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# 1 Mass-Movements in Cold and Polar Climates

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5

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## 14 **Abstract**

15       Mass-movements in cold and Polar climates often involve permafrost and ice and  
16 hence these two phenomena are recurring themes throughout this chapter. We  
17 consider specifically mass movements in glacial, periglacial and paraglacial settings.  
18 For each environmental setting we describe the types of mass movement that have  
19 been documented in the literature, describing the current research foci and remaining  
20 challenges. We specifically highlight that mass-movements in cold and Polar climates  
21 are particularly sensitive to climate change, because of the involvement of permafrost  
22 and ice in their triggering and dynamics and therefore we may anticipate an increasing  
23 frequency and increased hazard from these phenomena in the future in response to  
24 unprecedented rates of environmental change.

25

26 **Keywords:** Rapid mass movements, active-layer-detachment slides, retrogressive-  
27 thaw slumps, debris flows, rock/debris falls, rock/debris slides, rock/debris avalanches,  
28 deep seated gravitational slope deformation (DSGDS), slush flows, ice-flows, snow-  
29 avalanches, permafrost, glacial environments, periglacial environments, paraglacial  
30 environments,

# 31 1. Introduction

32 Mass movement refers to the processes of downslope movement of rock, debris  
33 and soil under the influence of gravity (Crozier, 2002; Hutchinson, 1968). Mass  
34 wasting, mass movement, slope movement or slope failure are commonly used as  
35 synonyms. Another commonly used term for discrete slope movements is “landslides”  
36 (Crozier, 1989), and these movements are set apart from processes of subsidence  
37 and creep that usually lack discrete failure boundaries. Mass movements are generally  
38 divided into slow and rapid mass movements. Slow mass movements evolve over  
39 extended periods of time from months to decades or longer, such as deep-seated  
40 gravitational slope deformations (e.g., Dramis and Sorriso-Valvo, 1995). They also  
41 include phenomena like solifluction, a process that originates from the action of freeze-  
42 thaw cycles that induce downslope displacement in general at a rate of at most  
43 1 m/year of soils in cold environments, where vegetation is lacking or sparse (e.g.,  
44 Andersson, 1906; Ballantyne and Harris, 1994; Washburn, 1980). In contrast, rapid  
45 mass movements usually refer to high-magnitude, low-frequency processes that can  
46 be extremely fast moving (up to the scale of 5 m/s; Hungr, 2007).

47 Mass movements are classified in a number of ways, for example, according to the  
48 type of material, the type of movement, the volume of displaced material, and the  
49 extent of the material’s spreading area (Cruden and Varnes, 1996; Hungr et al., 2014;  
50 Varnes, 1978). Varnes (1978) developed a classification system for slope movements,  
51 based on five possible types of movement (fall, topple, slide, spread and flow) and on  
52 the type of material (rock, debris and earth). In Varnes' (1978) classification, slow  
53 processes such as subsidence, creeping and solifluction are not accounted for, and  
54 are considered by other authors (Sidle and Ochiai, 2006). Here, we adopt the updated  
55 Varnes landslide classification from Hungr et al. (2014), where the most significant

56 changes from the initial Varnes' version are: i) the use of textural classes to replace  
57 the term "earth", which has no standard definition in either geological or geotechnical  
58 material description schemes; ii) the reintroduction of "ice" as type of material, because  
59 many destructive mass movements on mountain slopes contain varying proportions of  
60 ice; iii) the introduction of slow mass movements under the movement type of "slope  
61 deformation".

62 This chapter will specifically describe and discuss mass movements in cold and  
63 polar environments. We divide this chapter into three sections which consider the  
64 following geomorphological and environmental settings: glacial, periglacial and  
65 paraglacial. Glacial environments are those where the accumulation of ice and snow  
66 is favoured to such an extent that bodies of ice accumulate on hillslopes or even flat  
67 ground and then flow *en masse* (glaciers or ice caps).

68 Periglacial environments are those where 1) the process of ground-freezing and  
69 thawing is dominant (Tricart, 1956); 2) permafrost - defined as soil or rock with a  
70 temperature below 0°C continuously for >1 year (Dobinski, 2011; French, 2007; Muller,  
71 1945; Péwé and Brown, 1973; Subcommittee, 1988; Washburn, 1980) - is present.

72 Paraglacial environments are non-glacial and landforms in this type of environments  
73 are formed during and after the deglaciation (Slaymaker, 2004; Church and Ryder,  
74 1972). The paraglacial processes in non-glacial environments are though directly  
75 influenced by glaciation (Church and Ryder 1972).

76 In each section we will describe the types of mass movement that have been  
77 documented in each environment, describing their general geomorphic  
78 characteristics, behaviour/dynamics, and triggering factors. We will highlight  
79 remaining open research questions, or key areas of active research. Finally we discuss  
80 the role of climate change in modulating the hazard posed by the mass movements in  
81 these three environments.

## 82 2. Rapid Mass Movements in glacial environments

83 Mountainous cold and polar environments form steep-slope environments  
84 conducive to mass failures and to the presence of glaciers, hence it is not surprising  
85 that the two phenomena interact. There are four broad configurations of this  
86 interaction:

- 87 • Landslides can fall on top of glaciers without entraining a significant portion  
88 of the glacier itself so called “supra-glacial landslides”.
- 89 • Landslides can fall on top of glaciers and in so doing entrain a significant  
90 portion of the glacier ice, resulting in rock-ice avalanches.
- 91 • Glaciers themselves can fail catastrophically resulting in ice avalanches.  
92 Such ice avalanches can themselves entrain debris resulting in rock-ice  
93 avalanches.
- 94 • Glaciers can advance over and remobilise landside deposits (e.g., **Burki et**  
95 **al., 2009**; Cook et al., 2013). Although glaciers could be considered as a type  
96 of mass movement, they are specifically discussed in the “Cryospheric  
97 Geomorphology” volume of this encyclopedia, and hence this remobilisation  
98 will not be detailed further in this chapter. However, this type of reworking is  
99 important in the context of supraglacial landslide deposits and will be briefly  
100 mentioned in that context.

101 In the following sections we will discuss first supra-glacial landslides, then ice and  
102 rock-ice avalanches.

103

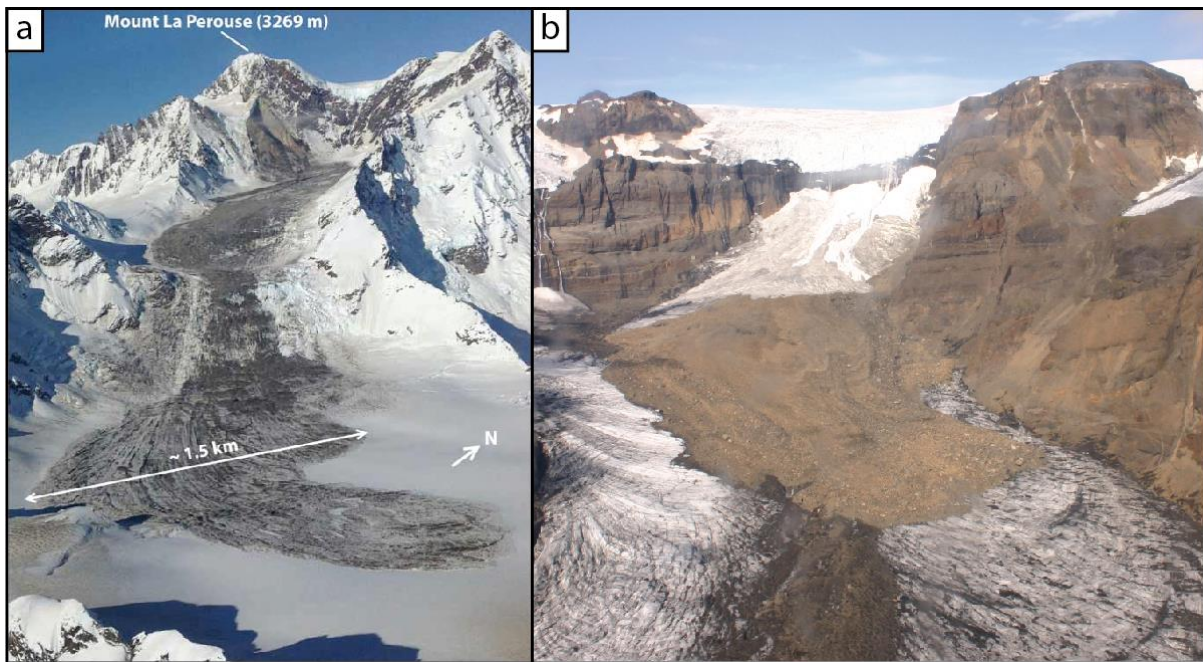
### 104 2.1 Supra-glacial landslides

105 Spectacular examples of large-scale rock avalanches falling onto glaciers are  
106 documented in the literature, mainly from the United States (Post 1965, Shreve 1966,

107 Dufresne et al. 2019, Crandell & Fahnestock 1965), the Himalayas (Hewitt 1988,  
108 2009a) and the New Zealand (Chinn et al. 1992). The main characteristic of such rock  
109 avalanches is the exceptionally long distances reached by the deposits over the ice  
110 surface. Several large-scale mass movements which have fallen onto glaciers during  
111 recent decades have been described and an international dataset with fall height,  
112 runout and volume has been developed (e.g. Schneider et al. 2011, Deline et al.  
113 2015a).

