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Surges of outlet glaciers from the Drangajökull ice cap, northwest Iceland

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A B S T R A C T

Surface elevation and volume changes of the Drangajökull surge-type glaciers, Reykjarfjarðarjökull and Leirufjarðarjökull, were studied by comparing digital elevation models that pre-date and post-date their most recent surges. Annual glacier-frontal measurements were used to estimate average ice velocities during the last surge of the glaciers. The observations show a distinct ice discharge, most of which was from the upper reservoir areas, down to the receiving areas during the surges. The surface draw-down in the reservoir areas was usually 10–30 m during the surges, while the thickening of the receiving areas was significantly more variable, on the order of 10–120 m. Despite a negative geodetic net mass balance derived from the digital elevation models, the reservoir areas have been gaining mass since the surge terminations. This surface thickening along with considerable ablation of the receiving areas will most likely return the glacier surface profiles to the pre-surge stage. Our results indicate that (a) greatest surface thinning in the upper reservoir areas of Drangajökull rather than proximal to the equilibrium line during Vatnajökull surges and (b) development of Drangajökull surges that resembles Svalbard surge-type glaciers rather than Vatnajökull surge-type glaciers. The contrasting surge characteristics could be explained by differences in glacier geometry, topography and substratum of the Drangajökull and Vatnajökull surge-type glaciers.

1. Introduction

Surge-type glaciers are different from non-surging glaciers because they do not respond directly to climatic changes. Therefore, surge-type glaciers are not considered good indicators of climate fluctuations on timescales of decades or perhaps centuries (Yde and Paasche, 2010). They are characterised by cyclic flow instability in which the glacier undergoes quasi-periodic fluctuations (Benn and Evans, 2010). Occasionally, there is a sudden period of fast flow, the surge, which takes from few months to several years. Between surges, there is a quiescent phase that can last from a few decades to hundreds of years (Meier and Post, 1969; Raymond, 1987; Harrison and Post, 2003). Surge-type glaciers can be warm-based or polythermal, and they cluster in certain areas such as Iceland, Svalbard, Novaya Zemlya, parts of West and East Greenland, Arctic Canada, Alaska–Yukon and mountain ranges in central Asia, most commonly where climatic conditions are characterised by mean annual temperature of 0–10 °C and annual precipitation of 200–2000 mm (Sevestre and Benn, 2015), indicating that climate alone does not control their location (Meier and Post, 1969; Raymond, 1987; Hamilton and Dowdeswell, 1996; Jiskoot et al., 2000; Murray et al., 2003; Sevestre and Benn, 2015).

One surge cycle consists of an active phase and a quiescent phase. During the active phase, the glacier undergoes a dramatic change in geometry and morphology (Þórarinsson, 1964, 1969; Meier and Post, 1969; Sharp, 1988; Harrison and Post, 2003). The ice velocity during surge is commonly on the order of 2–3 magnitudes higher than during the quiescent phase. Ice from the reservoir area of the glacier is discharged down-glacier to the receiving area during the active phase (Dowdeswell et al., 1995; Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005; Quincey and Luckman, 2014). During the quiescent phase, the ice is stagnant or flowing at a velocity lower than required to maintain the glacier

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size. Snow and ice accumulation in the upper area and ice melting in the lower area of the glacier gradually contributes to a steeper surface profile which is considered fundamental to return the glacier surface to the pre-surge state and subsequently enable a new surge (Raymond, 1987; Sharp, 1988; Dowdeswell et al., 1995; Eisen et al., 2001, 2005; Harrison and Post, 2003; Ingólfsson et al., 2016).

There are three basic models for the mechanics of rapid flow during surges: (1) basal motion where decoupling of a glacier from its bed enables fast ice flow through enhanced basal sliding across the ice/bed interface or very shallow subglacial deformation (Engelhardt and Kamb, 1998); (2) basal motion where fast ice flow is sustained by deformation of water-saturated subglacial sediment that is strongly coupled to the glacier (Alley et al., 1989; Bennett, 2003); (3) subglacial deformation as one primary mechanism for sustaining rapid ice flow and high sediment discharge where rapid ice flow is sustained by over-pressurised water causing decoupling at the bedrock beneath a thick sediment sequence that is coupled to the glacier (Kjær et al., 2006). It has further been proposed that deformation of subglacial sediment may impact the periodicity of surge-type glaciers (Boulton and Hindmarsh, 1987; Evans and Rea, 1999; Murray et al., 2003).

Considerable volumes of ice can be transported down-glacier during a surge, which often contributes to surface lowering in the order of 20–100 m in the reservoir areas and surface thickening of similar amounts in the receiving areas (Björnsson et al., 2003; Aðalgeirsdóttir et al., 2005; Sund et al., 2014). Such surges usually contribute to marginal advances of several hundreds of meters or even kilometres (Meier and Post, 1969; Raymond, 1987; Björnsson et al., 2003; Murray et al., 2003; Sund et al., 2009). Recent reviews of the glaciology and glacial geology of surging glaciers in Iceland are provided by Björnsson et al. (2003) and Ingólfsson et al. (2016).

Icelandic surge-type glaciers are both soft- and hard bedded, they overly a variety of volcanic substrates and in many cases geothermal areas. However, no specific relationship has been noticed between their location and substratum or geothermal areas (Björnsson et al., 2003). The active phase of the large Icelandic surge-type glaciers usually last for only a few months or 1–2 yrs (Sigurðsson, 1998; Björnsson et al., 2003; Ingólfsson et al., 2016). In previous studies all Icelandic glaciers are considered to be warm-based and therefore any surge mechanism related to a thermal transition has been ruled out (Björnsson et al., 2003).

