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Volcanic impacts on modern glaciers: a global synthesis

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Abstract

Volcanic activity can have a notable impact on glacier behaviour (dimensions and dynamics). This is evident from the palaeo-record, but is often difficult to observe for modern glaciers. However, documenting and, if possible, quantifying volcanic impacts on modern glaciers is important if we are to predict their future behaviour (including crucial ice masses such as the West Antarctic Ice Sheet) and to monitor and mitigate glacio-volcanic hazards such as floods (including jökulhlaups) and lahars. This review provides an assessment of volcanic impacts on the behaviour of modern glaciers (since AD 1800) by presenting and summarising a global dataset of documented examples. The study reveals that shorter-term (days-to-months) impacts are typically destructive, while longer-term (years-to-decades) are more likely protective (e.g., limiting climatically driven ice loss). However, because these events are difficult to observe, particularly before the widespread availability of global satellite data, their frequency and importance are likely underestimated. The study also highlights that because the frequency and nature of volcano-glacier interactions may change with time (e.g., glacier retreat may lead to an increase in explosive volcanic activity), predicting their future importance is difficult. Fortunately, over coming years, continued improvements in remotely sensed data will increase the frequency, and enhance the quality, of observations of volcanic impacts on glaciers, allowing an improved understanding of their past and future operation.

Keywords

Glaciers; volcanoes; volcanic impacts; glacier dimensions; glacier dynamics, hazards.

1. Introduction

Climate exerts a first-order control on the behaviour (i.e., dimensions and dynamics) of modern glaciers. For example, glaciers typically grow in response to reduced atmospheric temperatures and/or increased solid precipitation (Cook et al., 2005; Bolch, 2007). Despite this, other non-climatic factors also play a role. One notable example is volcanic activity, which can directly affect glacier behaviour, and/or modulate glacier response to climate forcing (Major and Newhall, 1989; Chapman et al., 2000; Kirkbride and Dugmore, 2003).

Volcanic activity is known to impact climate, which in turn regulates glacier behaviour (e.g., Hammer et al., 1980; Rampino and Self, 1993; McConnell et al., 2017; Cooper et al., 2018); however, the focus here is on the more direct volcanic impacts on glaciers. These include instances where glacier behaviour has been directly affected by subglacial heating, subglacial dome growth, subglacial eruptions, lava flows (supraglacial and subglacial), supraglacial pyroclastic density currents, supraglacial tephra deposition, floods and lahars, and the supraglacial deposition of other glacio-volcanic products. These types of interactions have received considerable interest since the 2010 subglacial eruption of Eyjafjallajökull, Iceland (Gudmundsson et al., 2012; Sigmundsson et al., 2010). This has included a focus on glacio-volcanic activity on Mars (Scanlon et al., 2014), , and consideration of the role of subglacial volcanic and geothermal activity in governing the future stability of ice sheets (de Vries et al., 2017; Iverson et al., 2017; Seroussi et al., 2017) (Section 3.3.). This latter aspect has received considerable attention over recent years due to the possibility that future subglacial volcanic activity might change the bed conditions of the West Antarctic Ice Sheet, potentially triggering, or contributing to, its rapid collapse and global sea level rise (Blankenship et al. 1993; Vogel et al. 2006; Corr and Vaughan 2008; de Vries et al., 2017). Interactions between volcanoes and glaciers have also developed as an area of interest because of their potential to result in hazards, including floods, lahars, and other debris flows, which have had devastating impacts on mountain communities globally during recent centuries (Blong, 1984; Pierson et al., 1990; Chernomorets et al., 2007; Tuffen, 2010). Finally, observed changes in glacier behaviour can serve as useful indicators of, and precursors to, periods of volcanic activity (e.g., observed ice loss was an early precursor to the 2009 eruption of Mount Redoubt, Bleick et al., 2013). Thus, monitoring, documenting, and, if possible, predicting volcanic impacts on glaciers is of global scientific and socio-economic importance (Pierson et al., 1990; Mazzocchi et al., 2010).

Instances where volcanic activity directly affects glacier behaviour (i.e., the focus of this paper) typically occur either because glaciers are located on active volcanoes, or because volcanic products (e.g., ash/tephra) interact with glaciers in adjacent (sometimes non-volcanic) regions. Though evidence for past volcanic impacts on glaciers is seen in the palaeo-record (Smellie and Edwards, 2016), establishing their importance for modern glaciers is challenging because glacio-volcanic regions are often remote and inaccessible. Thus, until recently, volcanic impacts on modern glaciers were rarely directly observed and were poorly understood. Fortunately, increased attention on glacio-volcanic regions, facilitated by rapid developments in remote sensing, has led to repeat observations and monitoring programs that are now elucidating some key aspects of volcano-glacier interactions (Curtis and Kyle, 2017). Major and Newhall (1989) made an early, and important, contribution to recognising the global impact of volcanic activity on snow and ice, with a particular focus on floods and lahars. Delgado Granados et al. (2015) provided a review of recent volcano-glacier interactions in South and Central America, and Smellie and Edwards (2016) provided a global review of volcano-glacier interactions, mainly (though not exclusively) focused on the palaeo (Quaternary) record. The present paper builds on these overviews by focusing specifically on the glaciological consequences of volcanic activity globally since AD 1800. Here, the term 'volcanic activity' is used to encompass explosive and effusive volcanic eruptions, as well as released geothermal energy and landscape changes (e.g., dome growth) induced by deep volcanic sources at active or dormant volcanoes.

2. Observed volcanic impacts on modern glaciers

Here, we outline how volcanic activity can directly influence the behaviour of modern glaciers. As alluded to in section 1., the volcanic processes considered are subglacial heat flow; subglacial dome growth, subglacial eruptions, lava flows (supraglacial and subglacial), supraglacial pyroclastic density currents, supraglacial tephra deposition, floods and lahars, and the supraglacial deposition of other glacio-volcanic products. Below we outline how each mechanism can affect glaciers, with reference to a global dataset of examples. This structure means that we focus on different mechanisms (and their glaciological impacts) discretely, while in reality (e.g., during a single eruption) many of the mechanisms likely act in conjunction (sometimes triggering one another). This can make it difficult to isolate or quantify the specific glaciological impact of a particular mechanism, and the fact that mechanisms may be operating in conjunction (with a combined glaciological impact) should be kept in mind throughout.