114 Landslides falling onto glaciers have deposits that differ from their counterparts that  
115 fall onto other terrain types due to the interaction with the ice. Most notably, their runout  
116 can be a lot greater than other landslides (e.g., Evans and Clague, 1994; Pirulli, 2009;  
117 Sosio et al., 2012). They tend to spread out, forming sheets of debris, which cover  
118 large areas (Figure 1; e.g., Delaney and Evans, 2014). Tongue-shaped, multi-lobed  
119 margins (Dufresne et al., 2019; Shreve, 1966; Shugar and Clague, 2011) and  
120 longitudinal furrows/ridges/bands (e.g., Dufresne and Davies, 2009) are other  
121 characteristics associated with these landslide deposits (Figure 1). A greater  
122 fragmentation of the rockmass is observed compared to non-glacial slides (Delaney  
123 and Evans, 2014). These observations are linked to the fact that the falling debris  
124 moves over a low-friction surface. Details on how the ice and snow incorporated in the  
125 landslide affect the physical processes which determine the landslide's motion are  
126 complex and debated (Pudasaini and Krautblatter, 2014; Schneider et al., 2011; Sosio  
127 et al., 2012). It should also be noted that not all mass wasting onto glaciers causes  
128 excessive runouts or distinctive deposits (e.g., Deline, 2009).

129 Similar to landslides in general, the triggers for such landslides are linked to a wide  
130 variety of factors, including seismic events (Shreve, 1966) and rockwall permafrost  
131 degradation (Coe et al., 2018), but of particular interest is the potential link with glacial  
132 retreat (see below).



134

135 *Figure 1: Examples of supraglacial rock avalanches. (a) On February 16, 2014*  
 136 *a rock avalanche detached from Mount La Perouse in Glacier Bay National Park,*  
 137 *Alaska, USA. Relief from head scarp to toe is about 1770 m. Photo taken on March 6,*  
 138 *2014 by Marten Geertsema. (b) The Morsárjökull rock avalanche in Iceland, which*  
 139 *occurred on 20<sup>th</sup> March 2007. Width of the deposit is ~350 m and the relief from head*  
 140 *scarp to toe is ~640 m. Photo taken in July 2007 by Matthew Roberts.*

141

142 Snowmelt is often one of the triggers of the ground-saturation that leads to  
 143 landslides (e.g., Moreiras et al., 2012), but the influence of the incorporation of snow  
 144 itself into the body of landslides is less-well studied. Yamasaki et al. (2014) found that  
 145 the incorporation and propagation over deep snow alone can increase the mobility of  
 146 a landslide. Dufresne et al. (2019) reported that that passive sliding on snow atop of a  
 147 glacier caused the Lamplugh rock avalanche to slide “silently”, or aseismically. This  
 148 mechanism has been inferred to be the origin of many “silent” supraglacial landslides



149 not detected by seismometers but instead in optical remote sensing (Smith et al.,  
150 2020).

151 The attempt to understand the dynamics of these landslides is partly driven by a  
152 need to understand the long runout of these landslides, as this can pose hazard to  
153 downslope communities and infrastructure, particularly if the landslide debris falls into  
154 the sea or terminal lake (e.g., Dufresne et al., 2018 a,b; Hubbard et al., 2005).

155 Understanding the magnitude-frequency of this type of landslide is important for  
156 understanding their role in landscape evolution at geological timescales, yet this is  
157 remarkably hard to quantify as deposits are usually unrecognisable in the landscape  
158 due to glacial reworking (e.g., Schleier et al., 2015). Debris that falls in the  
159 accumulation zone can be hidden by snowfall then rapidly sequestered into glaciers  
160 on the order of months (e.g., Dunning et al., 2015; Smith et al., 2020). Hence, deposits  
161 that fall on the ablation zone are transported down-glacier and contribute to the glacial  
162 deposits which are reworked beyond recognition at geological timescales. Some  
163 success has been obtained by monitoring seismic signals (e.g., Ekstrom and Stark,  
164 2013), but this only gives a snapshot of the situation at the present day. These  
165 landslides are on the continuum in our understanding of the role of mass movements  
166 in landscape evolution during glacial retreat (see Section 4 on mass movements in  
167 Paraglacial environments) and it is still an open question as to whether paraglacial or  
168 supraglacial mass movements play a greater role in shaping what we recognise today  
169 as the typical post-glacial landscape.

170 These landslides also play a role in conditioning the behaviour of glaciers and have  
171 therefore been studied by glaciologists in order to understand glacial response to  
172 changing climate and the sedimentary deposits (e.g., Reznichenko et al., 2011).  
173 Debris in the ablation zone can alter glacial dynamics by shutting off ablation (e.g.,  
174 D'Agata and Zanutta, 2007), or accelerating it. Debris can accelerate the glacier

175 motion (e.g., Shulmeister et al., 2009) and provoke surging behaviour (e.g., Hewitt,  
176 2009a). Equally, thick isolating debris can decelerate, or stall glacial movement.  
177 Supraglacial debris inputs by landslides can confound attempts to use moraines as  
178 markers of glacial advances/retreats and therefore limit their use as climate markers.  
179 This is because they confuse the sedimentary record via the introduction of significant  
180 lags between temperature changes and recorded “events”. The dynamic interplay  
181 between debris and ice within the glacial system is an active area of research and the  
182 role of mass movements in this interplay is yet another piece to add to the puzzle. The  
183 key drivers motivating research in this field are: understanding glacial meltwater  
184 generation for its contribution to sea level change, hydroelectricity generation and as  
185 a resource for potable water. Generation of glacial meltwater and sediment can also  
186 generate hazardous process-linkages and pose significant risk to infrastructure and  
187 habitation.

188

## 189 **2.2 Ice, Rock-ice and ice-rock avalanches**

### 190 ***2.2.1 Snow, ice avalanches and slushflows***

191 Snow and ice avalanches are a common type of rapid mass movement in cold and  
192 polar environments and are a critical driver of alpine glacial accumulation (e.g., Price  
193 et al., 2019). They take many forms because they can incorporate materials with a  
194 wide range of material properties (powder snow, sintered snow, ice, liquid water, rocks,  
195 debris, mud...) in varying proportions and can include phase changes e.g. from ice to  
196 water. From a geomorphology point of view many of these mass movements are  
197 transient as they do not supply significant sediment volume. Avalanche debris cones  
198 and strewn boulders are landscape elements that indicate the dominance of  
199 avalanches in downslope sediment transport (Figure 2b,c; e.g., Decaulne and

200 Sæmundsson, 2010; Jomelli and Francou, 2000; Luckman, 1992). Research into snow  
201 avalanches is mainly driven by a desire to understand the risk and design engineering  
202 solutions to mitigate it.



203  
204 *Figure 2: Examples of slush flow and avalanche deposits. (a) Freshly deposited*  
205 *slushflow from the Mt. Bleikkollubólstindur above the farm Hof in SE, Iceland. Photo*  
206 *taken by Brynja Kristjánsdóttir 2021. (b) Aerial view of two chutes in the Hólsádalur*

207 valley, northern Iceland, which host frequent snow avalanches whose deposits are  
208 found on the opposite side of the stream in the valley. The viewpoint of panel c is noted  
209 on the figure. Credit Loftmyndir ehf. (c) Field view of avalanche deposits within which  
210 pebbles from the stream below are found. A person is standing atop the deposits for  
211 scale. Photo taken by Þorsteinn Sæmundsson 2016.

212

213 Slush flows are mass movements that behave like debris flows (described fully in  
214 Section 3), but contain a significant amount of snow or ice (e.g., André, 1990; Decaulne  
215 and Sæmundsson, 2006; Elder and Kattelmann, 1993; Gude and Scherer, 1995; Héту  
216 et al., 2017; Larocque et al., 2001; Nyberg, 1989; Pérez-Guillén et al., 2019). They are  
217 characterised by levees and lobate tongues (Figure 2a) and have a longer runout than  
218 dry avalanches. They are capable of transporting clasts up to metres in size, although  
219 some slush flows contain only snow and are usually called “slush avalanches” or “wet  
220 avalanches” (Figure 2a; e.g., Bartelt et al., 2012; Eckerstorfer and Christiansen, 2012).  
221 Slush flows are triggered in a similar way to debris flows, by saturation of the sliding  
222 mass by snowmelt or precipitation. There has not yet been any research that permits  
223 the distinction of slush flow and debris flow deposits in the sedimentary record  
224 meaning their impact on the landscape at geologic timescales is broadly unknown.  
225 Slush flows are generally studied alongside snow avalanches or debris flows and form  
226 part of a spectrum of processes that can pose hazard to populations and infrastructure  
227 in steep mountainous environments.

