7.22 Mass-Movement Causes: Glacier Thinning

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7.22.1 Introduction

Mass movements can have a variety of causes. Earthquakes, permafrost degradation, wildfire, and heavy rainfall are common contributors to natural landslides. Tectonics, marked climate change, volcanic activity, and inherent weaknesses in soil and rock are some of the factors that may distinguish one region from another in terms of landslide activity, and indeed these sorts of factors contribute to global and regional landslide hot spots (Nadim et al., 2006). Glaciers also play a strong role in conditioning slopes to mass movement (Evans and Clague, 1994), and the impacts on hillslopes persist after glaciers disappear.

Paraglacial processes (i.e., glacially conditioned, nonglacial processes) were first introduced to the literature by Ryder (1971a, b) and defined by Church and Ryder (1972) to describe sedimentation rates on alluvial fans. Since then much work has been done to extend the concept of paraglacial geomorphology to other processes (see Ballantyne, 2002a, b and references therein) including landslides.

Glaciers play an important role in conditioning mountain slopes for mass movement. Mass movements in mountainous terrain are strongly influenced by the advance and retreat of glaciers (Evans and Clague, 1994). Many studies show that recently deglaciated areas are hot spots for a variety of landslide types (e.g., Bovis, 1982; Ballantyne, 2002a, b; Holm et al., 2004; Geertsema et al., 2006; Cossart et al., 2008). We note that glaciers are agents of erosion and deposition. Although some valleys are deepened by glaciers, others may be filled with glacigenic sediments. Both scenarios present challenges for slope instability. Here, we restrict our discussion to the effects of glacial erosion and thinning on mass movement.

We use the word ‘landslide’ as a generic term for a variety of mass-movement types and processes (Cruden and Varnes, 1996). Although we recognize that landslides are commonly complex, involving movements in both rock and soil (Geertsema et al., 2006), here we consider the influence of glacial thinning on landslides in soil and rock separately.

7.22.2 Landslides in Soil

Glaciers erode, redistribute, and deposit sediment. As glaciers retreat and thin, they expose both scoured bedrock and glacigenic sediment (Figure 1). Exposed soils can range from thin veneers over bedrock to complex packages of sediment that are tens of meters thick. Genetic materials can include till,
glaciolacustrine deposits, colluvium, glaciofluvial, and aeolian sediments. In some cases, the landforms are arranged in such a way that they become dams for temporary lakes. The unvegetated, and, in some cases, ice-cored sediments, are subject to a variety of hillslope processes (Ballantyne, 2002a) and can be readily mobilized by heavy precipitation, snow and ice melt, and dam-burst floods (Chiarle et al., 2007).

7.22.2.1 Exposure of Glacial Sediment

Retreating and thinning glaciers commonly expose extensive areas of bare sediment (Figure 1). Where soil is exposed on steep slopes, debris flows (Figure 2) are the main (most efficient) agents of erosion and sediment transport (Ballantyne, 2002a). Bulking of sediment to channels from surface erosion or debris slides generally leads to debris flows with longer travel distances. In addition, channelized runoff may entrain sediment and transform movements into debris flows. Although intense rainstorms and rapid snow melt are the principal triggers of debris flows, the available sediment supply, the young steep slopes, and the lack of vegetative cover set these recently deglaciated areas apart from other landslide-prone regions.

In a survey of 19 alpine basins in the Lillooet River valley, Holm et al. (2004) found that debris movements were concentrated along Little Ice Age lateral moraines and trimlines. Similarly, Zimmermann and Haeberli (1992) found that more than half of the debris flows triggered by intense rain in the Swiss Alps in 1987 initiated on steep, Little Ice Age vintage, till-covered slopes.

7.22.2.2 Paraglacial Dam Break Floods and Related Debris Flows

Retreating and thinning glaciers may produce a variety of (usually short lived) lakes (Clague and Evans, 1994). Glacial lakes may be dammed by moraines, landslides, alluvial fans, and the glaciers themselves. Outburst floods from glacier-dammed lakes (Figure 3) can happen annually, and the lakes may experience a jökulhlaup cycle (Geertsema and Clague, 2005). Conversely, lakes behind moraine dams do not refill once the dams are breached. Floods and debris flows from the escaping waters can be catastrophic because of the large volumes involved, the long travel distances, and their unpredictability.

Dam-burst floods tend to be much larger than nival floods; because river systems are rarely plumbed for extreme events, a suite of cascading effects may result, including secondary debris slides from valley walls and sometimes large debris flows. An example of such a suite of events comes from moraine-dammed Klattasine Lake in British Columbia. The lake drained catastrophically sometime between June 1971 and September 1973, releasing about $1.7 \times 10^9$ m$^3$ of water. The resultant flow dissected the moraine and mobilized large...
quantities of sediment from the channel and valley margins. As the sediment concentration increased, the flood rapidly transformed into a debris flow that traveled 8 km downstream (Clague et al., 1985). The debris flow destabilized adjacent valley slopes causing secondary landslides.

