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Joint-bounded crescentic scars formed by subglacial clast-bed contact forces: Implications for bedrock failure beneath glaciers

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ABSTRACT

Glaciers and ice sheets are important agents of bedrock erosion, yet the precise processes of bedrock failure beneath glacier ice are incompletely known. Subglacially formed erosional crescentic markings (crescentic gouges, lunate fractures) on bedrock surfaces occur locally in glaciated areas and comprise a conchoidal fracture dipping down-ice and a steep fracture that faces up-ice. Here we report morphologically distinct crescentic scars that are closely associated with preexisting joints, termed here *joint-bounded crescentic scars*. These hitherto unreported features are ca. 50–200 mm deep and involve considerably more rock removal than previously described crescentic markings. The joint-bounded crescentic scars were found on abraded rhyolite surfaces recently exposed (<20 years) beneath a retreating glacier in Iceland, as well as on glacially sculpted Precambrian gneisses in NW Scotland and various Precambrian rocks in Ontario, glaciated during the Late Pleistocene. We suggest a common formation mechanism for these contemporary and relict features, whereby a boulder embedded in basal ice produces a continuously migrating clast-bed contact force as it is dragged over the hard (bedrock) bed. As the ice-embedded boulder approaches a preexisting joint in the bedrock, stress concentrations build up in the bed that exceed the intact rock strength, resulting in conchoidal fracturing and detachment of a crescentic wedge-shaped rock fragment. Subsequent removal of the rock fragment probably involves further fracturing or crushing (comminution) under high contact forces. Formation of joint-bounded crescentic scars is favoured by large boulders at the base of the ice, high basal melting rates, and the presence of preexisting subvertical joints in the bedrock bed. We infer that the relative scarcity of crescentic markings in general on deglaciated surfaces shows that fracturing of intact bedrock below ice is difficult, but that preexisting weaknesses such as joints greatly facilitate rock failure. This implies that models of glacial erosion need to take fracture patterns of bedrock into account.

1. Introduction

The two main mechanisms of subglacial erosion of hard bedrock beneath glaciers and ice sheets are generally regarded to be abrasion and plucking (quarrying), with an added component of subglacial meltwater erosion (e.g., Drewry, 1986; Glasser and Bennett, 2004). A major challenge in understanding plucking has been the need to explain the failure (breaking) of hard bedrock by moving ice (e.g., Morland and Boulton, 1975; Iverson, 1991; Hallet, 1996). Ice has a shear strength and a compressive strength that is much lower than most intact rock types (see Table 1), and envisaging how such a weak material can fracture a much stronger material is difficult. More recent studies have now established that most, if not all, plucking occurs along preexisting fractures (joints) so that intact bedrock need not be fractured directly by ice above it (Rea, 1994; Dühnforth et al., 2010; Krabbendam and Glasser, 2011; Hooyer et al., 2012; Iverson, 2012). In essence, preexisting weaknesses such as joints (typically formed by uplift, cooling, or tectonic stresses) are merely exploited by the plucking process to further break apart the rock. However, the occurrence of crescentic markings—a family of small-scale erosional bedforms all characterised by conchoidal fracturing—suggests that fracturing of intact bedrock does occur below glaciers and ice sheets under certain circumstances.

Crescentic markings include *crescentic gouges*, *lunate fractures* and *crescentic fractures* (Chamberlin, 1888; Gilbert, 1906; Lahee, 1912; Harris, 1943; Okko, 1950; Dreimanis, 1953; Slocum, 1978; Wintges, 1985; Glasser and Bennett, 2004). Different names have been used to describe these (shown in brackets in Fig. 1); we prefer the descriptive term *crescentic markings* here as a collective term above the interpretational term friction cracks (cf. Harris, 1943). The term *chattermarks* is at times used as a general term (see discussion in Harris, 1943); more strictly they refer to a series of fractures confined to grooves (Benn and Evans, 1998), but the fractures may not necessarily be crescentic

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(Harris, 1943). Most crescentic markings have been reported on massive, unfoliated, or poorly foliated bedrock such as granite, sandstone, or quartzite (Chamberlin, 1888; Harris, 1943; Okko, 1950; Slocum, 1978; Wintges, 1985). There is general agreement that crescentic markings are formed by high clast-bed contact forces, exerted by large cobbles or boulders embedded in basal ice pressing onto the bed (Gilbert, 1906; Harris, 1943; Hallet, 1979; Ficker et al., 1980; Wintges, 1985; Drewry, 1986; Glasser and Bennett, 2004). The occurrence of crescentic markings may thus provide information about the subglacial conditions under which high clast-bed contact forces and bedrock failure can develop.

During work around the British Geological Survey Virkisjökull Glacial Observatory in southern Iceland, very well preserved, small- and medium-scale erosional bedforms were noted on a recently (<20 years) exposed bedrock high (Fig. 2). We noted numerous crescentic scars that are intimately related with preexisting, preglacial joints in the bedrock; an association that, to our knowledge, has not been reported before. Here, we define these as *joint-bounded crescentic scars*. In this paper we present a detailed analysis of these bedforms identified at Virkisjökull alongside the properties of the bedrock. Joint-bounded crescentic scars were also found on Precambrian gneiss in the NW Highlands of Scotland and on various Precambrian lithologies on the Canadian Shield of Ontario, Canada. These relict examples developed during Pleistocene glaciation(s). We go on to discuss how high clast-bed contact forces can be generated below ice; how these can result in stresses that exceed the strength of the bed adjacent to joints; we propose a formation mechanism for joint-bounded crescentic scars; and finally discuss how these bedforms can inform us about the circumstances under which bedrock failure of the glacier bed may occur.