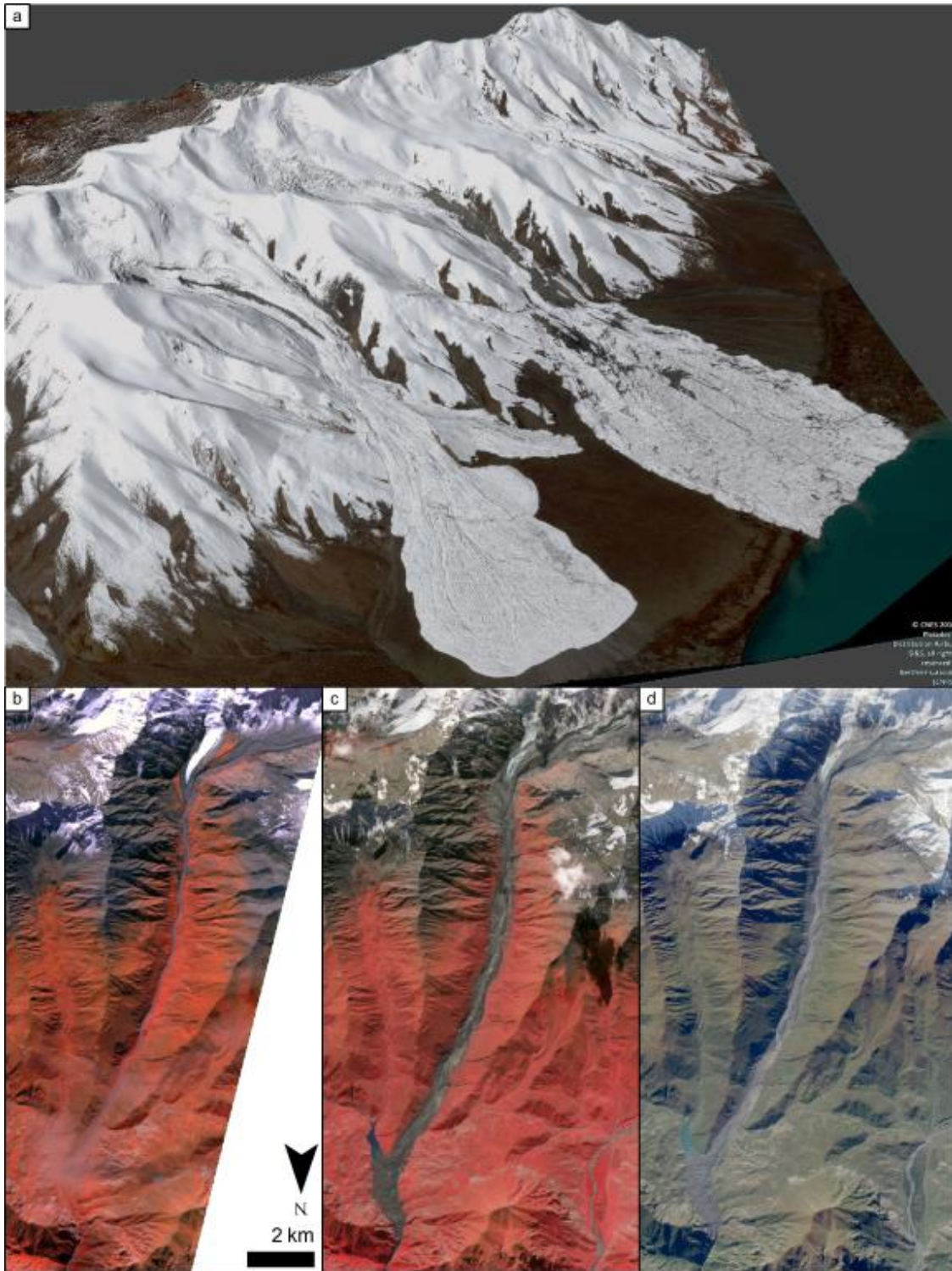
228

### 229 ***2.2.3 Ice, ice-rock and rock-ice avalanches***

230 Mass movements in glacial areas can originate from the glaciers themselves, the  
231 hillslopes, or both. Once initiated these mass movements can further interact with

232 downslope debris, snow, ice or rock causing complex events that can be hard to  
233 classify uniquely (e.g., Petrakov et al., 2008).

234       There have been three recent well-reported cases where large portions of glaciers  
235 detached to form rapid mass movements (Evans et al., 2009a; Falaschi et al., 2019;  
236 Gilbert et al., 2018; Kääb et al., 2018; Tian et al., 2017). For example, the collapse of  
237 two nearby glaciers in the Aru range in Tibet was widely reported in the media because  
238 of the remarkable visibility in satellite images (Figure 3a). However, these mass  
239 movements leave very little geomorphological evidence in the landscape, because  
240 their main constituent is ice which is rapidly ablated in the relatively warm locations  
241 where the deposits are laid down. Hence, the frequency of such failures in the geologic  
242 record is unknown. The limitation to a handful of present-day cases means that,  
243 although these types of failure pose considerable risk, there are many open questions  
244 as to how they are triggered.



245

246 *Figure 3: (a) Pléiades 3D rendering of the two glacier collapses in the Aru*  
247 *Mountains, Tibet with kind permission of Etienne Berthier at the CNRS. The fan in the*  
248 *foreground is ~2 km across. Image credit CNES Pléiades, Airbus Defence and Space.*

249 *(b, c and d) Mt. Kazbek in Southern Russia before and after a hanging glacier*  
250 *collapsed from the slope of Mt. Dzhimarai-Khokh onto the Kolka Glacier which*

251 *triggered an avalanche of ice and debris that went over the Maili Glacier terminus then*  
252 *slid more than 24 km burying villages in North Ossetia on 20<sup>th</sup> September 2002. (b, c)*  
253 *False-colour images from the Advanced Spaceborne Thermal Emission and*  
254 *Reflection Radiometer (ASTER) aboard NASA's Terra satellite. The deep reds show*  
255 *vegetated land surfaces, grey areas are bare rock, and white shows ice-cover. (d) True*  
256 *colour image from the Advanced Land Imager instrument on the EO-1 spacecraft.*  
257 *Images available from <https://earthobservatory.nasa.gov>*

258

259 Landscape change does occur where glaciers and rock detach together, or rock  
260 avalanches impact and incorporate glaciers or rock glaciers (Evans et al., 2009b;  
261 Haeberli et al., 2004; Schneider et al., 2011). Triggering of these large, complex events  
262 has been attributed to thermal perturbations from both climate and geothermal origins  
263 (Huggel, 2009; Huggel et al., 2012). These events occur on the flanks of volcanoes as  
264 well as in in uplifted mountain areas. They have no specific geomorphic signature and  
265 can display the attributes generally associated with rock avalanches (Figure 3b-d),  
266 debris avalanches, debris flows and supraglacial landslides, with their defining  
267 characteristic being their large mobility. The large-scale of these events and their  
268 mobility means they are both rare, but can be extremely hazardous (Evans et al.,  
269 2009b; Iribarren Anacona et al., 2015). The deposits of ancient rock-ice avalanches  
270 are reported in the literature, but only when the role of the ice is thought to have played  
271 a minimal effect on the resulting geomorphology (Robinson et al., 2014). It is unclear  
272 how important such large mass movements are in long-term landscape evolution,  
273 because their sedimentary signal is hard to detangle from other mass movements with  
274 which they share morphologic properties. The main research challenge lies in  
275 dissecting the dominant processes within the evolution of the mass movement to  
276 understand the mobility and potential hazard of such flows. Predicting where and when

277 such huge movements might occur remains the major challenge, as it does for rock  
278 avalanches in general.

279

### 280 **3. Mass Movements in periglacial environments**

281 Periglacial environments are sensitive to rapid changes in local climate, snow cover,  
282 topography, incoming radiation, ground and soil characteristics, which influence the  
283 variability of ground thermal regimes at local scale (e.g., Harris and Pedersen, 1998;  
284 Harris *et al.*, 2009; Thomas *et al.*, 2009; Hasler *et al.*, 2015). As average temperatures  
285 continue to rise, permafrost and permafrost ice are expected to degrade (Biskaborn *et*  
286 *al.*, 2019; Blunden and Arndt, 2020; Christiansen *et al.*, 2010; Gislås *et al.*, 2017;  
287 Harris *et al.*, 2009; Romanovsky *et al.*, 2010; Slater and Lawrence, 2013), particularly  
288 at high latitudes, where average yearly temperatures are increasing at faster rates than  
289 global averages (Alexeev and Jackson, 2013; Blunden and Arndt, 2020; Masson-  
290 Delmotte *et al.*, 2006; Screen *et al.*, 2012).

291 In the last decades, mass movements have been studied in relation to degrading  
292 permafrost conditions (Ashastina *et al.*, 2017; Blais-Stevens *et al.*, 2015; Fischer *et al.*,  
293 2006; Gruber and Haeberli, 2007; Haeberli *et al.*, 2017a; Huscroft *et al.*, 2003; Patton  
294 *et al.*, 2019a; Ravanel *et al.*, 2017). Mass movements are particularly common under  
295 periglacial conditions, as freeze-thaw weathering causes deterioration of rock,  
296 sediments and soils (e.g., Dobinski, 2011; Harris *et al.*, 2001). Frost action promotes  
297 mechanical fragmentation of rocks (Matsuoka and Murton, 2008; Prick, 2003), with  
298 resulting loose material available for transport (Van Vliet-Lanoë and Fox, 2018).  
299 Degrading-permafrost conditions reduce slope stability, as reaching  $-1.5^{\circ}\text{C}$  -  $0^{\circ}\text{C}$  or  
300 melting temperatures causes reduction in ground shear strength, loss of ice/rock  
301 interlocking, increase in the hydraulic permeability, and ice segregation (Davies *et al.*,



302 2001; Draebing et al., 2017; Draebing and Krautblatter, 2019; Gruber and Haeberli,  
303 2007; Krautblatter et al., 2013a; Matsuoka and Murton, 2008; Murton et al., 2006).