The locations of volcanoes mentioned in the text are illustrated in Fig. 1, and outlined in Table 1. Detailed information about each period of volcanic activity and associated glaciological consequences is presented in Supplementary Table 1 (alongside relevant citations), in a kmz. file for viewing and editing in Goole Earth™ (Supplementary data 1), and interactions are schematically illustrated in Fig. 2. We do not consider indirect impacts, such as glacier growth in response to volcanically triggered climatic cooling (e.g., Hammer et al., 1980; Rampino and Self, 1993; McConnell et al., 2017; Cooper et al., 2018), and do not directly consider seismic impacts on glaciers, though in many cases, seismic and volcanic activity may be linked.

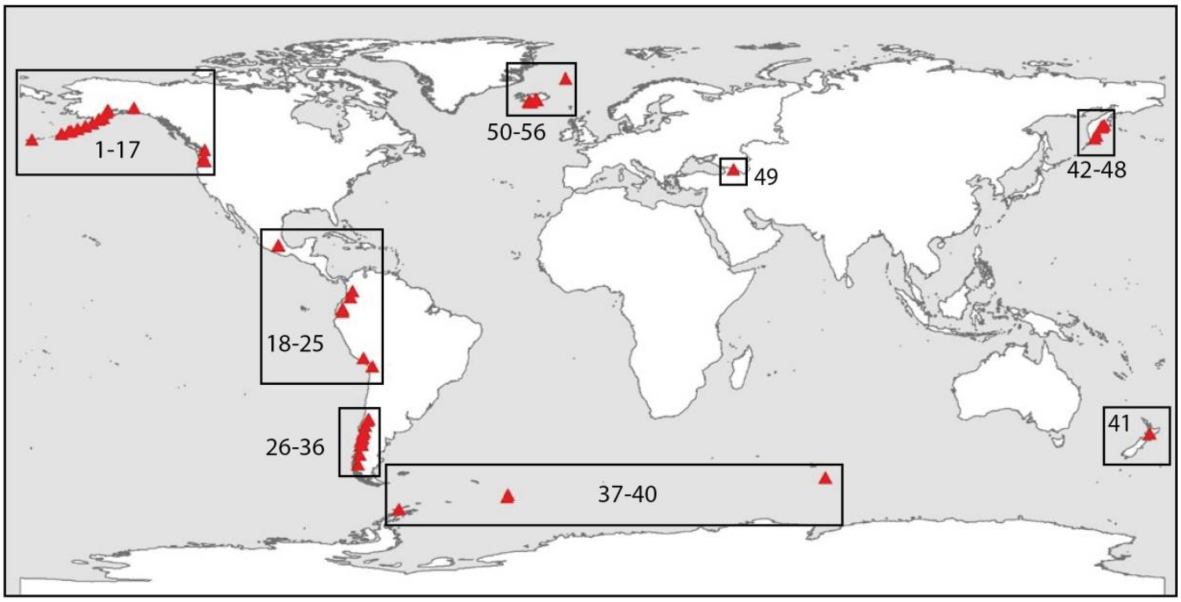

Fig. 1. Red triangles showing locations where the behaviour of modern (post AD 1800) glaciers has been affected by volcanic activity. Numbers refer to specific volcanoes, detailed in Table 1 and Supplementary Table 1 (where volcanic events and associated glaciological consequences are also described).

Table 1. Volcanoes and periods of volcanic activity discussed in this review. Detailed information is provided in Supplementary Table 1.

Volcano number	Volcano name	Location	Lat, Lon	Periods of activity
1	Great Sitkin	USA (Alaska)	52.08°N, 176.13°W	1945
2	Makushin	USA (Alaska)	53.89°N, 166.92°W	1983
3	Mount Westdahl	USA (Alaska)	54.52°N, 164.65°W	1978, 1991–92
4	Mount Shishaldin	USA (Alaska)	54.76°N, 163.97°W	1999
5	Mount Pavlof	USA (Alaska)	55.42°N, 161.89°W	2013
6	Mount Veniaminof	USA (Alaska)	56.17°N, 159.38°W	1983–84, 1993–95, 2013
7	Mount Chiginagak	USA (Alaska)	57.14°N, 156.99°W	2004–05
8	Novarupta	USA (Alaska)	58.27°N, 155.16°W	1912
9	Trident Volcanic group	USA (Alaska)	58.24°N, 155.10°W	1953–60
10	Mount Katmai	USA (Alaska)	58.26°N, 154.98°W	1912
11	Fourpeaked Mountain	USA (Alaska)	58.77°N, 153.67°W	2006
12	Mount Redoubt	USA (Alaska)	60.49°N, 152.74°W	1966–68, 1989–90, 2008–09
13	Mount Spurr (Crater Peak)	USA (Alaska)	61.30°N, 152.25°W	1953, 1992, 2004–06
14	Mount Wrangell	USA (Alaska)	62.00°N, 144.02°W	1899, 1964–ongoing, 1999
15	Mount Baker	USA (Cascade Arc)	48.78°N, 121.81°W	1958–76
16	Mount St Helens	USA (Cascade Arc)	46.20°N, 122.18°W	1980, 2004–06
17	Mount Hood	USA (Cascade Arc)	45.37°N, 121.70°W	1853–1869, 1907
18	Iztaccíhuatl	Mexico	19.18°N, 98.64°W	Late 20th century
19	Popocatépetl	Mexico	19.02°N, 98.63°W	1994–2001
20	Nevado del Ruiz	Columbia	4.90°N, 75.32°W	1985
21	Nevado del Huila	Columbia	2.93°N, 76.03°W	2007–12
22	Cotopaxi	Ecuador	0.68°S, 78.44°W	1877
23	Tungurahua	Ecuador	1.47°S, 78.44°W	1999–2001
24	Nevado Sabancaya	Peru	17.78°S, 71.85°W	1986–88, 1990–98
25	Volcán Guallatiri	Chile	18.42°S, 69.09°W	Late 20th Century
26	Tinguiririca	Chile	34.81°S, 70.35°W	1994, 2006/07
27	Volcán Peteroa (Planchón-Peteroa)	Chile	35.27°S, 70.58°W	1963–91, 1991, 2004–07, 2010–11
28	Nevados de Chillán	Chile	36.86°S, 71.38°W	1973–86
29	Volcán Llaima	Chile	38.69°S, 71.73°W	1979, 1994, 2008
30	Volcán Villarrica	Chile	39.42°S, 71.93°W	1971, 1984–85, various

