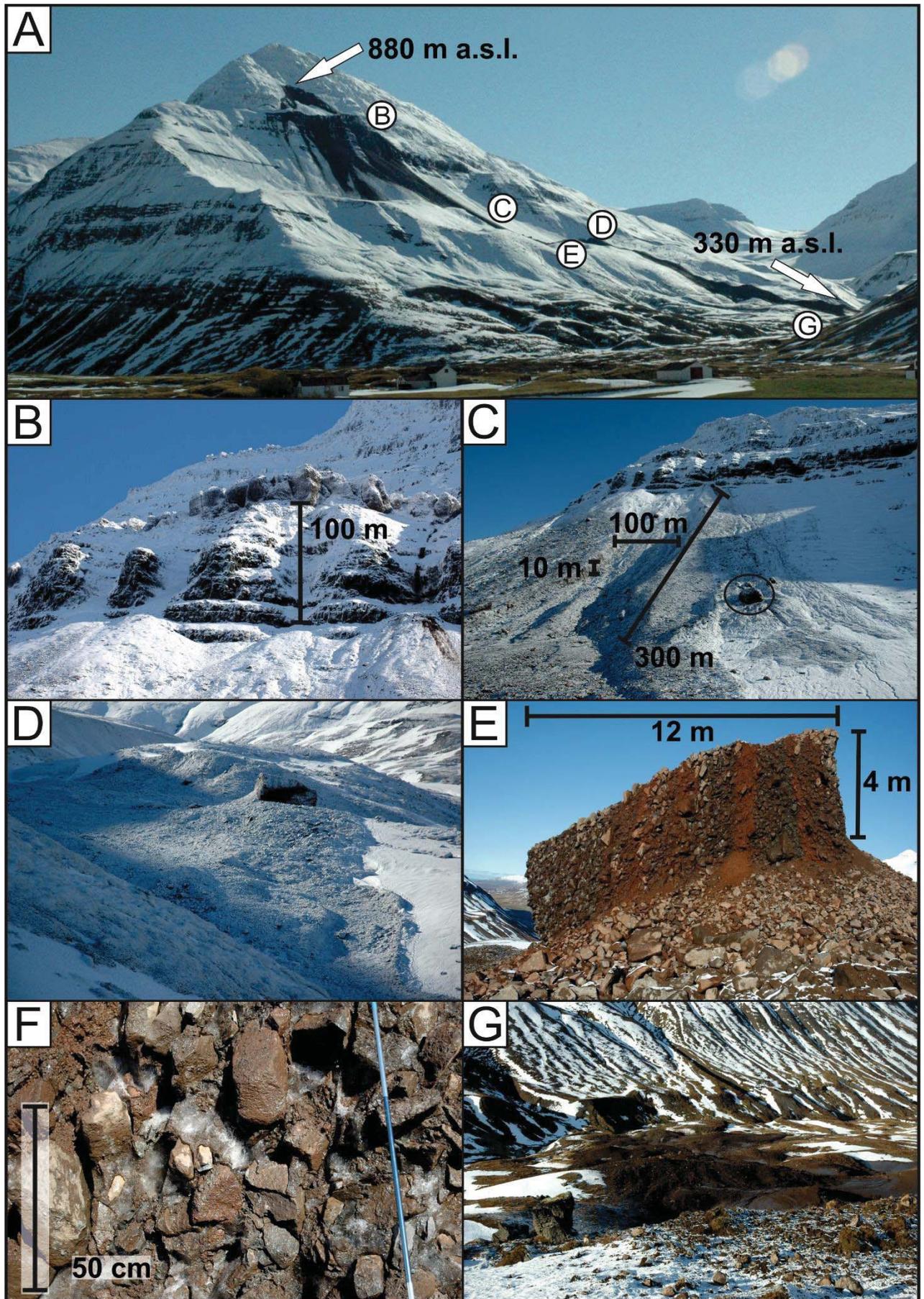
295 **4.2. The debris slide morphology**

296 The debris slide detached from a main scarp at an elevation of 880 m a.s.l. (Fig. 4A). The
297 source material of the landslide was a colluvial cone composed of talus deposits perched
298 on a topographic bench at 790 m a.s.l. (Fig. 3C), and lying against a rockwall composed
299 of lava and tephra layers of the Tertiary Basalt formation dipping less than 10° towards
300 SSW. The colluvial material comprises a mixture of materials derived from the above
301 rockwall (basalt with intercalated sediment layers) and under nominal conditions (with or
302 without ice) should lie stable against the rockwall under the angle of repose.

303 The colluvial material partially slid off the steep rockwall ledge at the edge of the bench,
304 as described by the local residents. From field observations and aerial photographs we
305 estimate that the horizontal displacement in the upper part of the cone was around 40 m
306 (Fig. 4A). Later observations revealed that the bedrock was exposed from beneath the
307 colluvium as it slid (at the time of the slide it was hidden by a thin veneer of mud, which
308 was later cleaned by rainfall). Part of the colluvial deposits remained perched on the

309 topographic bench after the event (Fig. 3C, 4B). Based on field observation, we
310 estimated that the thickness of the frontal part of the still ice-cemented colluvial cone that
311 was preserved at the edge of the bench after the slide was around 20-30 m thick (Fig.
312 4B). As the colluvial cone slid off the rockwall edge, large blocks of ice-cemented debris
313 broke off and fell onto the talus slope located below the topographic bench (Fig. 3C). Fig.
314 3B shows decametre-scale blocks of ice cemented colluvium that have remained on the
315 bench. A 300 m long, 80-100 m wide and 8-10 m deep channel was carved into the talus
316 slope deposits (Fig. 3C, 4C), entraining further material to the landslide. This channel
317 shows that the landslide materials must already have been fluidised (saturated) in order
318 to scour out this channel.