304 Different types of mass movements can occur in terrains affected by permafrost,  
305 which in high-latitude lowland permafrost regions is classified into zones based on its  
306 geographic extent and continuity. Permafrost is defined continuous where frozen  
307 ground is present at 90-100% localities, while it is defined discontinuous where units  
308 of frozen ground (50-90%) are separated by unfrozen areas. Sporadic and isolated  
309 permafrost is restricted to limited areas, covering respectively 10-50% and 0-10% of  
310 the terrain considered as permafrost (e.g., Black, 1954; French, 2007; Péwé, 1983).  
311 Permafrost is also present at lower latitudes at high elevation, it is called mountain  
312 permafrost, and it is characterised by an extreme spatial variability (Gruber and  
313 Haeberli, 2007; Haeberli et al., 2011; Péwé, 1983; Stearns, 1966). Mass movements  
314 in periglacial environments can be distinctive of a particular class of permafrost or  
315 ubiquitous in various permafrost terrains, and we will specifically discuss: active-layer-  
316 detachment slides, retrogressive-thaw slumps, debris flows, rock/debris falls,  
317 rock/debris slides, rock/debris avalanches.

318

### 319 **3.1 Active-layer-detachment slides**

320 Active-layer-detachment slides are localised slope failures (Figure 4) that are  
321 confined to the permafrost active layer in continuous and discontinuous ice-rich  
322 permafrost terrains (Leibman, 1995; Lewkowicz, 1990; Lewkowicz et al., 1992;  
323 Lewkowicz and Harris, 2005a,b). They fall in the suite of landforms resulting from the  
324 process of thawing of ice-rich permafrost or melting of massive ground ice called  
325 thermokarst (French, 2007; Kokelj and Jorgenson, 2013). They occur due to a  
326 reduction in shear strength resulting from high porewater pressures and low effective

327 strengths, due to rapid thawing of ice lenses at the base of the active layer or in the  
328 upper part of permafrost in fine-grained soils with low hydraulic conductivity (French,  
329 2007; Lamoureux and Lafrenière, 2009; Lewkowicz, 2007). The movements generally  
330 involve shallow (~1 m deep) failures over a thin shear zone or shear plane on very low-  
331 angled slopes (Kokelj and Jorgenson, 2013; Lewkowicz and Harris, 2005a). Few direct  
332 observations exist of in-action active-layer-detachment slides, and they develop over  
333 a period of minutes to several days by a single almost instantaneous sliding movement  
334 or by progressive movements of a few metres per hour, reaching runout distances of  
335 up to several hundred metres (Lewkowicz, 2007; Lewkowicz and Harris, 2005a).  
336 Active-layer-detachment slides usually occur in clusters (Balsler et al., 2009; Couture  
337 and Riopel, 2008; Lamoureux and Lafrenière, 2009), generally after warm summers,  
338 rapid spring thaw, high snowmelt, prolonged summer precipitation, or forest or tundra  
339 fires (Gooseff et al., 2009; Lamoureux and Lafrenière, 2009). Active-layer-detachment  
340 slides can enhance slope sediment and solute yields in stream systems (Lamoureux  
341 and Lafrenière, 2009; Lewkowicz, 2007; Lewkowicz and Kokelj, 2002), impact the  
342 patterns of tundra vegetation communities (e.g., Cannone et al., 2010; Frost et al.,  
343 2018; Verdonen et al., 2020) and mobilise carbon sequestered in shallow permafrost  
344 (e.g., Abbott et al., 2015; Beamish et al., 2014; Turetsky et al., 2019; Woods et al.,  
345 2011). An increase in active-layer-detachment slides has been observed also in the  
346 western Antarctic Peninsula region, where permafrost has reached degrading  
347 temperatures (– 0.4 to – 3.1 °C; because the area has been affected by the greatest  
348 degree of warming in the past 50 years (Bockheim et al., 2013; Guglielmin, 2012;  
349 Vieira et al., 2008). Research is particularly focussed on the role these mass  
350 movements have in releasing carbon contributing to the positive feedback mechanism  
351 of the warming climate releasing carbon trapped in continuous permafrost, which goes  
352 on to exacerbate future warming and so on. Although areas with continuous permafrost

353 are sparsely inhabited, these mass movements can pose risk to infrastructure (e.g.  
354 Behnia and Blias-Stevens 2018), but also to archaeological investigations (Walls et al.  
355 2020).



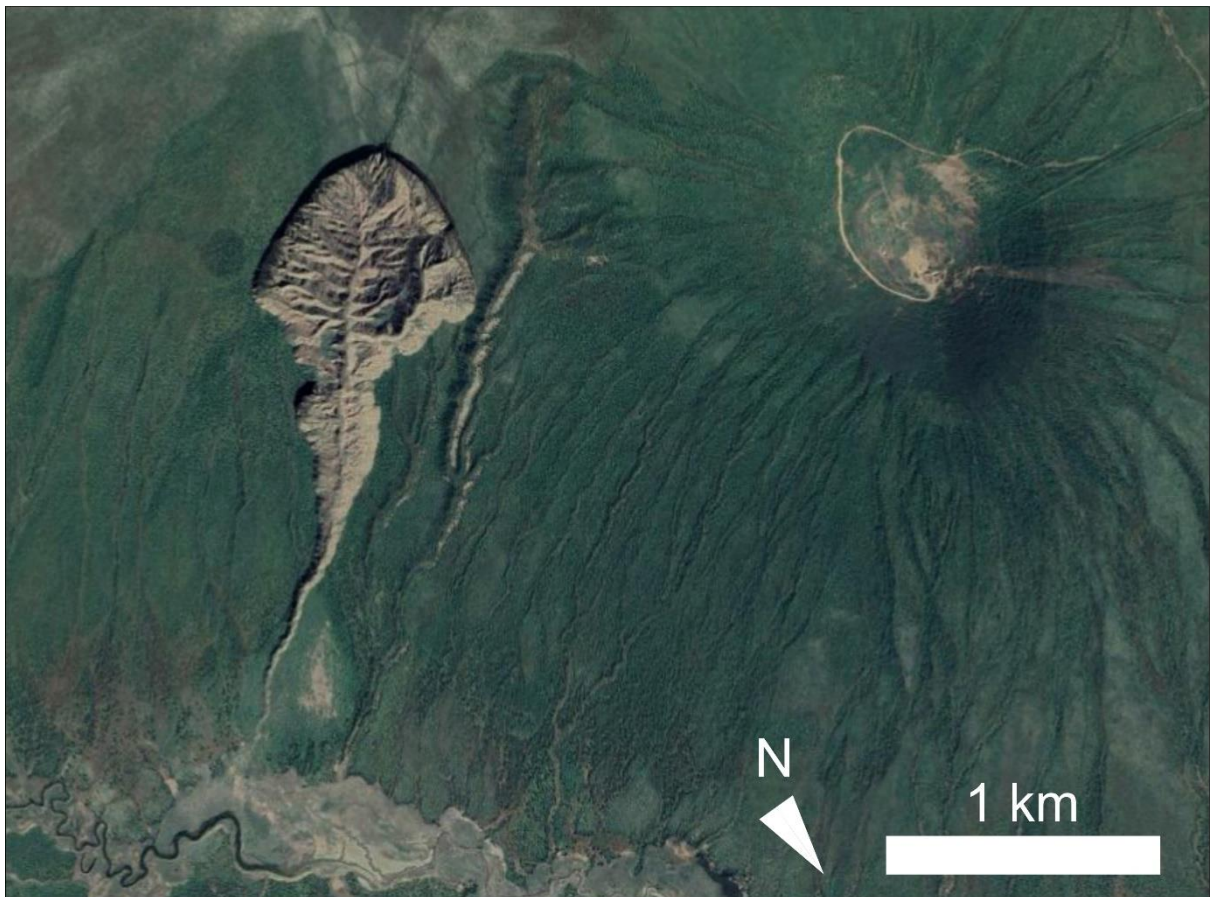
356  
357 *Figure 4 - Active-layer detachment slide (perimeter marked with white dashed line),*  
358 *Fosheim Peninsula, Ellesmere Island, Nunavut, Canada. Modified after Lewkowicz,*  
359 *2007.*

360

### 361 **3.2 Retrogressive-thaw slumps**

362 Retrogressive-thaw slumps are semi-circular depressions resulting from thawing of  
363 ice-rich permafrost terrains (Burn and Friele, 1989; French, 2007; Lantuit and Pollard,  
364 2008). They occur where ice-rich permafrost is exposed to thaw, and are generally  
365 triggered by wave action along coasts, undercutting erosion of banks along streams  
366 or lakes, or active-layer-detachment slides (Lantuit and Pollard, 2008; Patton et al.,  
367 2019b). Thaw slumps cannot be classified as “rapid movements” — they can develop  
368 over decades (Lacelle et al., 2010) — but phenomena like falls or flows can transport  
369 thawed material from the slump headwall as secondary movements (Lantuit et al.,  
370 2012; Murton, 2001). They can cover vast areas, from 0.05 up to 0.8 km<sup>2</sup>, as in the  
371 case of the Batagaika mega-thaw-slump (Figure 5), which in the last thirty years has  
372 increased its size by three times, reaching an area of 0.78 km<sup>2</sup> in 2018 (Vadakkedath  
373 et al., 2020). Thaw slumps expose ground ice to thaw, modify the landscape, and can  
374 erode and transport thawed material to lakes, valley-bottoms or coastal zones (Kokelj  
375 et al., 2015; Lafrenière and Lamoureux, 2019; Lantuit et al., 2012; Murton et al., 2017).