31	Puyehue-Cordón Caulle	Chile	40.59°S, 72.12°W	2011
32	Volcán Calbuco	Chile	41.33°S, 72.61°W	1961
33	Volcán Michinmahuida	Chile	42.79°S, 72.44°W	2007–08
34	Volcán Chaitén	Chile	42.84°S, 72.65°W	2008
35	Volcán Hudson	Chile	45.90°S, 72.97°W	1971, 1991, 2011
36	Volcán Lautaro	Chile	49.02°S, 73.55°W	Various 20th Century
37	Deception Island	Sub-Antarctic	62.97°S, 60.65°W	1969
38	Bristol Island	Sub-Antarctic	59.04°S, 26.53°W	1935–1962
39	Mt Belinda	Sub-Antarctic (Montagu Island)	58.42°S, 26.33°W	2001–07
40	Mawson Peak	Sub-Antarctic (Heard Island)	53.11°S, 73.51°E	2006-08
41	Mount Ruapehu	New Zealand	39.28°S, 175.57°E	1995–96, 2007
42	Mutnovsky	Russia (Kamchatka)	52.45°N, 158.20°E	2000, ongoing
43	Avachinsky	Russia (Kamchatka)	53.26°N, 158.83°E	1945, 1991
44	Tolbachik	Russia (Kamchatka)	55.82°N, 160.38°E	1975–76, 2012–13
45	Bezymianny	Russia (Kamchatka)	55.98°N, 160.59°E	1955–57
46	Klyuchevskoy	Russia (Kamchatka)	56.06°N, 160.64°E	1944–45, 1953, 1966–68, 1977–80, 1982–83, 1984–85, 1985–86, 1986–90, 2005–10
47	Ushkovsky	Russia (Kamchatka)	56.07°N, 160.47°E	1959–60, 1982–84
48	Shiveluch	Russia (Kamchatka)	56.65°N, 161.36°E	1964
49	Mount Kazbek	Russia/Georgia (Caucasus)	42.70°N, 44.52°E	Various 20th and 21st Century (possible)
50	Beerenberg	Norway (Jan Mayen Island)	71.08°N, 8.16°W	1970–72
51	Grímsvötn	Iceland	64.42°N, 17.33°W	1934, 2004, 2011
52	Gjálp fissure	Iceland	64.52°N, 17.39°W	1996
53	Bárðarbunga	Iceland	64.64°N, 17.53°W	2014
54	Katla	Iceland	63.63°N, 19.05°W	1918
55	Eyjafjallajökull/ Fimmvörðuháls;	Iceland	63.63°N, 19.62°W	2010
56	Hekla	Iceland	63.98°N, 19.70°W	1947

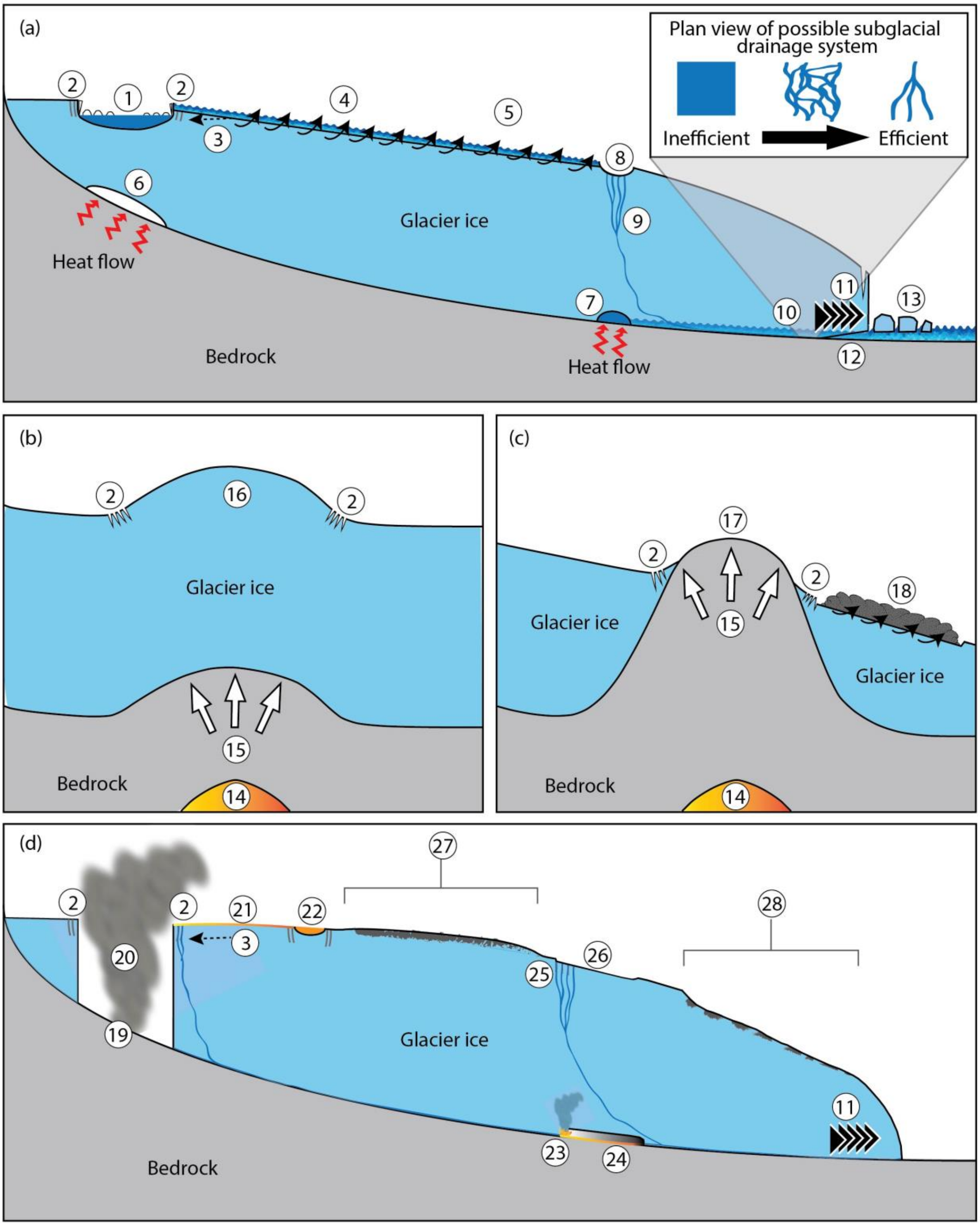

Fig. 2. Schematic illustrations of the impacts that different forms of volcanic activity can have on glacier behaviour. Different forms of activity include (a) enhanced subglacial heat flow and flood/lahars, (b) Subglacial volcanic dome growth, (c) volcanic dome extrusion and a pyroclastic density current, and (d) a subglacial volcanic eruption, lava flows, and supraglacial tephra deposition. Numbers in this figure refer to different processes or features. (1) Lake filled supraglacial cauldron,