319 At 495-505 m a.s.l. at the foot of the talus slope, we observed that both loose muddy
320 debris and blocks of ice-cemented deposits comprised the landslide deposits (Figs.
321 4C,D,E). We infer that these ice cemented blocks had fallen from the topographic bench,
322 as they had the same aspect and size of those still perched in the source area (Fig. 4B).
323 These blocks ranged from less than 1 m up to 12 m wide and from less than 1 m up to 10
324 m tall. The ice-cemented blocks were composed of layers of sand to boulder-size (up to
325 50 cm in long-axis) angular rock fragments (Fig. 4F), interbedded with layers composed
326 of clay to sand-size materials. The red colour of the horizons of fine material (Fig. 4E)
327 derives from the original source material, namely paleo-soils interbedded in the lava
328 layers composing the Tertiary Basalt Formation. Nine days after the failure, from visual
329 inspections we estimated a content of ground ice that was cementing the blocks of
330 around 15-20% of the total volume (Fig. 4F), i.e. pore-filling ice, rather than excess ice.
331 No massive ice deposits were observed.



333 Figure 4 – (A) An overview of the western side of the Móafellshyrna Mountain after the debris slide. The
334 uppermost and lowermost elevation limits of the debris slide path are labelled in m a.s.l. as are the
335 locations of panels B to G. (B) The ice-cemented blocks perched, up to 20 m in thickness on the 100 m
336 high rockwall. (C) The erosional area in the lower talus slope below the rockwall. Note the large block of
337 ice-cemented deposits on the right. (D) Landslide debris and an ice-cemented block resting at 495-505 m
338 a.s.l. (E) The same large block of ice-cemented deposits as in D, around 12 m wide and 4 m high. (F)
339 Close up of the block in E showing stratified deposits of coarse angular clasts with an icy matrix. (G)
340 Looking downwards onto the deposits in the terminal part of the slide. Photos: Þ. Sæmundsson, A taken on
341 the 20th September and B-G taken on the 29th September 2012.

342 Some of the landslide deposits came to rest below the talus slope at 495-505 m a.s.l.
343 (Fig. 4D), where the average slope angle is less than 10°, but the rest traversed down the
344 mountain slope, finally stopping at 330 m elevation (Fig. 4G), a few meters from the river
345 Galtará draining the bottom of the valley (Fig. 3). The part of the material that continued
346 down the mountain side is composed of clay to boulder-size material which followed a
347 well-defined path with distinct lateral boundaries. The deposit boundaries are upstanding
348 with 1-2 m of relief. The lack of sorting in the deposits, their lobate planform and their
349 upstanding boundaries are consistent with this landslide being classed as a debris side
350 transforming to a debris avalanche from the Hungr *et al.* (2001, 2014) classification, i.e.
351 loose debris fluidised by the inclusion of water exceeding the pore-space (Iverson 1997)
352 without a well-defined central channel and generally shallow motion. This shows that
353 water volumes of 20-40% of the final deposits were required to mobilise this flow and that
354 it was not a dry rock avalanche.

355 Based on aerial photographs the dimensions of the upper colluvial cone perched on the
356 bench was around 80 m in height and around 150 m wide at the lower end prior to the
357 slide. Based on photographs taken after the slide the total movement of the cone was
358 estimated around 40 m and the frontal part of the cone perched on the bench was 150 m
359 wide and about 20-30 m thick (Fig. 4B). Based on these dimensions we estimated that
360 the volume of the deposits that broke off the frontal part of the cone and fell down onto
361 the lower talus had a volume up to ~120 000 – 180 000 m³. From the dimensions of the
362 channel carved into the lower talus on the southern side of the slide (with 80-100 m,
363 depth 8-10 m, length 300 m) (Fig. 4C) we calculated that an additional volume of ~192
364 000 - 300 000 m³ of material was mobilized. Combining these volumes gives an estimate
365 of the total volume of material mobilized in the event of 312 000 - 480 000 m³. From GPS
366 measurements the areal extent of the landslide is ~0.3 km², giving an average depth of

367 deposits of 1-2 m, consistent with our direct observations of thickness at the deposit-
368 boundaries.