376 Retrogressive thaw slumps are showing an accelerated activity due to global warming  
377 in continuous and discontinuous permafrost terrains, from western Canadian Arctic to  
378 Siberia (e.g., Ashastina et al., 2017; Jones et al., 2019; Segal et al., 2016). Thousands  
379 of retrogressive thaw slumps have formed in the last 30 years due to extreme  
380 temperatures in High Arctic (Banks, Island, Canada), and are modelled to rise to over  
381 10,000 per decade by 2075 (Lewkowicz and Way, 2019). Hence, similarly to active-  
382 layer detachment slides current research focuses on the impact climate change is  
383 having on these mass movements.



384  
385 *Figure 5 - The Batagaika mega-thaw slump in the Chersky Range area, Sakha*  
386 *Republic, Russia. CNES/Airbus Image from Google Earth, 6<sup>th</sup> October 2019.*

387

### 388 3.3 Debris Flows

389 Debris flows are rapid, downslope flows of poorly sorted debris mixed with water  
390 (e.g., Iverson, 1997), and are common on the slopes of periglacial environments,  
391 including in the European Alps to Iceland (Figure 6), from the Tibetan Plateau to  
392 Antarctica (Decaulne and Sæmundsson, 2006b; Morino et al. 2019; Cannone et al.,  
393 2006; Oliva and Ruiz-Fernández, 2015; Sattler et al., 2011; Yu et al., 2020). Debris  
394 flows can travel over long distances and transport large amounts of material at great  
395 speed (Zimmermann and Haeberli, 1992). They are characterised by a central  
396 channel, the construction of levees on one or both sides of channel, and a terminal  
397 depositional lobe or lobes (e.g., Costa, 1984). Steep slopes, loose materials and wet  
398 conditions are favourable for the occurrence of debris flows (Lewin and Warburton,  
399 1994). This combination of factors is extremely common in periglacial mountain  
400 environments, where climatic warming during the 19th and 20th centuries has  
401 indirectly increased the frequency of debris flows (Decaulne and Sæmundsson,  
402 2006b; Haeberli et al., 1990; Rebetz et al., 1997; Zimmermann, 1990; Zimmermann  
403 and Haeberli, 1992). This increased frequency is not directly linked to the thaw of  
404 perennially frozen debris in the source areas (i.e., meltwater producing a source of  
405 fluid), but it has been shown that there is a link between the initiation zones and the  
406 thickening of the active layer as a reaction to the increase in atmospheric temperatures  
407 (Sattler et al., 2011). This has also been observed where the interface between frozen  
408 and unfrozen ground seems to control the depth of movement in debris flows (Huscroft  
409 et al., 2003). Active-layer failures may result from increased concentration of water  
410 above the permafrost table, generating debris flows and mass failures (Kummert et al.,  
411 2018; Lugon and Stoffel, 2010; Marcer et al., 2020), particularly in formerly frozen  
412 sediment mantled slopes (Davies et al., 2001; Fischer et al., 2013; Haeberli, 2013).  
413 Recently deglaciated terrains have an abundance of debris often perched in unstable

414 positions that are particularly susceptible to remobilisation by debris flows (Ballantyne  
415 2002).

416 Debris flows often pose hazard to human habitation and infrastructure and the main  
417 challenge lies in developing tools to effectively predict where and when they will be  
418 triggered in addition to understanding their runout. On one hand this requires, a deeper  
419 understanding of the physical mechanisms where significant effort is being put into  
420 focussed monitoring and developing sophisticated physics-based models. And on the  
421 other hand it also requires the development simple yet effective tools to enable public  
422 authorities and other stakeholders to mitigate the hazard posed by debris flows.

423



424

425 *Figure 6 - The debris flows above in the Gleid arhjalli slope above the town of*  
426 * safj r ur, Westfjords, Iceland (Decaulne and S mundsson, 2006b; Morino et al.*  
427 *2019).*

428

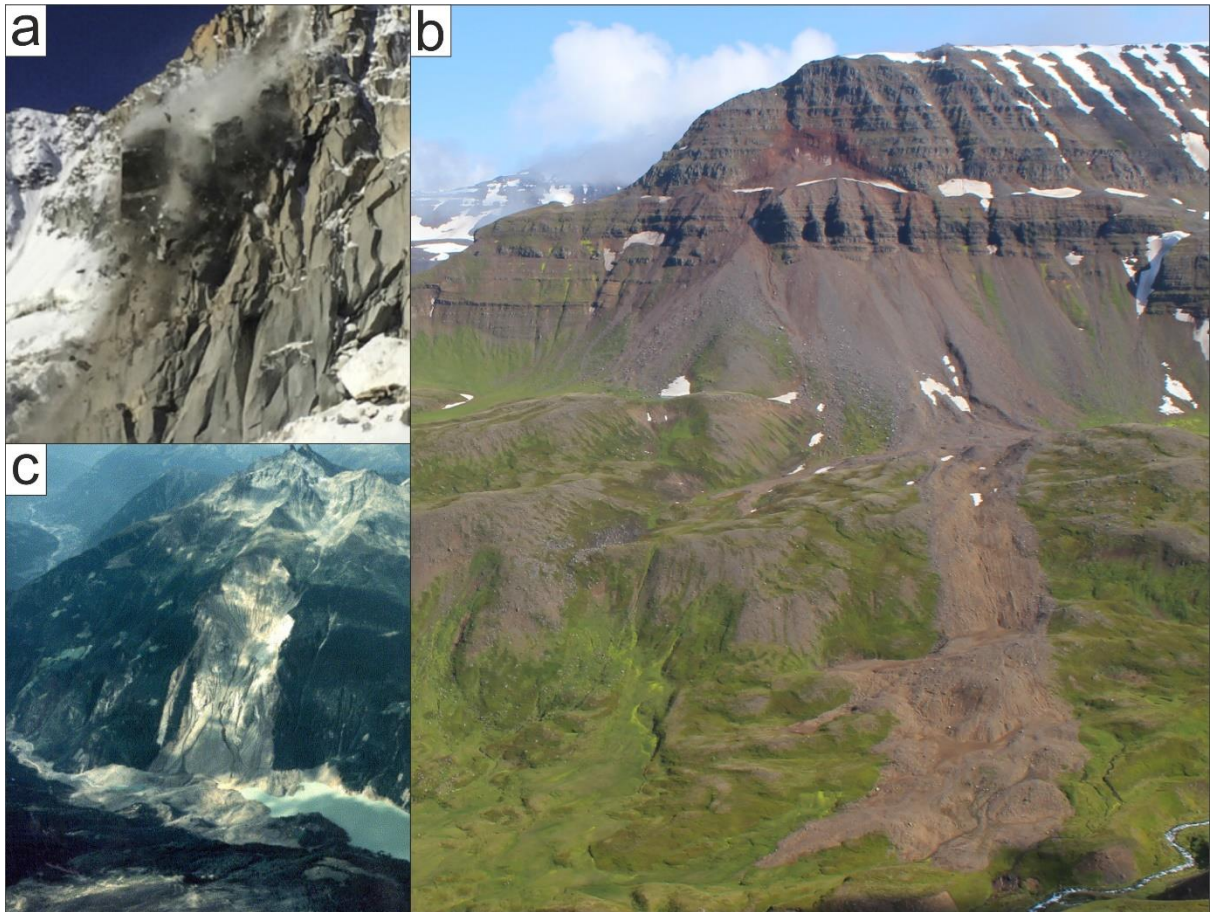
### 429 **3.4 Rock/debris falls**

430 Rock/debris falls are mass movements on steep slopes (Fig. 7a), rock/debris slides  
431 (Fig. 7b) are mass movements that occur along discrete surfaces, and rock/debris  
432 avalanches are large-volume ( $>1\text{M m}^3$ ) mass movements that cause fragmentation  
433 during rapid transport (Fig. 7c) (Cruden and Varnes, 1996; Hungr et al., 2014). These  
434 phenomena are frequent in mountain glacial and periglacial environments, usually in  
435 paraglacial settings where glacial over-deepening, stress-release fracturing after  
436 removal of glacial ice (Ballantyne, 2002; McColl and Davies, 2013), glacial  
437 debuttressing (Blikra et al., 2006; Cossart et al., 2014), degrading permafrost (Clague  
438 et al., 2012a; Deline et al., 2015a; Fischer et al., 2006; Gruber and Haeberli, 2007;  
439 Krautblatter et al., 2013b; Magnin et al., 2020; Morino et al., 2019; Sæmundsson et  
440 al., 2018), and cooling-warming cycles (Draebing, 2020) can all act to destabilise rock  
441 walls and debris source areas. These factors can induce changes in the stress field  
442 and expose previously insulated surfaces to altered mechanical and thermal erosion  
443 (Draebing et al., 2017; Haeberli, 1997; Jia et al., 2015; Krautblatter et al., 2013b;  
444 Wegmann et al., 1998), which can consequently favour mass wasting processes. The  
445 increase in temperatures in the last decades due to climate change has resulted in  
446 increased activity of such rock/debris falls, slides and avalanches; a correlation  
447 between exceptional heatwaves and rockfalls (Fig. 7a) has been established (Huggel  
448 et al., 2012; Ravanel et al., 2017; Ravanel and Deline, 2008). When permafrost  
449 degrades, physical changes also occur in the rock mass and in loose deposits (Allen  
450 and Huggel, 2013; Draebing et al., 2017; Gruber and Haeberli, 2007; Harris et al.,  
451 2009; Ravanel et al., 2017). After permafrost thaw, the shear strength of rock/debris  
452 significantly reduces with warming (minimum factor of safety is between  $-1.5^\circ\text{C}$  and