Absence of end moraines, lateral moraines, and annual moraines and little glacial geomorphological imprints on the plateau that have been deglaciated and not affected by the glacial rivers since the Little Ice Age (LIA) about 500–650 m a.s.l. around the southern perimeter of Drangajökull, might indicate polythermal conditions during the LIA and perhaps at present (Brynjólfsson et al., 2014). This might reflect a thin, frozen glacier margin having limited influence on its forefield and substratum, or contrasting sedimentary and geomorphological conditions of the low-relief topography around the southern perimeter of Drangajökull, compared to the surge-type outlet glaciers draining the northern half of Drangajökull (Brynjólfsson et al., 2014).

Notably, while the glacier advance during the two most recent Drangajökull surges lasted 5–7 yrs (Sigurðsson, 1998; Björnsson et al., 2003; Brynjólfsson et al., 2015a), the active phase of the large Icelandic surge-type glaciers usually last for only a few months or 1–2 yrs (Sigurðsson, 1998; Björnsson et al., 2003; Ingólfsson et al., 2016). This relatively long surge phase of Drangajökull resembles the 3–4 yrs long active phase of surge-type cirque glaciers in Iceland (Brynjólfsson et al., 2012) and 3–10 yrs long active phase of Svalbard surge-type glaciers which has been considered strongly related to their polythermal conditions (Dowdeswell and Hamilton, 1991; Hamilton and Dowdeswell, 1996; Murray et al., 2003; Hambrey and Glasser, 2012; Sevestre et al., 2015).

The reason for Drangajökull outlets displaying surge behaviour so different from surging glaciers of the other Icelandic ice caps remains to be fully understood. Bedrock substrate or flow rate has hitherto not been considered obvious variables controlling surge frequencies during the surge phase of Icelandic surge-type glaciers, although Brynjólfsson et al. (2015a) suggested it could not be excluded that the Miocene basalts below Drangajökull, which are different from the predominantly Pliocene–Pleistocene bedrock of other Icelandic ice caps, could at least partly explain the surge behaviour. Svalbard surging glaciers, recently suggested to be over 700 in number (Farnsworth et al., 2016), do occur on a variety of subglacial lithologies, from igneous and metamorphic Precambrian–Paleozoic basement rocks to sedimentary and igneous rocks of Paleozoic–Cenozoic age (Jiskoot et al., 1998; Murray et al., 2003; Farnsworth et al., 2016), but generally share the characteristics of long surge cycles and duration of the active surge phase. The geothermal heat flux below Drangajökull is comparable to the heat flux below Brúarjökull and Múlajökull (a very active surging outlet of Hofsjökull), estimated to be between 100–200 mW/m^2 (Hjartarsson, 2015), so that the very different surge dynamics and kinematics can probably not be explained by differences in geothermal heat flux. Recent studies of Drangajökull have focused on geomorphology, glacial history since the Last Glacial Maximum (LGM), surge dynamics and recent areal changes of the ice cap (Principato, 2003, 2008; Principato et al., 2006; Þrastarson, 2006; Brynjólfsson et al., 2014, 2015a, 2015b). Differencing of Digital Elevation Models (DEMs) is a well-established methodology to quantify volume changes of glaciers (e.g., Magnússon et al., 2005, 2015; Schomacker and Kjær, 2007, 2008; Sund et al., 2009, 2014, Abermann et al., 2010; Kjær et al., 2012; Schomacker et al., 2012; Jóhannesson et al., 2013). Time series of DEMs and other remotely sensed data are also commonly used to identify glacier surges and quantify their velocity, surface, volume and areal changes during the surges (Fischer et al., 2003; Magnússon et al., 2005, 2015; Frappé and Clarke, 2007; Sund et al., 2009, 2014; Quincey et al., 2011).

Shuman et al. (2009) compared a GPS derived Digital Elevation Model (DEM) with series of repeated satellite profiles across Drangajökull, indicating up to 1.5 $\mathrm{m\,a^{-1}}$ surface lowering at the location of the satellite profile in the years 2003–2007. However, ablation stake measurements indicate positive mass balance of the whole ice cap in 2005–2007, indicating that the satellite profile is not representative for the whole ice cap (Shuman et al., 2009).

Comparison of DEMs since c. 1990 and from 2011, indicates about 8 m average surface lowering of the ice cap in the period 1990–2011 (Jóhannesson et al., 2013). Obvious ice discharges during the most recent surges of the three outlet glaciers are reflected as much more thinning of their reservoirs and distinct thickening of the receiving areas of each outlet glacier as described by Jóhannesson et al. (2013). Recently, Magnússon et al. (2015) calculated the geodetic mass balance of Drangajökull in six time steps back to 1946 based on DEMs from aerial photographs. They demonstrated a mean mass balance rate of the ice cap of -0.25 ± 0.04 $\mathrm{m\,w.e.\,a^{-1}}$ in the period 1946–2011. However, they also observed high decadal variability in the mass balance with, e.g., a positive mass balance rate of 0.07 ± 0.08 $\mathrm{m\,w.e.\,a^{-1}}$ in 1975–1985 and 0.26 ± 0.11 $\mathrm{m\,w.e.\,a^{-1}}$ from 1985–1994. Thus, the mass balance of Drangajökull does not appear to follow exactly the increasing temperature trend in the Arctic (Miller et al., 2010). However, on longer timescales, Drangajökull is thinning and its surface area has decreased since the LIA and during the last decades (Jóhannesson et al., 2013; Brynjólfsson et al., 2014, 2015a; Magnússon et al., 2015).