Outbursts floods from englacial and subglacial lakes can be insidious because the water pockets are difficult to detect. An englacial lake with a volume of 65,000 m$^3$ was recently identified with Tête-Rousse Glacier on Mont Blanc, France. Local authorities, concerned about the threat an outburst flood posed to several villages in the valley below, decided to drain the water through vertical holes drilled into the glacier (Chiarle et al., 2011). In 1892, a 200,000-m$^3$ englacial lake drained in the same area, triggering an 800,000-m$^3$ debris flow. The flow, which ran through the village of Saint Gervais and killed 175 people, is one of the worst glacier-caused disasters in Europe (Vincent et al., 2010).

### 7.22.2.3 Melting of Ground Ice

Dead glacier ice (ground ice) is common around the margins of retreating glaciers. Even when it is no longer in equilibrium with climate, such ice may persist primarily because of the insulating effect of overlying debris. Slow thawing of the ice, particularly under a warming climate, can contribute to debris flow initiation. Although many flows are triggered by rainfall, some may occur simply from melting of the ice. Mass movements on rupture surfaces of buried glacial ice are similar to those that involve ice-rich permafrost. The ice acts as a barrier to water movement and, therefore, concentrates flow. Meltwater and rainfall can be concentrated along a frozen rupture surface.

Ice-rich debris and ice lenses have been observed in many debris flow initiation zones (Chiarle et al., 2007). Thawing of ground ice likely contributed to a debris flow on 29 July 2005 in Val di Fosse in the eastern Italian Alps. Melting of a buried ice mass, exposed in a 20-m-long detachment zone at 3000 m asl, triggered a 15,000-m$^3$ debris flow (Chiarle et al., 2011). Similarly, Mattson and Gardner (1991) documented 25 landslides in ice-cored moraines in the Alberta Rockies. The landslides initiated at the ice–debris interface and were triggered by rainfall as well as melting of ice.

We expect thawing of ground ice to increase in importance as a paraglacial contributor to alpine debris flows as glaciers continue to thin. Resulting debris flow activity may lag behind glacier retreat because of the lack of direct contact of the buried ice with the atmosphere. Once debris is removed by an initial landslide, or becomes thinner, we expect melting of ice to accelerate and to lead to more debris flows.

### 7.22.3 Landslides in Rock

Glacier dynamics can have profound effects on the stability of rock slopes. Both glaciation and deglaciation weaken rock masses (Augustinus, 1995; Ballantyne, 2002a; Cossart et al., 2008). This happens in a four-step process: (1) deepening and widening of valleys (glacial erosion); (2) ice loading resulting in the development of stress fractures; (3) debuttressing (during deglaciation); and (4) stress release as manifested by joint expansion (Figures 4 and 5).

Landslides resulting from glacially conditioned, rock mass instability include rock topples and falls, slow deep-seated slope deformation, and catastrophic rock avalanches. Rockfall (cliff collapse) and slow rock slope deformation (and sagging) can transform into long runout rock avalanches (Figure 5) (Chigira, 1992; Geertsema et al., 2006). The spatial and temporal association of rock landslides with previously glaciated areas suggests that glacial conditioning of rock slopes plays an important role in their occurrence.

### 7.22.3.1 Glacial Conditioning of Rock Masses

Eroding glaciers deepen valleys and steepen valley walls, particularly where ice flow is parallel to the long axis of valleys. The deepening of valleys increases rock wall heights, which, in turn, elevates the self-weight shear stresses of the rock mass (Bovis, 1982; Ballantyne, 2002a) and may increase tensile stresses in toe slopes (Augustinus, 1995).

Glacial loading, in addition to having regional isostatic impacts, imparts stresses on both valley walls and floors (Ballantyne, 2002a). Fracture orientation from glacial loading can be distinguished from structural joint sets because of a similar orientation to glacier movement (e.g., Bovis, 1990). According to Benn and Evans (1998), glacier loading stresses vary at different scales. At the regional scale stress is a function of ice thickness and surface gradient, at the mountain slope scale stresses increase toward the valley floor, and at the site level stresses are greater on stoss faces of rock ridges. Cossart et al. (2008) calculated stresses at these various scales in an investigation of glacial debuttressing on landslides in southeastern France. They calculated values of up to 9000 and 300 kPa for normal and longitudinal ice-loading stresses, respectively.

Glacial debuttressing is one of the most important consequences of deglaciation in mountain environments (Ballantyne, 2002a) and involves the expansion of joint sets in...
As glaciers retreat and thin valley walls, they cease to support valley walls. Most rock masses display some degree of elasticity and will rebound to some degree in response to the removed load. Note that this is different from isostatic rebound. When rock slopes are exposed by glacier thinning, relaxation of tensile stresses in the rock mass is released to a degree dependent on the amount of residual strain energy and the modulus of elasticity of the rock (Ballantyne, 2002a). This rebound in response to debuttressing may result in the opening of joints (Figure 4). The joints develop independently of regional structure – the response is to glacial loading – and correspond to the direction of ice flow. This paraglacial relaxation of rock masses conditions them for mass movement.