453 0°C) (Davies et al., 2001). Ice segregation and volume expansion subsequent to  
454 increased temperatures can lead to failures (Gruber and Haeberli, 2007). Elevated  
455 water pressure and reduced frictional strength can result from meltwater or ground-  
456 water flow in previously frozen masses (Harris, 2005). Ice-rich permafrost occurs in  
457 loose-deposits landforms such as rock glaciers, ice-cored moraines and talus slopes  
458 (Dramis et al., 1995; Isaksen et al., 2000; Kenner et al., 2017; Lukas et al., 2005;  
459 Monnier and Kinnard, 2015), and its degradation can cause a reduction in  
460 cementation, which can lead to the occurrence of falls, slides (Fig. 7b) or avalanches  
461 (Fig. 7c; Dramis et al., 1995). Rock and debris falls, slides and avalanches directly  
462 attributed to this mechanism have been reported from numerous locations in cold  
463 climates around the globe, including Alaska, Canada, European Alps, Greenland,  
464 Himalaya, New Zealand, Norway, Russia (Deline, 2009; Evans et al., 2009a,b,c;  
465 Fischer and Huggel, 2008; Frauenfelder et al., 2018; Gruber et al., 2017; Huggel et al.,  
466 2012, 2005; Magnin et al., 2019; Patton et al., 2019a; Sosio et al., 2008; Svennevig et  
467 al., 2019). Much current research is focussed on mitigating the risk from such events  
468 and is reliant on understanding and modelling the influence of increasing temperature  
469 on the state of the permafrost, the influence of the associated hydrology which finally  
470 leads to the destabilisation of rock mass.

471





472

473 *Figure 7 - Rapid mass movements caused by permafrost degradation. a. The Tour*  
474 *Ronde rock fall of 27<sup>th</sup> August 2015, Mont Blanc Massif, France (modified after*  
475 *Ravanel et al., 2017); b. The Móafellshyrna debris slide of 20<sup>th</sup> September 2012,*  
476 *Tröllaskagi peninsula, Iceland (Sæmundsson et al. 2018, Morino et al. 2019); c. The*  
477 *Val Pola rock avalanche of 28<sup>th</sup> July 1987, Central Alps, Italy (modified after Crosta et*  
478 *al., 2004).*

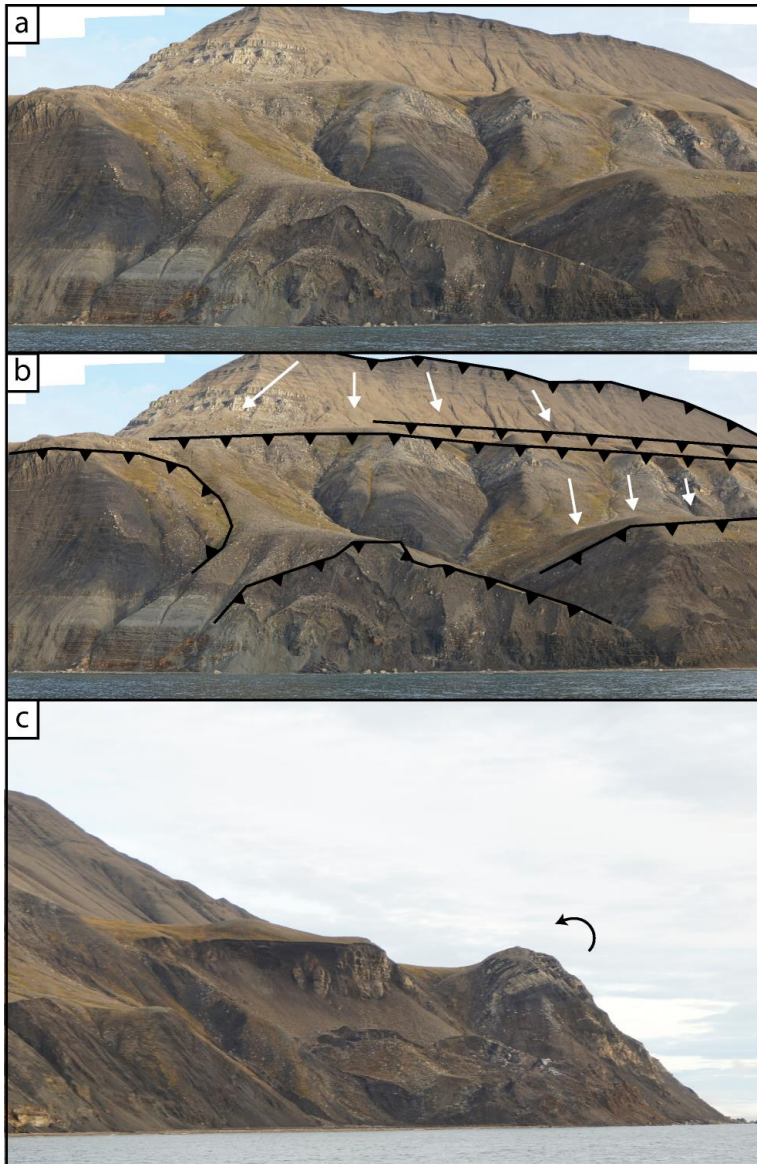
479

#### 480 **4. Rapid Mass Movements in Paraglacial Environments**

481 Paraglacial environments represent the relatively rapid adjustment from deglaciated  
482 landscape to non-glacial conditions (Ballantyne 2002). A wide range of subaerial  
483 processes are active in these environments which can be active over different  
484 timescales and lead to different rates of landscape change (evolution). Many of these

485 processes are intimately associated with periglacial environments, already described  
486 in Section 3. Periglacial conditions usually only persist for a relatively short period after  
487 deglaciation, and here we focus on (large scale) Mass Movements that are not  
488 necessarily related to periglacial environmental conditions, yet occur in a paraglacial  
489 setting.

490 Massive rock slope failures are as frequent or more frequent in glaciated basins  
491 than in equivalent ice-free mountainous terrain (McSaveney 2002, Geertsema et al.  
492 2006, Hewitt 2009a Deline et al. 2015a). Complex interactions between several factors  
493 can affect the occurrence of mass movements according to Pacione (1999); inherent  
494 factors (e.g. lithology, rock structure, slope form, topographic relief), preparatory  
495 factors (e.g. weathering, de-buttressing, climate change), triggering factors (e.g.  
496 earthquake, rainstorm) and factors that may affect mobility (e.g. glacial surface)  
497 (Agliardi et al. 2012, Deline et al. 2015b). According to Ballantyne (2002) the response  
498 of a rock slope to glacier down wasting has been reported to result in large rock  
499 avalanches, large-scale, progressive and slow rock mass deformation, and frequent  
500 rock fall events. Slope steepening caused by glacial erosion and unloading or de-  
501 buttressing due to glacial retreat often leads to rock slope failures (Agliardi et al. 2012,  
502 Holm et al. 2004, Lane et al., 2017, Stoffel & Huggel 2012). Thermal and mechanical  
503 alteration can result from cycles of ice loading and unloading and exposition of rock to  
504 new thermal regimes due to the glacier retreat (Cody et al., 2020; Coquin et al., 2015;  
505 Evans and Clague, 1994; Grämiger et al., 2017; McColl, 2012; McColl and Draebing,  
506 2019; Mercier et al., 2017; Wegmann et al., 1998). Finally, permafrost fluctuations  
507 throughout the Holocene is likely a first order control on past and modern rock slope  
508 failures (Hilger et al., 2021).



509

510 *Figure 8 A Deep seated gravitational slope deformation at Forkastningsfjellet in*  
 511 *Spitsbergen documented by Kuhn et al. (2019, 2021). (a) Unannotated view towards*  
 512 *the blocks that are sliding downslope. (b) Annotated view showing the location of the*  
 513 *main sliding surfaces that define the different blocks with white arrows indicating*  
 514 *displacement directions. (c) Side-on view of the same slope showing the rotation of*  
 515 *the block nearest the coast. Images kindly provided by Reginald Hermanns.*

516

517 Deep seated gravitational slope deformation (DSGSD, also called "sagging" or  
 518 "sackung", Figure 8) are one of the adjustment processes which can be detected in

519 paraglacial setting and can be categorized as mountain slope deformation of steep,  
520 high mountain slopes, manifested by scarps and cracks, but lacking a defined rupture  
521 surface and slow movement rates (Hungre et al., 2014). DSGSD are large mass  
522 movements with long duration slow motion processes that can experience irregular  
523 rates of activity characterized by periods of rapid reactivating (Pánek et al. 2013) which  
524 often result in high volume and long-runout events (Hewitt et al., 2008; Hewitt,  
525 2009a,b,c; Mege and Bourgeois, 2011; Pánek & Klimeš, 2016; Pedrazzini et al., 2013).  
526 The first observations of the geomorphic role and impact of DSGSD dates back to the  
527 1940s, but recent developments in remote sensing, e.g. InSAR have made it easier to  
528 detect these slow movements which has led to increasing understanding and  
529 recognition of DSGSD (e.g. Ambrosi and Crosta, 2006; Colesanti and Wasowski,  
530 2006; Hilley et al., 2004; Saroli et al., 2005; Teshebaeva et al. 2019).