with ice breaking from vertical cauldron walls. (2) Zones of crevassing and fracturing. (3) Local reversal in ice flow direction. (4) Supraglacial flood (formed by lake drainage) transitioning to a lahar. (5) Supraglacial channel/canyon formed by ice melt, erosion and entrainment (by a flood/lahar). (6) Subglacial cavity caused by enhanced heat flow. (7) Subglacial lake formed by enhanced heat flow. (8) Supraglacial depression formed above a subglacial lake. (9) Crevasses (above a subglacial lake) acting as a route for supraglacial meltwater drainage to the bed. (10) Subglacial meltwater drainage with inset plan view of different possible drainage styles. (11) Glacier advance/acceleration due to subglacial meltwater drainage (more widespread if drainage is inefficient). (12) Glacier front uplifted by subglacial flooding. (13) Ice blocks torn from a glacier front by subglacial flooding. (14) Magma upwelling. (15) Subglacial dome growth. (16) Glacier doming. (17) Dome extrusion through a glacier. (18) Pyroclastic density current (due to dome collapse) and associated supraglacial channel (formed by ice melt, erosion and entrainment). (19) Subglacial eruption. (20) Ice crater. (21) Supraglacial lava flow and associated melt channel. (22) Supraglacial lava ponding and associated crevasse bounded melt pit. (23) Site of small and/or phreatic eruption (with lava fountain). (24) Subglacial lava flow and associated melt cavity. (25) Crevasses (above the site of a small subglacial eruption) acting as a route for supraglacial meltwater drainage to the bed. (26) Supraglacial depression formed above a subglacial lava flow melt cavity. (27) Reduced melt under thick and/or continuous supraglacial tephra. (28) Increased melt under thin and/or discontinuous supraglacial tephra. These illustrations are not to scale, with some aspects exaggerated to highlight specific phenomena. For reasons of simplicity/clarity, the illustrations also depict glaciers with shallow surface gradients, whereas many volcano-occupying glaciers are steeper (though the processes shown here are likely to operate for both steep and shallow glaciers).

2.1. Enhanced subglacial heat flow

Enhanced subglacial heat flow (geothermal heating) (Fig. 2a) can occur without associated volcanic activity, or prior-to, during, or after periods of activity. The main glaciological consequence is subglacial melt. This can lead to the formation of subglacial cavities or lakes (Fig. 2a, points 6 & 7), which can, in turn, cause subsidence and fracturing of the ice surface (Gudmundsson et al., 1997)

(Fig. 2a, points 1 & 8). Increased subglacial melt can also cause glacier retreat (Salamatin et al., 2000) or advance/acceleration (Sturm et al., 1991; Rivera et al., 2012) (Fig. 2a, point 11). Here (below), we consider documented examples of these different outcomes.

2.1.1. Surface subsidence and fracturing

Instances where enhanced subglacial heat flow has caused the subsidence and fracturing of glacier surfaces are documented from many glacio-volcanic regions globally, including the USA, Chile, sub-Antarctic Islands, Kamchatka, and Mexico (Supplementary Table 1). In many cases, subglacial heating initially results in the formation of supraglacial melt pits/holes (Fig. 3a), which become enlarged to form cauldrons (Fig. 3b) as subglacial heating continues (e.g., Mount Wrangell 1964–ongoing; Mount Baker 1975–76; Mount Makushin 1983; Mount Redoubt 1989–1990, 2009; Mount Chiginagak 2004–05; Mount Spurr 2004–06). These cauldrons can be filled by lakes (Fig. 3c) (e.g., Mount Wrangell 1964–ongoing; Mount Baker 1975–76; Mount Veniaminof 1983–84; Mount Chiginagak 2004–05; Mount Spurr 2004–06), which often act to increase the cavity size as ice falls/calves from the vertical or overhanging ice walls (e.g., Mount Baker 1975–76; Mount Spurr 2004–06) (Fig. 2a, point 1). Lakes can entirely or partially drain, forming large englacial and/or subglacial channels, which result in further surface depressions and/or melt holes (e.g., Mount Spurr 2004–06). Lakes can also overflow, and generate supraglacial channels (Smellie, 2006). Sometimes, drainage events are large enough to trigger floods and lahars (Section 2.7.) (e.g., Mount Chiginagak 2004–05), though not always (e.g., Mount Baker 1975–76; Mount Veniaminof 1983–84).

Ice surrounding supraglacial ice cauldrons is often fractured, forming encircling arcuate/concentric crevasses (e.g., Mount Veniaminof 1983–84; Mount Spurr 2004–06) (Fig. 3d; Fig. 2a, point 2). These crevasses are partly a reflection of localised reversals in ice flow direction, as ice begins to flow back towards the cauldrons, perpetuating melt (Fig. 2a, point 3). In some cases, enhanced subglacial heat flow can result in the formation of ice surface fissures (e.g., Deception Island 1969; Volcán Iztaccíhuatl during the late 20th century). These can be an important route for supraglacial water to drain to the glacier bed (Fig. 2a, point 9), where it accumulates along with subglacially-derived meltwater and potentially promotes ice advance and/or acceleration (Section 2.1.3.).