369 **4.3 Antecedent conditions**

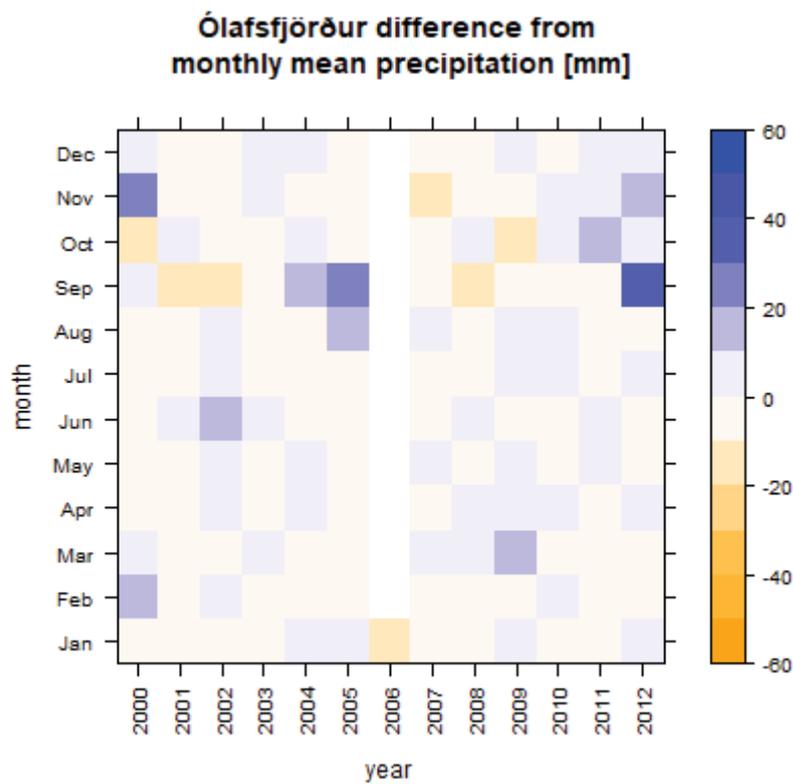
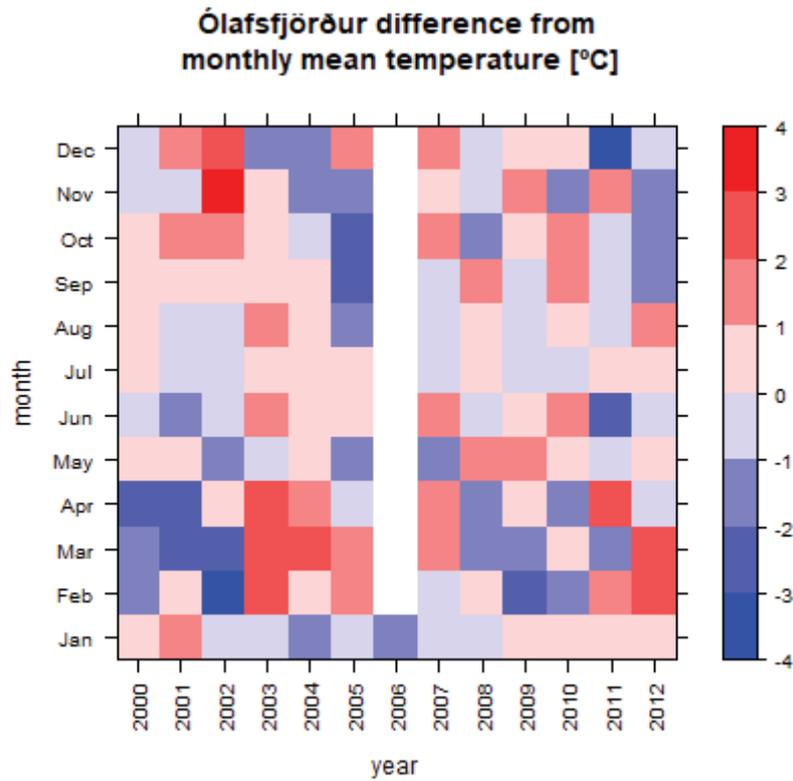
370 *4.3.1 Precipitation*

371 The spring and summer months preceding the Móafellshyrna event were dry, and the
372 autumn was unusually wet (Figs. 5, 6) (Jónsson 2013). From April until 28th August 2012,
373 dry conditions prevailed in the outer part of the Tröllaskagi peninsula, with only one day
374 with precipitation greater than 10 mm: 23rd to 24th July, when 70 to 90 mm of rain was
375 recorded at Siglufjörður and Ólafsfjörður weather stations (Fig. 7).