531 DSGSD have been observed in various rock types and are generally characterized  
532 by discontinuous or poorly defined lateral boundaries (Ambrosi and Crosta, 2006;  
533 Varnes et al., 1990). According to Agliardi et al. (2001) diagnostic features of DSGSD  
534 can include: “morpho-structures (doubled ridges, scarps, counterscarps, etc.) similar to  
535 those observable, at a smaller scale, in cohesive soils landslides; size of the  
536 phenomenon comparable to the slope; present day low rate of displacement (mm/y, in  
537 alpine and prealpine areas); presence of minor landslides inside the deformed mass  
538 and ancient collapses of the lower part of the slope”.

539 DSGSD are not considered as imminent hazardous phenomena because of their  
540 slow movement rate, but a part of the creeping mass can evolve into multiple  
541 sequential failures or into fast large catastrophic slope failures such as rockslides and  
542 rock avalanche (Agliardi et al. 2012). These large-scale events, even though they are  
543 infrequent, can cause hazardous conditions e.g. on glaciers and surrounding areas  
544 (e.g. Deline et al. 2015b; Dufresne et al. 2019; Hewitt et al. 2011), in fjord environments

545 were rock-slope failures can generate tsunamis (Böhme et al. 2011; Kuhn et al. 2019)  
546 and in valley systems were large rock-slope failures can dam rivers (Oppikofer et al.,  
547 2020; Strom and Korup, 2006).

548 Many case studies of DSGSD have been described in various settings around the  
549 world, but it is difficult to understand their controlling factors due to the variety of  
550 geologic and environmental settings which they occur in (Agilardi et al. 2012). They  
551 are widespread in orogenic mountain ranges (Ambrosi and Crosta, 2006; Agliardi et  
552 al. 2009; 2012; Korup, 2006;), and have been described from the Central-Eastern Alps  
553 (Forcella, 1984; Forcella and Orombelli, 1984; Zischinsky, 1966), Norway (Böhme et  
554 al. 2013; Oppikofer et al. 2017; Schleier et al. 2016), Spitsbergen (Figure 8) (Kuhn et  
555 al. 2019; 2021), Iceland (Coquin et al. 2015; 2019) and other places.

556 Research challenges, on mass movements in Paraglacial environments and in  
557 some extend in Periglacial environments, have during the last decades focused more  
558 and more on the hidden hazard which DSGSD can cause to human habitation in these  
559 environments. Recent and future developments in remote sensing will without doubt  
560 deepen our understanding on the physical mechanisms and triggering factors for  
561 DSGSD. These developments will improve our monitoring and hazard analyses.

562

## 563 **5. Climate change and hazard of mass movements in cold and** 564 **polar climates**

565 According to the definition of (Varnes, 1984), landslide hazard is defined as the  
566 probability that a landslide event of a certain magnitude occurs within a period of time  
567 in a given area. Landslide risk - as any other risk - refers to the expected degree of  
568 loss due to a landslide event, and it is the result of the product of the landslide hazard  
569 and the vulnerability, with the latter being the degree to which a community, a structure,

570 a service or an area is likely to be damaged by a hazardous phenomenon (UNDHA,  
571 1992). Mass movements can be extremely dangerous, as they are processes that can  
572 severely threaten human life and affect infrastructures (Crozier and Glade, 2005;  
573 Geertsema et al., 2009; Haque et al., 2016; Kjekstad and Highland, 2009; Petley,  
574 2012). Slope failures can involve different types of materials and can have a great  
575 diversity of velocities and fluid contents, making them hard to understand and produce  
576 the appropriate countermeasures.

577 Even if high level of uncertainty of global-climate predictions and the lack of spatial  
578 resolution of the available downscaled projections, there is a strong theoretical basis  
579 for increased landslide activity as a result of climate change (Borgatti and Soldati,  
580 2010; Crozier, 2010; Gariano and Guzzetti, 2016; Haque et al., 2019; Wood et al.,  
581 2020). There are few studies (Coe, 2020; Coe et al., 2018; Schlögel et al., 2011;  
582 Uhlmann et al., 2013) that systematically evaluate and compare the magnitude,  
583 frequency, and extent of landslide events in glacial and periglacial environments on  
584 the long term (Allen et al., 2011; Bessette-Kirton and Coe, 2020; Hock et al., 2019).  
585 Cold environments are particularly sensitive to variations in atmospheric temperatures  
586 (Clague et al., 2012b; Huggel et al., 2012) that lead to an enhanced reaction of slopes  
587 to glacial retreat and permafrost degradation (e.g., Gruber et al., 2017; Gruber and  
588 Haeberli, 2007; Harris et al., 2009; Hock et al., 2019; Kos et al., 2016; Patton et al.,  
589 2019a; Schoeneich et al., 2011). As result of climate change, mass movements in  
590 cryospheric regions can be triggered by warming periods, freeze-thaw, rapid  
591 snowmelt, intense and/or prolonged precipitation (Huggel et al., 2012, 2010; Stoffel  
592 and Huggel, 2012). From a hazard perspective, understanding how these factors will  
593 impact failures' frequency and magnitude is becoming increasingly important.

594 In glacial environments, retreat and thinning of glacial ice reduces lateral support of  
595 steep rock walls (Ballantyne, 2002). This can generate a reduction in slope stability by

596 debutressing steep valley walls (Lane et al., 2017), that are impacted by modifications  
597 in the stress field and subject to mechanical and thermal alteration, and consequently  
598 slope instabilities (Cody et al., 2020; Coquin et al., 2015; Evans and Clague, 1994;  
599 Grämiger et al., 2017; McColl, 2012; McColl and Draebing, 2019; Mercier et al., 2017;  
600 Wegmann et al., 1998). Failures linked to glacial retreat can range from small-scale  
601 rock fall to catastrophic rock/debris avalanches. As glacial environments are highly  
602 dynamic because of increasing atmospheric temperatures, the landslide hazard in  
603 these regions is more difficult to assess because of the changing zones of instability  
604 and initiation (e.g., Evans and Clague, 1988; Geertsema et al., 2006a,b; Kääb et al.,  
605 2005). The consequences of landslides in glacial environments can be more  
606 unexpected and severe than the events themselves, generating hazardous  
607 consequences because of dam-creation (Fan et al., 2020; Oppikofer et al., 2020;  
608 Strom and Korup, 2006), or tsunami if they reach the water (Dahl-Jensen et al., 2004;  
609 Dai et al., 2020; Dufresne et al., 2018a). A better understanding of changes in slope  
610 stability caused by the loss of glacier ice could improve the definition of landslide  
611 hazard zones, aiding to protect areas and populations exposed to these hazards  
612 (Haeberli et al., 2017b; Hock et al., 2019).

613 Climate change is causing permafrost degradation at high latitudes and in mountain  
614 regions (e.g., Blunden and Arndt, 2020; Slater and Lawrence, 2013), and this is  
615 confirmed by local monitoring permafrost studies in the European Alps, Alaska,  
616 Canada, Siberia, Iceland, Greenland, northern Europe, the Tibetan Plateau, the Andes  
617 and the Antarctic Peninsula (Christiansen et al., 2010; Czekirda et al., 2019; Gislås et  
618 al., 2017; Harris et al., 2009; Panda et al., 2014; Pastick et al., 2015; Romanovsky et  
619 al., 2010; Westermann et al., 2017; Zou et al., 2017). Among the consequences of  
620 permafrost degradation, the most hazardous are mass movements like rock falls,  
621 debris flows, rock/debris avalanches (Darrow et al., 2017; Deline et al., 2015a;

622 Huscroft et al., 2004; Patton et al., 2019a; Schoeneich et al., 2011). Mountains host  
623 12% of the global population (Schild, 2008), and about a quarter of the permafrost on  
624 Earth is hosted in mountainous terrains (Gruber, 2012). Therefore, mass movements  
625 in periglacial mountain environments pose at direct risk the local population and  
626 infrastructure (e.g., Haeberli, 2013; Schoeneich et al., 2011), and can impact local  
627 economies and tourism, communication and transport systems and power generation  
628 industries (e.g., Arenson and Jakob, 2017; Duvillard et al., 2019; Mourey et al., 2019;  
629 Pröbstl-Haider et al., 2016). The unpredictability of the extension and speed of  
630 permafrost degradation makes the understanding of landslide processes in mountain  
631 periglacial environments even more urgent, in order to predict and mitigate their  
632 hazardous consequences.