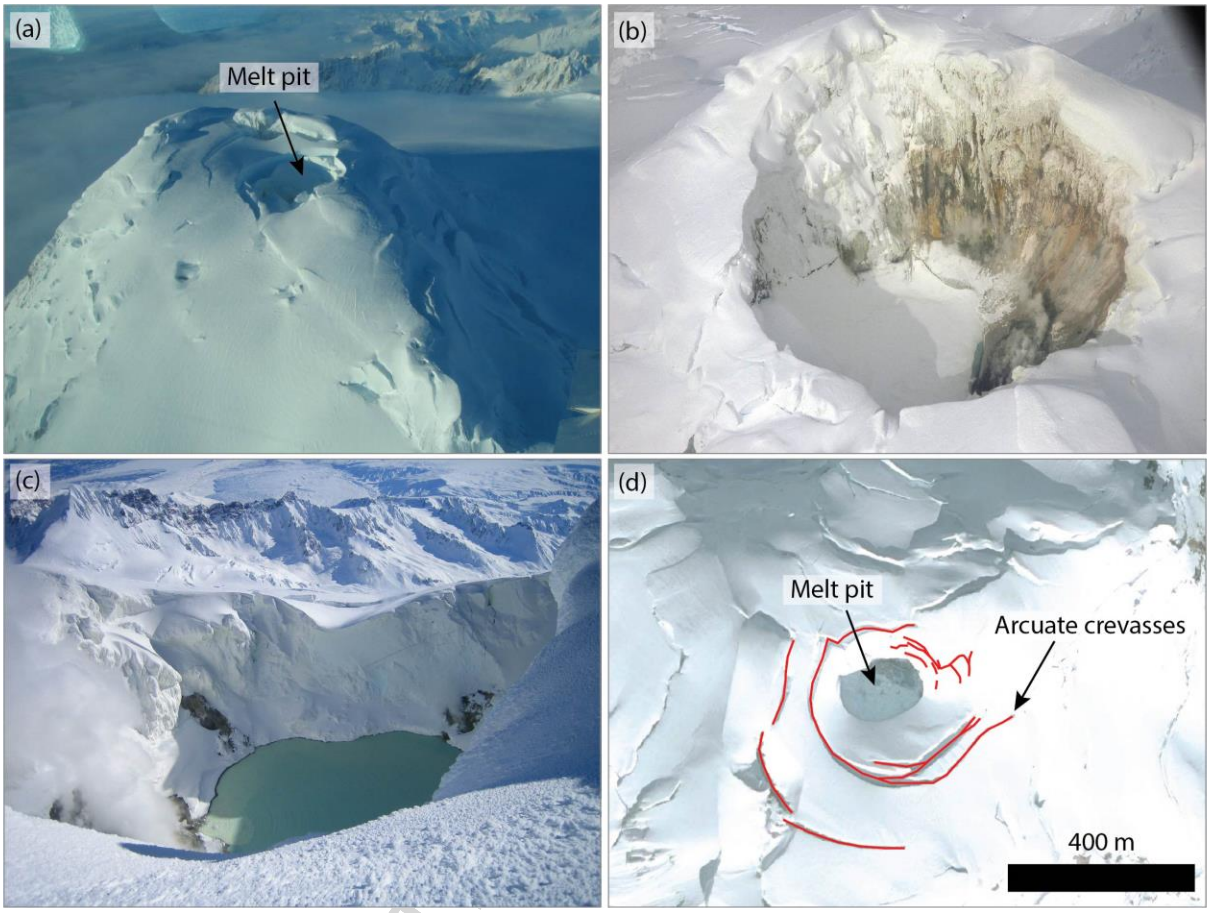

Fig. 3. Evidence of ice surface subsidence and fracturing during the 2004–06 period of enhanced subglacial heat flow at Mount Spurr. In this example, a summit melt pit (a), enlarged to form a summit ice cauldron (b), occupied by a lake (c). During formation of the melt pit, adjacent ice became fractured, with encircling arcuate crevasses (marked as red in ‘d’). (a) Photograph taken by M.L. Coombs (AVO/USGS) on October 30, 2004. Image obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=5). (b) Photograph taken by R.G. McGimsey (AVO/USGS) on June 28, 2007. Image obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=13305). (c) Photograph taken by N. Bluett (AVO/USGS) on March 22, 2006. Image obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=9657). (d) DigitalGlobe™ image, taken on August 11, 2004, viewed in GoogleEarth™.

2.1.2. Glacier retreat

Instances where enhanced subglacial heat flow has caused observable glacier retreat are comparatively rare (Supplementary Table 1). One example comes from Volcán Villarrica, where, over recent years, Pichillancahue-Turbio Glacier retreated at a faster rate than others in the region, partly due to enhanced geothermal heating (Rivera et al., 2012). A second example comes from Volcán Iztaccíhuatl, where geo- and hydro-thermal heat flow caused accelerated melt and glacier retreat during the late 20th century (Delgado Granados et al., 2005). For example, Ayoloco Glacier experienced a rapid increase in the rate of area loss (i.e., between 1958 and1982 the glacier lost 12% of its area, but between 1982 and 1998 lost 43%), and Centro Oriental Glacier almost entirely disappeared due to geothermally-driven increased melt over this period (Delgado Granados et al., 2005).

2.1.3. Glacier advance and/or acceleration

Instances where enhanced subglacial heat flow has caused (or appears to have caused) glacier advance and/or acceleration are documented in Alaska, Chile, Kamchatka, and the Caucasus (Supplementary Table 1). For example, at Mount Wrangell, there was increased heat flux following a major regional earthquake in 1964. As a result, the three glaciers which emanate from the volcano's North Crater (Ahtna Glacier, and South and Centre MacKeith Glaciers) have advanced since 1965 (at a rate of 5–18 m a^{-1}), unlike others on the volcano or elsewhere in the Wrangell Mountains (Sturm et al., 1991). It is assumed that this advance resulted from meltwater, which drained down the northeast flank of the volcano and lubricated the glacier bed (Sturm et al., 1991). These glaciers have also come to show little seasonal variation in their surface velocity, unlike most glaciers not subject to volcanic impacts (Iken and Bindschadler, 1986; Bartholomew, 2011), thus supporting the idea that volcanically-produced meltwater is driving changes in flow conditions (Sturm et al., 1991). At Volcán Peteroa, subglacial geothermal heating before phreatomagmatic explosions in 1991 and 2010 caused subglacial melt, increased basal sliding, and glacier advance during the 1963–1990 and 2004–2007 periods (Liaudat et al., 2014). At Volcán Michinmahuida, subglacial geothermal heating a few months prior to the 2008 eruption of Volcán Chaitén (~ 15 km to the west) is thought to have caused glacier advance and acceleration (Rivera et al., 2012). For example, Glaciar Amarillo retreated ~ 76 m yr^{-1} between 1961 and 2007, but advanced 243 ± 49 m between November 2007 and September 2009 (coinciding with the period of subglacial geothermal heating), after which, glacier retreat resumed (Rivera et al., 2012). At Ushkovsky**,** strengthening of seismic (and perhaps volcanic) activity is thought to have caused Bilchenok Glacier (which emanates from the NW corner of the ice-filled caldera) to advance by 1050–1150 m and 700–800 m in 1959–1960 and 1982–84, respectively (Muraviev et al., 2011, 2012; Muraviev and Muraviev, 2016). Finally, at Mount Kazbek, periods of increased subglacial volcanic and/or geothermal activity may have caused the acceleration, advance, and destabilisation of local glaciers at various periods during the 20^{th} and 21^{st} centuries (Chernomorets et al., 2007).