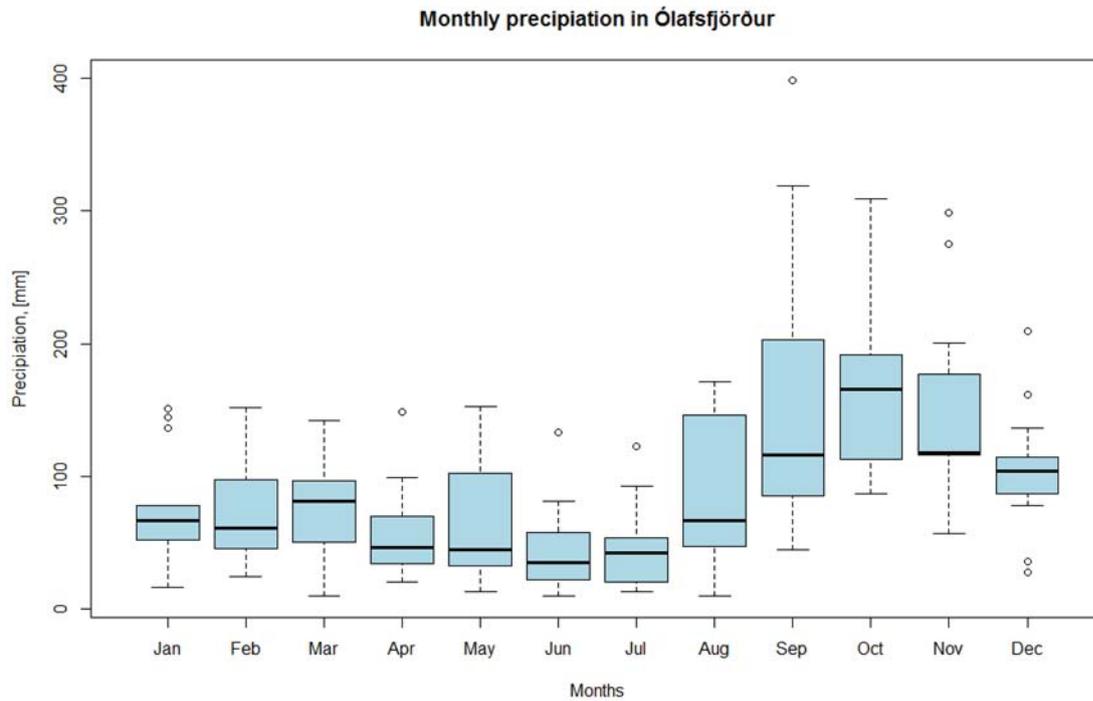
376 From 20th August to 20th September around 1/3 of the precipitation for 2012 fell in the
377 area (~400-550 mm). For comparison, the average annual precipitation in the town of
378 Ólafsfjörður is ~400 mm for the period 2000-2012 (Figs. 5 & 6). In detail, from 28th August
379 to 8th September the cumulative precipitation in Siglufjörður was 190 mm and 120 mm in
380 Ólafsfjörður, with an additional 30 to 40 mm precipitation at these stations on the 3rd, 6th
381 and 7th September. From 9th to 11th September an unseasonal and severe snowstorm hit
382 the north eastern and northern parts of the country (Jónsson 2013, Hermannsdóttir
383 2012). Following this snowfall, around 100 mm rain was measured in only two days at
384 Siglufjörður and almost 150 mm at Ólafsfjörður. The precipitation continued from 11th to
385 17th September either as snow, sleet or rain at these weather stations. From September
386 17th to 19th less than 10 mm of precipitation was recorded, but at the time of the event
387 540 mm precipitation had been recorded in Siglufjörður and 490 mm at Ólafsfjörður
388 weather stations since 23rd July, which corresponds to 40-45% of the mean annual
389 precipitation from 2000-2012 (Fig. 5). The monthly precipitation data from Ólafsfjörður
390 station from 2000 to 2012 show that September is the month with maximum precipitation
391 for any given year, with a range between 70 and 250 mm (Figs. 5 & 6). The year of 2012,
392 however, had precipitation exceeding the average for this month (Fig. 5)

393 Unfortunately, there is no weather station located in the mountainous region of
394 Tröllaskagi area and as previously mentioned there is no weather station in Fljótin area.
395 This means that we cannot report absolute precipitation data for the Móafellshyrna site,
396 but it is reasonable to assume that on a month by month basis the trends should be
397 similar to those of surrounding weather stations. Predicting whether precipitation falls as
398 snow or rain at Móafellshyrna is outside the scope of this study and is complicated by a