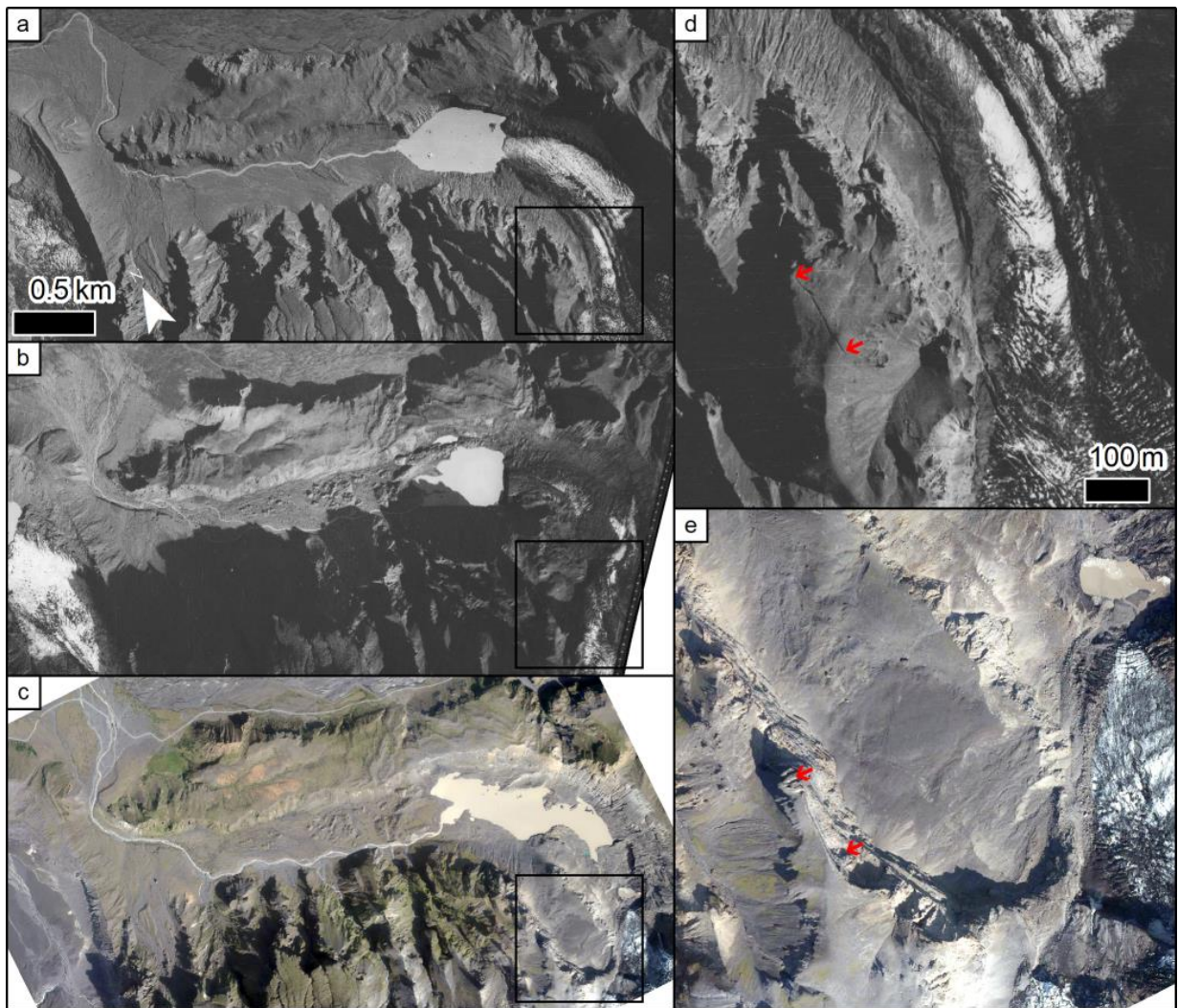
633 Due to their magnitude and mobility, rock/ices avalanches are also particularly  
634 hazardous in both glacial and periglacial environments (Davies and McSaveney, 2012;  
635 Evans and Clague, 1988; Geertsema et al., 2006a, 2006b; Hungr and Evans, 2004).  
636 Several examples with volumes ranging up to millions of cubic metres include the  
637 rock/ice avalanche from Monte Rosa in the western Italia Alps in 2007 (Fischer et al.,  
638 2011), the rock/ice avalanches at Mount Munday in 1997 and at Kendall Glacier in  
639 1999 in British Columbia (Delaney and Evans, 2014; Geertsema et al., 2006a), the  
640 2005 Mt. Steller rock/ice avalanche in Alaska (Huggel et al., 2008), the 2002 rock/ice  
641 avalanche at Kolka/Karmadon, Russian Caucasus, with the latter one causing the  
642 death of ~140 people (Evans et al., 2009c; Haeberli et al., 2004; Huggel et al., 2005).

643 Hazard related to mass movements in both glacial and periglacial environments  
644 include also catastrophic phenomena such as glacial lake outburst floods (GLOFs).  
645 GLOFs are floods that originate from lakes formed behind a landslide dam (Fan et al.,  
646 2020; Oppikofer et al., 2020; Strom and Korup, 2006) or dammed by a frontal moraine  
647 or glacial ice (Carey et al., 2012; Clague and Evans, 2000; Cook et al., 2016; Haeberli



648 et al., 2016; Harrison et al., 2018; Hubbard et al., 2005; Richardson and Reynolds,  
649 2000; Veh et al., 2019; Westoby et al., 2014). The lakes can drain catastrophically for  
650 several reasons, including mass movements into the lake (Figure 9) (Costa and  
651 Schuster, 1988; Korup and Tweed, 2007). Their influence on settlements and infra-  
652 structure can be devastating (e.g. Kjartansson 1967; Dahl-Jensen et al., 2004; Stoffel  
653 & Huggel, 2012; Dai et al., 2020; Deline et al., 2015b; Dufresne et al., 2018a, b). In  
654 regions like the Himalaya, GLOFs have been recognised as the dominant long-term  
655 drivers of fluvial erosion and valley evolution (Cook et al., 2018). They are extremely  
656 destructive due to their magnitude and unpredictability (Khanal et al., 2015; Mool et  
657 al., 2011; Schwanghart et al., 2016; Shrestha et al., 2010). GLOFs' frequency and  
658 magnitude are thought to be increasing in response to climate change (Bajracharya  
659 and Mool, 2009; Bolch et al., 2011; Dussaillant et al., 2010; Harrison et al., 2018; Lutz  
660 et al., 2016). This is due to the fact that outlet glaciers have thinned and retreated from  
661 their basins, resulting often in unstable slopes combined with accelerated formation  
662 and rapid growth of proglacial lakes (e.g. Guðmundsson et al. 2019; IPCC, 2012).

663



664

665 *Figure 9. The Steinsholtsjökull rockslide and the GLOF in the Steinsholtsdalur valley*

666 *in 1967. The extent of panels d and e are shown as black boxes on panels a-c. (a)*

667 *Aerial photograph taken in August 1960 (b) Aerial photograph taken in September*

668 *1967, eight months after the rock slide. (c) Modern image showing clearly that the*

669 *headwall of the slide was about 900 m long. (d) Detailed view showing a 100 m long*

670 *fracture in the Innstíhaus above the Steinsholtsjökull outlet glacier. (e) Same view as*

671 *in panel d, but using the modern image shown in panel c to illustrate the amount of*

672 *material mobilised. Aerial images are from the National Land Survey of Iceland.*

673

674 Landslides and mass movements such as active-layer-detachment slides and thaw

675 slumps in degrading-permafrost terrains can impact on the environments in which they

676 occur. The physical and ecological structure of hillslopes and fluvial networks can be  
677 affected by an increased sediment, nutrients, and soil displacement of mass  
678 movements (Abbott and Jones, 2015b; Bowden et al., 2008; Gooseff et al., 2009;  
679 Osterkamp et al., 2009; Turetsky et al., 2019; Zongxing et al., 2016). These  
680 phenomena can generate disturbance to terrestrial and aquatic ecosystems (Bowden  
681 et al., 2008; Callaghan et al., 2011; Cannone et al., 2010; Osterkamp et al., 2009;  
682 Racine et al., 2004). An increase in mass movements occurrence in cold and polar  
683 terrains due to permafrost thaw because of the warming climate may also accelerate  
684 the release to the atmosphere of Pleistocene sequestered carbon (Belshe et al., 2013;  
685 Grosse et al., 2011; Masyagina et al., 2019; Pautler et al., 2010; Schuur et al., 2009,  
686 2008; Tarnocai et al., 2009). Thaw-slumps and active-layer-detachment-slides activity  
687 in the Arctic has enhanced in the last decades (Jorgenson and Osterkamp, 2005; Luo  
688 et al., 2019), generating alterations in the water quality of nearby lakes and streams  
689 (Kokelj et al., 2013; Lafrenière and Lamoureux, 2019; Segal et al., 2016), vegetation  
690 and soils (Khomutov and Leibman, 2014; Lantuit et al., 2012), and trace gases (Abbott  
691 and Jones, 2015).

692

## 693 **6. Conclusions**

694 In this chapter, we emphasized the major types of rapid mass movements that  
695 reflect the most characteristic expression of rapid down-slope sediment transport in  
696 cold and polar environments under glacial, periglacial and paraglacial conditions.  
697 Many of these rapid mass wasting processes are not exclusive to the glacial,  
698 periglacial and paraglacial domains, but we highlighted that permafrost and ice are the  
699 critical agent in their development. We focussed on the specific attributes that  
700 distinguish these mass movements from ones in other environments and the

701 associated specific research challenges. Cryospheric terrains in mountainous and  
702 high-latitude environments are the most susceptible to slope stability problems, as  
703 warming temperature cause permafrost and ice to degrade. These changes if  
704 maintained or worsen can have significant implications for hazard analyses and  
705 adaptation strategies in these environmentally sensitive areas of the world.

706

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