While the periods of glacier advance/acceleration outlined above likely occurred due to enhanced meltwater accumulation at the ice-bed interface, which reduced basal drag and promoted basal sliding and advance, the exact cause of these events is often unclear. In many cases, multiple causes may have acted together. For example, the advance of Glaciar Amarillo at Volcán Michinmahuida between November 2007 and September 2009 may have been caused by combination of increased subglacial heating and supraglacial tephra deposition (Rivera et al., 2012).

2.1.4. Overall glaciological impacts of enhanced subglacial heat flow

The most common and conspicuous glaciological impact of enhanced subglacial heat flow is the subsidence and fracturing of glacier surfaces (Fig. 2a, points 1, 2 & 8). This occurs in response to the formation of subglacial melt-cavities and lakes (Fig. 2a, points 6 & 7). Subsidence and fracturing not only reflect changes in glacier geometry, but potentially have additional, indirect, impacts on glacier behaviour. In particular, sites of supraglacial subsidence can cause local reversals in ice flow direction (Fig. 2a, point 3), and fracture zones are a potential route for meltwater drainage to the glacier bed (Fig. 2a, point 9). The accumulation of meltwater at the glacier bed (from subglacial or supraglacial sources) is potentially the most important glaciological consequence of enhanced subglacial heat flow, since basal lubrication can facilitate glacier advance and/or acceleration (Fig. 2a, points 10 & 11). Documented examples of advance/acceleration in response to enhanced subglacial heating are certainly more common than examples of glacier retreat. However, making clear (unequivocal) links between subglacial heating and changes in glacier behaviour is difficult, since geothermal heating is often poorly monitored and subglacial environments are notoriously difficult to observe.

2.2. Subglacial dome growth

Magma upwelling can result in ground deformation, and the growth of (often hot) lava domes (Melnik and Sparks, 1999) (Fig. 2b & c, points 14 & 15). Periods of dome growth can occur without associated volcanic activity, or before, during, or after periods of activity. Subglacial dome growth can cause deformation and fracturing of the ice surface (Fig. 2b & c, points 2 & 16). In some cases, lava domes can extrude through the overlying ice (i.e., they become subaerial), where they cause further glacier displacement and fracturing (Fig. 2c, points 2, 15 & 17). Some extruded domes are also susceptible to gravitational collapse and/or destruction by explosions (dome growth and destruction can occur repeatedly), resulting in (hot) supraglacial pyroclastic density currents (Fig. 2c, point 18) (Section 2.5.). Here (below), we consider documented examples of subglacial dome growth (both with and without subsequent extrusion) and associated glaciological impacts.

2.2.1. Subglacial dome growth, ice deformation and fracturing

There are two notable examples of glacier deformation due to subglacial dome growth, without extrusion through the overlying ice (Supplementary Table 1). The first is from Mount St Helens, when, prior to the major 1980 eruption, minor eruptions and bulging resulted in crevassing and ice avalanches on overlying glaciers (Brugman and Post, 1981). More recently, at Nevado del Huila in 2007–12, subglacial domes formed between the South and Central peaks, and caused deformation of the overlying glacier surface (Delgado Granados et al., 2005).

2.2.2. Dome extrusion and glacier displacement

There are only two documented examples of volcanic domes extruding through overlying glaciers (Supplementary Table 1). The earliest observation, based on a single photograph, comes from Great Sitkin, where, in 1945, a dome formed beneath, and then emerged through, the caldera glacier, resulting in a melt hole surrounded by bulged and crevassed ice (Simons and Mathewson, 1955). A better-documented example comes from Mount St Helens, where, following the major 1980 eruption, the ice-free summit crater became occupied by a ~ 1 km^2, up to 200 m thick glacier (Schilling et al., 2004; Walder et al., 2008). An eruption beneath this glacier in 2004–06 resulted in the formation of solid lava spines (parts of a subglacial dome), which extruded through the ice in the southern part of the Crater Glacier (Walder et al., 2007, 2008, 2010). As a result, the glacier was split into two parts, East Crater Glacier and West Crater Glacier, which were then squeezed between the growing lava dome and the crater walls (Walder et al., 2007, 2008). Because of this squeeze, the surfaces of the two glaciers buckled, forming multiple crevasses, and both glaciers locally doubled in thickness (at a rate of 0.6 m d^{-1}) (Walder et al., 2008). During this period, associated ice melt was limited, though both glaciers lost some volume. Since dome growth has stopped (in 2006), the glaciers have thinned in their upper reaches, and thickened in their lower (as ‘normal’ flow has resumed and ice has been redistributed downslope), and the terminus of East Crater Glacier has advanced (Walder et al., 2008).

2.2.3. Overall glaciological impacts of subglacial dome growth

The glaciological implications of subglacial dome growth are likely more important, and more conspicuous, for small/thin glaciers (e.g., mountain glaciers) than for large ice masses (e.g., continental ice sheets). Documented examples of domes deforming glaciers remain comparatively rare, and this is particularly true of dome extrusion. Despite this, in cases of subglacial dome growth and extrusion, the glaciological consequences can be extreme. This is most clearly documented at Mount St Helens, where the behaviour of a developing crater glacier(s) was severely disrupted by dome extrusion in 2004–06 (Section 2.2.2). The fracturing of ice surfaces in response to subglacial dome growth might also have indirect implications for glacier behaviour, since meltwater pathways to the glacier bed are potentially opened. Despite this, we are not aware of documented examples of glacier advance or acceleration in response to dome growth.

2.3. Subglacial eruptions

Subglacial volcanic eruptions (explosive or effusive) can have both thermal and mechanical impacts on overlying glaciers (Fig. 2d), leading to (i) the formation of ice craters, and associated fractures (Fig. 2d, points 20 & 2); (ii) partial glacier destruction; (iii) complete glacier destruction; and (iv) glacier advance/acceleration (Fig. 2d, point 11). Documented examples of these different impacts are considered below.