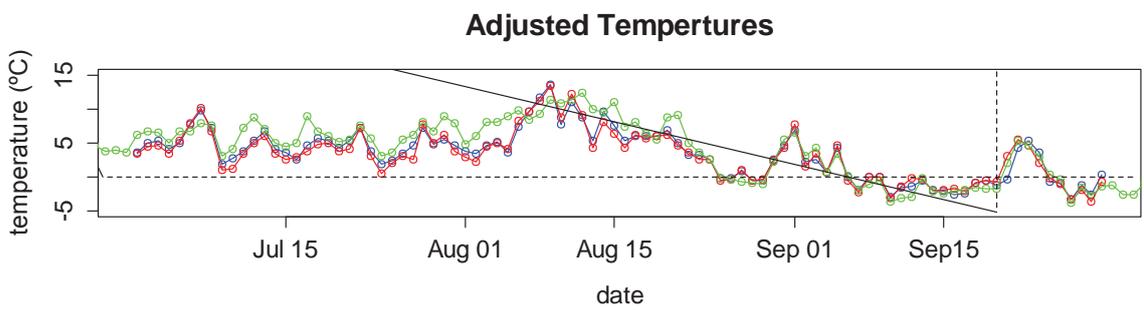
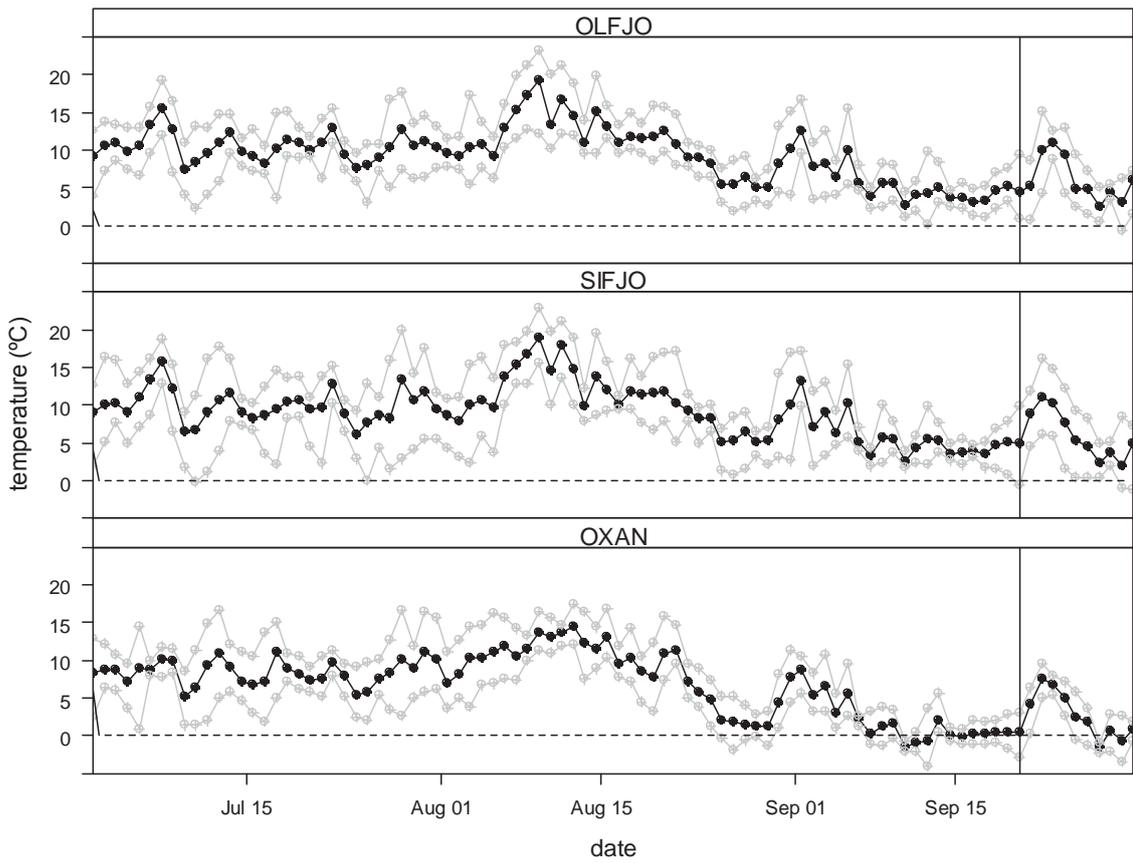
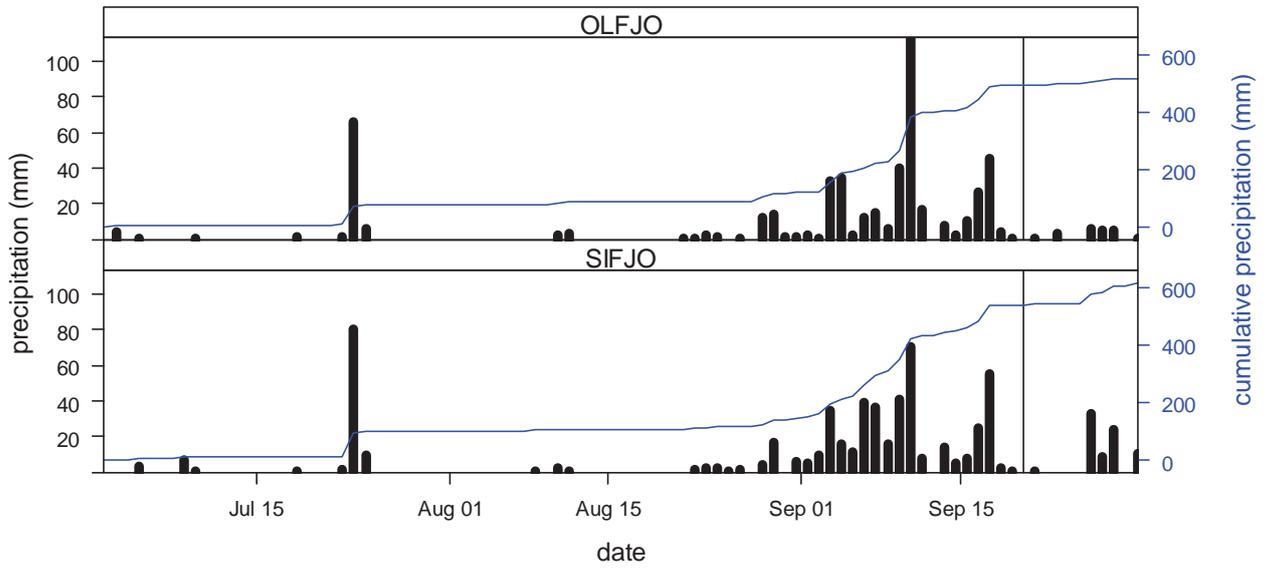
399 number of factors including snow drifting, wind-dependant snowmelt and variable
400 orographic effects dependant on wind direction (Brynjólfsson and Ólafsson 2008) – see
401 Section 1.3. Snow was visible on the ground on the day of the slide and was less than
402 1m thick from eye witness accounts (equivalent to ~100 mm of precipitation, depending
403 on snow density) Therefore, the majority of the precipitation received up to 20th
404 September had been absorbed by the ground. The fact that snow was present on the
405 ground argues against a sudden influx of water into the ground via snowmelt, known to
406 trigger other mass wasting phenomena in Iceland (e.g. Decaulne et al. 2005). Hence, we
407 did not pursue an analysis of the wind data from the weather stations, because this would
408 only be important if melt or precipitation were the primary triggers for the slide.
409



410
 411
 412 Figure 5 – Matrix plots of the difference between the average monthly temperature (top)
 413 and precipitation (bottom) and the average value for that month for the period 2000-2012
 414 for the Ólafsfjörður station (Data supplied by the IMO in 2016).