2. Methods

Detailed geological and geomorphological observations were made in the study area in Iceland, including orientations of glacial striae and bedrock joints, along with depth and angle of joint-bounded crescentic scars. On a number of level bedrock surfaces, vertical photos were taken with a scale, carefully oriented to true north. These photos were georeferenced in ArcGIS in an artificial coordinate system, in essence creating a georeferenced outcrop-scale aerial photo from a height of ca. 2 m (Fig. 3). Glacial striae and the outline and long axes of crescentic scars were digitised on the georeferenced images. From these digital lines, the orientation and length were extracted (using ArcTools and a spreadsheet); these are presented in rose diagrams made using Stereonet V8 software (Cardozo and Allmendinger, 2013). Joint analysis uses the principle of the *circle inventory method* (Davis and Reynolds, 1996, p. 720) to prevent any directional bias, but was speeded up by using the georeferenced outcrop photos. Circles of known area were drawn in the GIS and all joint traces were digitised within that particular circle (see also Krabbendam and Bradwell, 2014). Joint orientations can then be easily extracted from the data set and presented in rose diagrams, using the same method as the striae.

Schmidt Hammer rebound values are a function of rock hardness and show an empirical exponential relationship with uniaxial compressive strength (Aydin and Basu, 2005) and can thus be regarded as a proxy for intact rock strength. Schmidt Hammer (type-N) rebound measurements were taken on outcrops away from edges and joints, as well as on a number of large (>1 m) boulders. Only smooth surfaces were tested. Ten blows per site were performed in slightly different positions. Anomalous values, typically on the low side, were rejected; the average of the remainder were taken as the rebound measurement.

3. Glaciology, geology, and geomorphology background, Virkisjökull, Iceland

Virkisjökull is an outlet glacier draining the Öraefajökull Ice Cap in southern Iceland, which largely covers the Öraefajökull volcano, one of the largest active volcanoes in Iceland. Öraefajökull volcano consists mainly of basaltic hyaloclastite (volcanic breccias formed by lava-ice and lava-water interaction), basaltic tuffs, basalt lava flows, and minor rhyolitic intrusions (Prestvik, 1979; Stevenson et al., 2006). Virkisjökull has experienced jökulhlaups, most recently during an eruption in 1362 CE (Thorarinsson, 1958). Bradwell et al. (2013) provided a detailed glaciological description of Virkisjökull and its recent history of rapid retreat (ca. 40 m y^{-1}; 1995–2012), which has revealed a bedrock high on the northwest margin of the glacier ablation zone (Figs. 2A, B, 4A). This bedrock high comprises the main study area; it is elongate (300 m wide, 100 m long) broadly parallel to ice flow, and varies in height between 20 and 40 m. The top surface of the bedrock high is ca. 100 m below the maximum level attained by the glacier surface in the Little Ice Age, and the lee side is ca. 1 km from the Little Ice Age

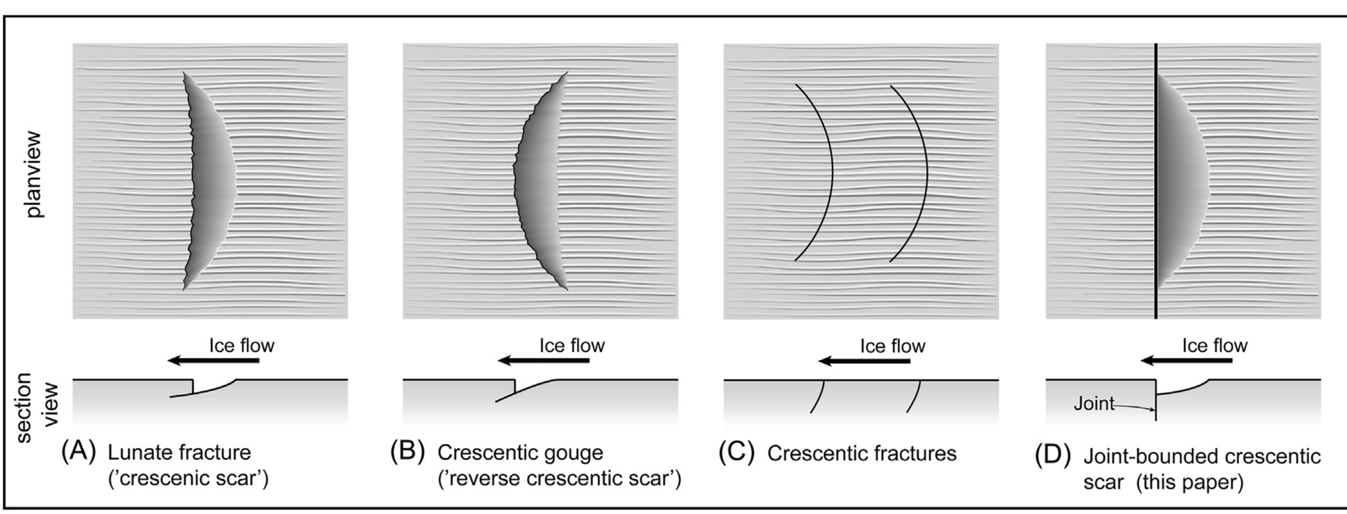

Fig. 1. Crescentic markings, in plan view (top) and cross-section (bottom); modified after Embleton and King (1975), with terminology of Prest (1983) in brackets. (A) lunate fracture or crescentic scar; (B) crescentic gouge or reverse crescentic scar; (C) crescentic fractures, without removal of bedrock; (D) joint-bounded crescentic scar (this paper). Note different relation between conchoidal fracture and subvertical fracture/joint. The conchoidal fracture invariably dips in the direction of ice flow.