2.3.1. Ice craters and fractures

The formation of ice craters (Fig. 4a & Fig 2d, point 20) is perhaps the most common glaciological consequence of subglacial volcanic eruptions. Examples have been documented in many glacio-volcanic regions globally, including Alaska, Chile, Columbia, Iceland, Kamchatka and in the Sub-Antarctic Islands (Supplementary Table 1). These craters are typically hundreds of metres deep and wide, and reflect notable ice loss. For example, between the 14th and 20th of April 2010, ~10% (~0.08 km^3) of the pre-eruption caldera ice at Eyjafjallajökull was destroyed by crater formation (Magnússon et al., 2012). As with ice cauldrons (Section 2.1.1.), craters are often surrounded by concentric crevasses, reflecting local reversals in ice flow direction (e.g., Nevado del Ruiz 1985; Volcán Hudson 1991, 2011; Gjálp 1996) (Fig. 4a; Fig. 2d, point 3), but not where the ice is comparatively thin (e.g., Eyjafjallajökull 2010).

Above less explosive vents and bedrock fissures (particularly those experiencing phreatic eruptions) (Fig 2d, point 23), smaller melt-pits and ice-fissures can form, often surrounded by heavily crevassed and deformed ice (e.g., Mount Westdahl 1991–92; Fourpeaked Mountain 2006) (Fig. 4b). Ice fissures can be kilometers long and hundreds of metres wide. They are a potential route for supraglacial water to drain subglacially (Section 2.1.1.) (Fig. 2d, point 25), but are also a means by which subglacial or englacial meltwater (which is often heated) can emerge at the surface and contribute to further supraglacial melt (potentially triggering lahars and/or floods) (e.g., Nevado del Huila 2007–12). The heavy fracturing of glaciers during subglacial eruptions can also leave them susceptible to subsequent erosion if, for example, they are later swept by pyroclastic density currents (Section 2.5.). Notable examples of heavily crevassed ice surfaces due to subglacial eruptions include the upper section of Chestnina Glacier following suspected volcanic activity at Mount Wrangell in

1999 (McGimsey et al., 2004); the southern part of the continuous ice rim at Mount Spurr, where ice was eroded into pinnacles following an eruption in 1953 (Juhle and Coulter, 1955); and the glacier on the west flank of Nevado del Huila as a result of eruptions in 2007–12 (Worni et al., 2012). In some cases, it has been suggested that heavy crevassing reflects volcanically induced glacier advance, even though the behaviour itself may not have been observed (e.g., on the upper section of Chestnina Glacier at Mount Wrangell in 1999).

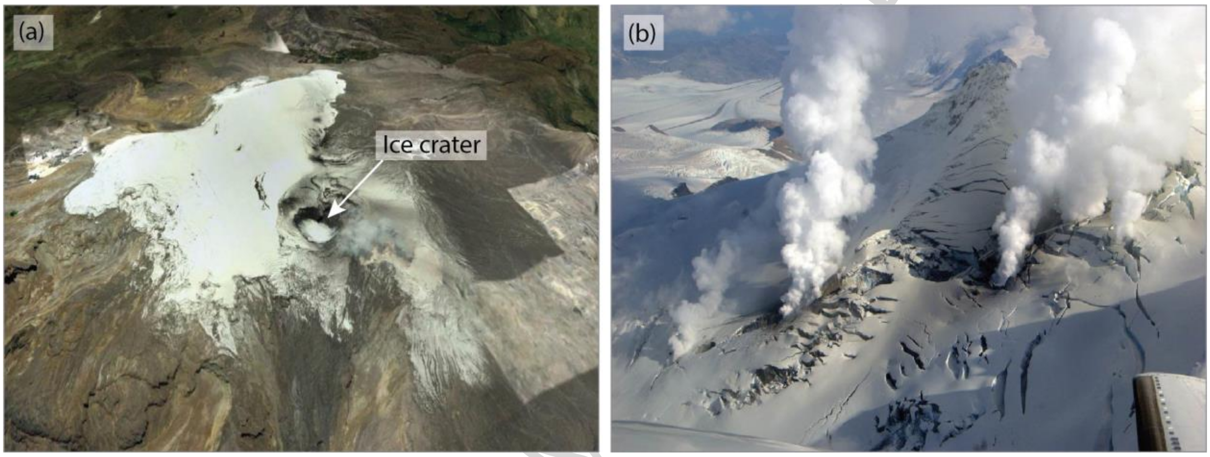

Fig. 4. Consequences of subglacial volcanic eruptions. (a) An ice crater (Arenas Crater) at Nevado del Ruiz, which opened repeatedly during multiple late 20th century eruptions. DigitalGlobe™ image, taken on November 10, 2015, viewed in GoogleEarth™. (b) Supraglacial crevasses, ice fissures and melt pits formed in 2006 above active subglacial vents at Fourpeaked Mountain. Photograph taken by C. Read (AVO/USGS) on September 24, 2006. Image obtained from the AVO/USGS database (http://www.avo.alaska.edu/image.php?id=11205). Description based on Neal et al. (2009).

2.3.2. Partial glacier destruction

The partial destruction of glaciers during subglacial volcanic eruptions is relatively common (Supplementary Table 1), and typically involves glacier beheading (i.e., the destruction of part of a glacier's accumulation zone). For example, White River Glacier, now Coalman Glacier, was partially beheaded during an eruption of Mount Hood between 1894 and 1912 (Lillquist and Walker, 2006). Shoestring, Forsyth, Ape, and Nelson Glaciers were beheaded during the 1980 eruption of Mount St

Helens (Brugman and Post, 1981) (Fig. 5). The glacier draining the NW flank of Volcán Hudson was partially beheaded during an eruption in 1971, as 50–80% (60 km^2) of the volcano's intra-caldera ice was destroyed (Fuenzalida, 1976; Rivera and Bown, 2013). Tolbachinsky Glacier was partially beheaded (i.e., the surface area of intra-caldera ice reduced from 1.54 to 0.5 km^2) during the 1975–76 eruption of Tolbachik (Vinogradov and Muraviev, 1982), and the 1982–83 eruption of Klyuchevskoy partially beheaded Kellya Glacier (Vinogradov and Muraviev, 1985). Beheading of this type is often a direct result of volcanic blasts, but can also occur during summit collapse. A notable example comes from Alaska, where the eruption of Novarupta in 1912 led to the collapse of the glacier-clad summit of Mount Katmai (~ 9 km to the east) (Hildreth and Fierstein, 2000, 2003, 2012). This collapse formed a ~ 5.5 km^3 summit caldera (Fig. 6) and partially beheaded Metrokin Glacier and Knife Creek Glaciers 3 and 4 (Hildreth and Fierstein, 2012). This glacier beheading left ice cliffs surrounding the crater rim (these cliffs were effectively the upper-ends of each glacier), with ice avalanches frequently falling into the crater, where the ice soon melted (Hildreth and Fierstein, 2012). Over a period of decades, these ice cliffs slowly thinned and retreated from the caldera rim (due to ice flow and melting). Part of the icefield outside the caldera experienced a reversal in flow direction as an ice tongue advanced into the caldera, ultimately terminating at (and calving into) the caldera-occupying lake (Fig. 6b).