415
 416 Figure 6 – A boxplot of the precipitation data from the Ólafsfjörður station for each month between the
 417 years 2000 to 2012 (Data from IMO 2016). The end of the dotted lines is where the max and min values of
 418 precipitation were measured for each month, excluding the outliers that are displayed as dots. The blue
 419 boxes is where the 50% of accumulated measured precipitation falls and the black lines are the medians of
 420 each month.



423 Figure 7- Top: daily precipitation (black bars), cumulative precipitation (blue lines) for temperature
424 measurements from the Siglufjörður (SIFJO) and Ólafsfjörður (OLFJO) weather stations. Middle: daily
425 temperature data from the Siglufjörður (SIFJO), Ólafsfjörður (OLFJO) and Öxnadalsheiði (OXAN) weather
426 stations. Bottom: mean temperature data for all three stations adjusted to 880 m altitude of Móafellshyrna
427 (red= SIFJO, green= OXAN and blue= OLFJO). All data from 1st July to 30th September 2012 (Data
428 obtained from IMO in 2016).

429

430

431 4.3.2 Temperature

432 The temperature patterns in 2012 were also unusual, the summer and spring were
433 unusually warm on the whole, but the autumn was particularly cold (Fig. 5; Jónsson
434 2013). The average temperature measured in the town of Ólafsfjörður was in 10.4°C July
435 and 10.9°C in August in 2012. The average temperatures for these months for the period
436 2000-2012 are 9.8°C and 9.6°C respectively. The average temperature for September
437 2012 was on the other hand 5.7°C compared to the average temperature of 7.2°C for
438 2000-2012.

439 The average daily air temperature from 6th to 20th September at the Siglufjörður and
440 Ólafsfjörður stations ranged from 3 to 6°C and at the Öxnadalsheiði weather station
441 fluctuated around zero, but lowered to approximately -3°C the night before the slide (Fig.
442 7). Our corrected temperature data indicate average daily temperatures of at the altitude
443 of the Móafellshyrna slide of around -1 to -2°C in the days preceding the slide and hence
444 night-time temperatures would have been even lower. During the evening of 19th
445 September, a drop below 0°C in the atmospheric temperature in the mountains was
446 measured in the Öxnadalsheiði weather station (Fig. 7). The snow on the ground at the
447 time of the Móafellshyrna debris slide shows that similar sub-freezing conditions also
448 prevailed at this altitude prior to the slide. These low temperatures combined with the
449 snow cover are strong evidence that sudden influx of water from precipitation was not the
450 trigger for the Móafellshyrna slide.

451 4.3.3 Earthquake sequence

452 In the Eyjafjarðaráll graben, three earthquakes with magnitudes M 4.2 to M 4.5 were
453 registered on 19th and 20th September (Gudmundsson *et al.*, 2014). Their epicentres are
454 located 25-27 km north-northeast of the town of Siglufjörður, and 60 km north-northeast
455 of the Móafellshyrna site (Fig. 4). On the morning of 19th September, one day prior to the

456 slide, two earthquakes with magnitudes M 4.5 and M 4.3 occurred at 07:57 and 08:28
457 respectively. A number of smaller aftershocks continued to occur throughout the day
458 following these earthquakes. Another earthquake of magnitude M 4.2, occurred at 9:27
459 on the 20th September, approximately three hours prior to the first observations of the
460 Móafellshyrna slide (Fig. 4).

461

462 **5. Discussion**

463 The Móafellshyrna debris slide is a rare example of a gravitational mass movement
464 where three triggering factors may have contributed to the failure of the slope: heavy
465 precipitation, earthquake activity and ground ice degradation (via rising average surface
466 temperatures). We infer that precipitation was the main preparatory factor for this debris
467 slide for the following reasons. Heavy prolonged precipitation was recorded across the
468 area, where nearly half of the usual annual precipitation fell in less than one month (Fig.
469 5). Many case studies have shown that high magnitude water input, either by rainfall
470 (Rapp and Nyberg 1981) or snowmelt (Decaulne *et al.* 2007), leads to oversaturation of
471 soil directly triggering debris flows and shallow landslides. These studies also point out
472 the role of intense rainfall as a preparatory (rather than direct trigger) factor to failure (e.g.
473 Rapp, 1964, 1985 and 1995; Johnson and Rahn, 1970; Johnson and Rodine, 1984;
474 Rapp and Nyberg, 1981; Addisson, 1987; Innes, 1989; Luckman, 1992; Becht, 1995;
475 Cannon and Reneau, 2000; Sæmundsson *et al.*, 2003, Decaulne *et al.* 2007,
476 Sæmundsson and Decaulne 2007, Guzzetti *et al.* 2007). However, the role of water
477 infiltration in triggering of shallow landslides and debris flow in permafrost areas is rarely
478 well documented (Harris and Gustafsson, 1993). We do not favour sudden water input as
479 a direct trigger of the Móafellshyrna debris slide, because snow was present on the
480 ground at the time of the failure. However, we think this is a necessary pre-condition for
481 the failure to occur.