7.22.3.2 Landslides on Glacially Conditioned Rock Slopes

Although glaciers can condition rock slopes for failure, stability is still largely controlled by lithology and structure. Augustinus (1995) in New Zealand and Holm et al. (2004) in British Columbia observed few landslides in glacially debuttressed plutonic rock in contrast to abundant landslides in weaker, sedimentary, metamorphic, or volcanic bedrock. In some cases, it becomes difficult to distinguish the influences of tectonic stresses, faulting, extension, compression, mountain building, river incision, volcanic influences, and similar factors from the effects of glacial debuttressing (see, e.g., Hermans et al., 2001; Prager et al., 2008). Nonetheless, the evidence supporting a relationship between glacial debuttressing and the failure of rock slopes is strong (e.g., Panizza, 1973; McSaveney, 1993; Hancox and Perrin, 1994; Abele, 1997; Ballantyne et al., 1998; Fort, 2000; Holm et al., 2004; Strom, 2004; Geertsema et al., 2006; Cossart et al., 2008; Hewitt et al., 2008).

Paraglacial rockfall contributes to the development of talus aprons at the base of cliffs. Many authors (see Ballantyne, 2002a) have shown that present-day rockfall accumulation rates are much too small (order of magnitude too small) to account for the amount of postglacial talus accumulation. The conclusion is that rockfall rates were much greater in the early paraglacial phase of the postglacial period. In a few cases, rockfall (cliff collapse) can transform into long runout rock avalanches.

Another response of rock slopes to glacial debuttressing is slow deformation, or gravitational sagging, or spreading. These slow movements commonly display horst and graben topography, gulls, and uphill-facing scarps. Bovis (1982, 1990) confirmed that $3 \times 10^7$ m$^3$ of rock was deforming as a result of post-Little Ice Age glacial debuttressing following downwasting of Affliction Glacier in British Columbia. Similarly, Evans and Clague (1994) reported on post-Little Ice Age sackung development above Melburn Glacier in the St. Elias Mountains of British Columbia. Slow rock deformation can be a precursor to catastrophic rock slope failure (Chigira, 1992; Geertsema et al., 2006). In some cases, similar lithologies can result in different types of rock failure. A slope in the northern Rocky Mountains of British Columbia responded to debuttressing with a slow, deep-seated rotational slide as well as a rock avalanche (Figure 6).

Rock avalanches are generally characterized by their large volumes, long runout distances, and low travel angles. Many rock avalanches after the Little Ice Age occur on rock slopes above glaciers (McSaveney, 1993; Evans and Clague, 1994; Holm et al., 2004). For example, two-thirds of the rock avalanches in northern British Columbia that occurred between 1973 and 2003 (Geertsema et al., 2006) occurred on cirque...
walls where glaciers have thinned up to several hundred meters. Some early Holocene rock avalanches are measured in volumes of tens of cubic kilometers. Examples occur in many parts of the world (e.g., Hancox and Perrin, 1994; Hewitt et al., 2008), and most of these gigantic events occurred very shortly after deglaciation.

Further evidence for paraglacial deturbation of rockslides is provided by an exhaustion model of Cruden and Hu (1993) applied to the Canadian Rockies. The model assumes that a limited number of failure sites exist and that failure occurs only once at a given site. This causes the probability of failure to diminish exponentially over time and goes a long way to explain some of the immense prehistoric landslide deposits. Of course, we do see evidence of repeated rock avalanches in the same areas, but their model provides a good general framework. The other problem with rock avalanches is that they may happen after a long period of conditioning by slope deformation. So while rockfall and slope deformation may be immediate responses to glacial thinning, some rock avalanches require more specific conditions – which develop overtime (e.g., Abele, 1997). Ultimately, some mid-Holocene rock avalanches may still be occurring as paraglacial responses to deglaciation.

7.22.4 Conclusions

Glaciers play an important role in conditioning landscapes for mass movement. Glaciers rearrange and override sediments, only to expose them to elements when the glaciers recede. The exposed and commonly steep soils are rapidly modified by erosion and debris slides and flows. Outburst floods from temporary lakes also occur, and these may also transform into debris flows. Not only do glacial loads impart stress fractures to bedrock, the stress release and associated slope relaxation can result in rockfall, slow deep-seated slope deformation, and (under the right circumstances) rock avalanches. As glaciers continue to thin in the twenty-first century, we can expect continued associated landslide activity.

References

Biographical Sketch

Marten Geertsema, a geomorphologist, has worked for the British Columbia Forest Service since 1985. He is interested in terrain hazards and their causes, including past, present, and future hazard regimes. Marten has degrees in soil science and earth science from the University of Alberta, and a doctorate in physical geography from the University of Utrecht in the Netherlands. He is a professional geoscientist and a professional agrologist, and holds adjunct professor positions at the University of Northern British Columbia and Oklahoma State University.

Marta Chiarle is a geomorphologic researcher with the Research Institute for Geo-hydrological Protection of Turin, of the Italian National Research Council. Marta has also worked as a visiting scientist with the US Geological Survey in Colorado, and Simon Fraser University in British Columbia. She is interested in mountain hazards and climate change, and has a special interest in alpine debris flows and periglacial geomorphology. Marta has a degree in Earth Science from the University of Turin, and a doctorate in Geological Engineering from the Politecnico di Torino, Italy.