482 The direct influence of the seismic activity and associated ground acceleration on the
483 motion of the debris in the source area seems unlikely, as the event did not occur
484 immediately following any seismic event. Earthquakes are a common triggering factor for
485 landslide activity and are considered as a major cause for landslides worldwide (e.g.
486 Keefer, 1984, 1994, 2002, Malamund *et al.* 2004). Yet, no other debris flow or rockfall
487 activity was reported on the northern part of the Tröllaskagi area on 19th or 20th

488 September, which might be expected if ground acceleration were sufficient to trigger
489 mass movement. However, the short time interval (3hrs) between the last earthquake
490 and the failure indicates a possible connection between the slide and the seismic events.

491 Selby (1993) argued that “stability of the slope, orientation of the earthquake in relation to
492 the slide mass, earthquake magnitude, focal depth, seismic attenuation and after-shock
493 distribution” are factors that determine whether earthquakes trigger landslides. According
494 to Keefer (1984), the maximum area likely to be affected by landslides in a seismic event
495 increases from approximately 0 km² at M=4.0 up to 500.000 km² at M=9.2. According to
496 Malamund *et al.* (2004) the lowest earthquake-magnitude is M 4.3 +/-0.4. for triggering
497 gravitational mass movements. Tatard *et al.* (2010) state that earthquakes of M4 and
498 lower have little or no influence on landslide triggering. Nevertheless, several studies
499 (e.g., Sassa *et al.*, 2007, Walter and Joswig, 2008) mention that small earthquakes
500 (maximum M 3.6 in southern Italy according to Del Gaudio *et al.*, 2000) and repeated
501 shocks can influence hydrogeological settings and can possibly cause landslides,
502 sometimes with delay between the earthquake and the mass movement. Jibson *et al.*
503 (1994) also discuss delayed landslide movements, from larger earthquakes (M 7.0), and
504 state that the simplest explanation for the delay is a change in the ground-water
505 conditions. Based on the above mentioned studies, it is unlikely that an earthquake of M
506 4.3 was the only triggering factor for the Móafellshyrna debris slide, having taken place
507 60 km away from the epicentre. On the other hand, since the Móafellshyrna slide
508 occurred only three hours after a seismic event, the seismic sequence is likely playing
509 some indirect role.

510 The ground water flow system of the colluvial cone composing the source material of the
511 Móafellshyrna landslide is expected to be very limited. This is due to several factors: i)
512 the catchment area above the source area is not very large (around 350 m long); ii) the
513 colluvial cone is confined uphill by a vertical rockwall, and downhill by the edge of the
514 topographic bench; iii) the presence of ground ice cementing the deposits; iv) the sub-
515 horizontal dipping of the bedrock layers where the deposits are perched. However, it has
516 been shown that talus slopes can contain multiple and distinct groundwater flow systems
517 beneath or within them, and that they have a rapid and localised response to precipitation
518 and melt inputs (Roy and Hayashi, 2009; McClymont *et al.*, 2010). One component of the
519 groundwater flow in the colluvial cone of Móafellshyrna may originate in the pervasive
520 system of sub-horizontal and sub-vertical discontinuities affecting the bedrock. If a

521 groundwater system was present before the failure, the response of the water table
522 should be rapid. Since seismic activity can release water by coseismic liquefaction or
523 consolidation of loose sediments (e.g., Manga *et al.*, 2003; Montgomery *et al.*, 2003), a
524 change in the hydrogeological equilibrium of the colluvial cone caused by seismic activity
525 could have contributed to the occurrence of the failure.

526 Field evidence from the Móafellshyrna debris slide strongly suggests the involvement of
527 ground ice thaw in triggering the event. Because the month of the event had
528 temperatures lower than average and the days prior to the event were mostly below zero
529 Celsius (as evidenced by snow on the ground), we do not think thaw water from the
530 ground ice in the perched talus contributed significantly to the event. However, longer
531 term, deeper thawing caused by an annual rise in temperature and therefore a shift in the
532 permafrost table, including an anomalously warm preceding summer, we believe is a
533 more likely contributor. In recent years, there has been an increasing interest worldwide
534 in the influence of climate warming and associated decline of mountain permafrost on the
535 occurrence of mass wasting phenomena (e.g., Rebetz *et al.*, 1997; Gruber *et al.* 2004;
536 Gruber and Haeberli, 2007; Fischer *et al.*, 2006; Sattler *et al.*, 2011; Stoffel and Huggel,
537 2012; Damm and Felderer, 2013; Stoffel *et al.*, 2014). The increasing frequency of rapid
538 mass movements, such as debris flows, debris slides, rock falls and rock avalanches, in
539 mountainous areas have been linked in several cases with mountain permafrost
540 degradation (e.g., Clague *et al.*, 2012; Wirz *et al.*, 2013; Barboux *et al.*, 2015; Darrow *et*
541 *al.*, 2015; Haeberli *et al.*, 2016). Loss of ice-cementation, the presence of segregated ice,
542 increased hydrostatic pressure and the associated reduction of shear strength can all
543 lead to reduction of stability with increasing atmospheric temperature via permafrost
544 degradation (e.g., Gruber and Haeberli, 2007; Krautblatter *et al.*, 2013; Pogliotti *et al.*,
545 2015). Although, these previous studies have focused on the stability of massive rock
546 masses, a similar (perhaps exaggerated) effect might be expected in ice cemented talus.

547 The increase of mean annual temperature, which has been observed in Iceland over the
548 last few decades (Björnsson *et al.* 2008), should be leading to degradation of
549 discontinuous permafrost in Iceland, which is thought to be present in the Tröllaskagi
550 peninsula (Etzelmüller *et al.*, 2007a). Our observations of the ice-cemented deposits
551 shows that the Móafellshyrna slide originated from deposits containing pore-filling ground
552 ice and equally that these deposits were still frozen at the time of the slide. Together
553 these argue for a permafrost origin for this ground ice. The increasing average

554 temperatures over the last decades (Björnsson *et al.* 2008, Jónsson 2013) before the
555 event may have initiated the degradation of ground ice in the talus cone where the slide
556 initiated, but not from the top-down, but from the base-up. This thawing may have: i)
557 lubricated of the base of the cemented colluvial cone, ii) lowered the effective friction
558 angle (cohesion), and hence iii) caused the slow movement of the colluvial cone perched
559 on the bench and the sliding along a detachment surface of the whole ice-cemented
560 mass. The warming of the rock mass onto which the colluvial deposit was previously
561 cemented could have been brought about by a combination of propagation of the thermal
562 wave through the rock mass from the warmer southeast-facing side (e.g., Noetzli *et al.*
563 2007) and the delivery of warmer liquid water (derived from the intense precipitation) to
564 the talus-rock interface from the south-westward dipping strata. Hence, the rupture
565 occurred beneath the permafrost table. Perhaps the ice-cemented colluvium was in effect
566 forming an underground “ice dam” that was holding back water saturated debris until its
567 own weight and the seismic shaking caused it to fail. However, we cannot substantiate
568 this link with certainty as we lack direct temperature measurements in the talus cone. Our
569 hypothesis is supported by the slow widening of the tension crack at the top of the source
570 area as observed by the eye-witnesses and the fact that the landslide was fluidised (a
571 water content higher than expected for such a small catchment) even as it fell down the
572 talus slope (causing the channelized erosion).

573 Ice-cemented deposits have been observed in two other landslides in Iceland, e.g. on the
574 Torfufell mountain (source area at ~750 m a.s.l.) and the Árnesfjall Mountain (source
575 area at ~350 m a.s.l.). These further events provide additional evidence to support our
576 hypothesis that the lower limit of permafrost degradation extends to lower altitudes. The
577 source zones for the Móafellshyrna and the Torfufell slides are at the lower elevation limit
578 of discontinuous mountain permafrost in northern Iceland (i.e., 840 m a.s.l.; Etzelmüller *et*
579 *al.*, 2007a). On the other hand, the source zone of the slide in the Árnesfjall Mountain is
580 at an unexpected much lower elevation, which shows that the knowledge of mountain
581 permafrost in Iceland is incomplete. The setting of talus perched on benches is not a rare
582 situation in Iceland because of the sub-horizontal basalt layers create topographic
583 benches on which loose material can accumulate. Hence investigating whether those
584 with permafrost conditions, particularly above inhabited areas, contain ground ice and
585 establishing its condition, should be a priority.

586 **6. Conclusions**

587 The debris slide in the Móafellshyrna Mountain began with a slow movement of a
588 perched colluvial cone, as described by the local residents of Þrasastaðir. This colluvium
589 is composed of stratified ice-cemented deposits at 840 m elevation and large blocks and
590 boulders broke off the frontal part of the cone and fell onto the talus slope below. The
591 mass movement transformed into a rapid debris slide, travelling down the mountainside
592 with the final deposits coming to rest at 330 m a.s.l. The total volume of the slide is
593 estimated to be around 312 000 – 480 000 m³ (including the initial mass and mass added
594 via bulking), covering an area of 0.3 km².

595 We suggest that heavy precipitation prior to the slide was the main preparatory factor,
596 with over 400 mm of precipitation recorded in one month prior to the event after an
597 unusually dry summer season. The influence of seismic activity is unclear, but the close
598 temporal association between the last earthquake series and the failure suggests that the
599 shaking could have pre-conditioned the landslide weakening the cohesion between the
600 ice-cemented colluvium and the bedrock and/or changing of the hydrology. The presence
601 of ice-cementing the source colluvium at 880 m confirms the presence of discontinuous
602 mountain permafrost at that elevation. We suggest that the partial thaw of these deposits
603 was a trigger for the failure for three reasons: i) the landslide followed an usually warm
604 spring and summer, ii) mean annual air temperatures are generally increasing in Iceland
605 and iii) the colluvial cone initially slid as a single cohesive mass suggesting basal
606 lubrication/melting. The fact that two other recent landslides contain similar ice cemented
607 deposits suggests that mountain permafrost degradation could be more prevalent in
608 triggering landslides in Iceland than has previously been thought.

609 The ice-cemented deposits within the slides of the Móafellshyrna, Torfufell and Árnesfjall
610 Mountains have highlighted the need for a more detailed understanding of the distribution
611 and condition of mountain permafrost within perched talus deposits in Iceland. Future
612 studies should focus on the relationship between rapid mass wasting processes and the
613 degradation of mountain permafrost in such deposits in Iceland. These three landslides
614 occurred in uninhabited areas, but future similar landslides might not, and therefore they
615 could pose a potential hazard to society and infrastructure in the island.

616

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628

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