It is notable that despite dramatic changes to glacier accumulation areas during glacier beheading (particularly at Mount Katmai in 1912 and Mount St Helens in 1980), these events rarely result in quick observable glacier retreat, likely because associated deposits (including tephra) act to insulate glacier surfaces, leading to stagnation (Sections 2.6.2. and 2.8.1.). However, over the longer-term (i.e., over decades), some beheaded glaciers have retreated or disappeared entirely (e.g., Shoestring, Nelson, Forsyth, and Dryer Glaciers at Mount St Helens).

In addition to cases of beheading, subglacial eruptions have directly destroyed parts of glacier ablation zones (tongues). For example, the 1977–80 eruption of Klyuchevskoy destroyed part of Shmidta Glacier's tongue (Muraviev et al., 2010, 2011; Muraviev and Muraviev, 2016). However, such cases are very rare, likely because glacier tongues rarely overlie volcanic vents (since eruptions tend to occur near the summits of volcanic edifices).

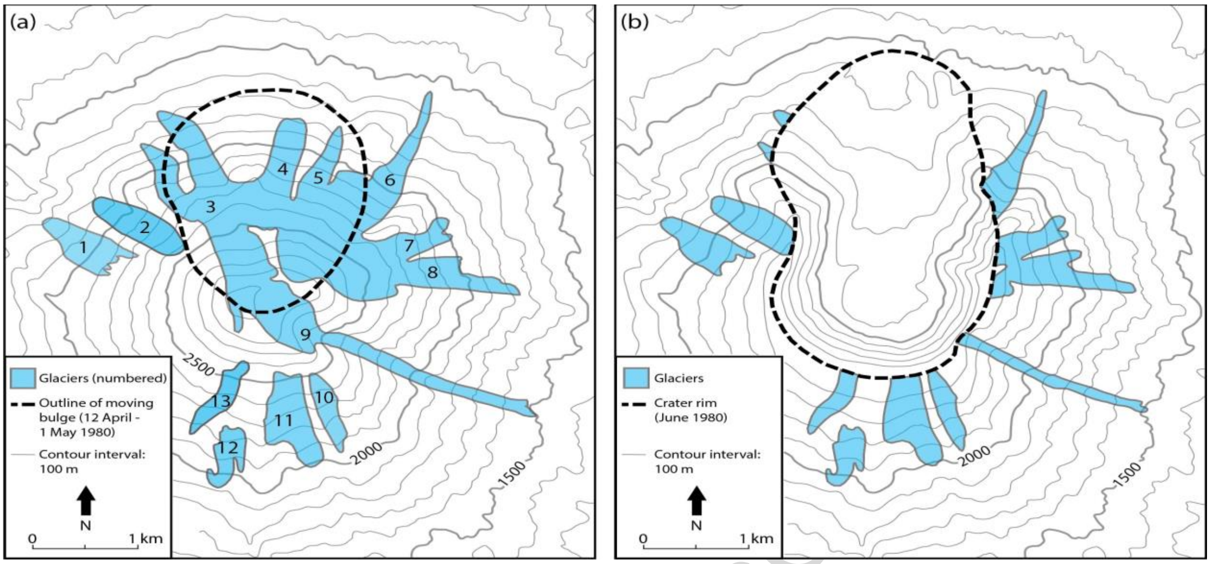

Fig. 5. Glaciers occupying Mount St. Helens (a) before, and (b) after the 1980 eruption. Numbered glaciers are: (1) Talus, (2), Toutle, (3) Wishbone, (4) Loowit, (5) Leschi, (6) Forsyth, (7) Nelson, (8) Ape, (9) Shoestring, (10) unnamed, (11) Swift, (12) unnamed, and (13) Dryer. The eruption beheaded some glaciers, and completely destroyed others. Figure based on Brugman and Post (1981).

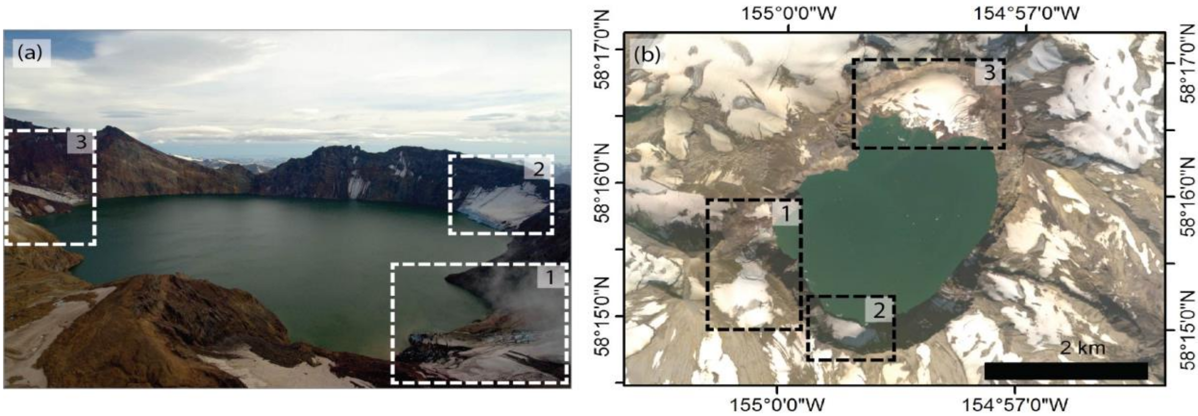

Fig. 6. Lake-filled caldera at Mount Katmai, formed due to summit collapse during the 1912 eruption of Novarupta (~ 9 km to the west). This figure shows an ice tongue extending from outside the SW margin of the caldera (dashed box 1), and two ‘new’ intra-caldera glaciers, one at the crater’s southern margin (dashed box 2), and one at its northern margin (dashed box 3). (a) Photograph taken by C. Read (AVO/USGS) on September 8, 2009, and obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=19191). (b) RapidEye™ image (September 1, 2016).

2.3.3. Complete glacier destruction

Cases of complete glacier destruction due to subglacial eruptions are rare (Supplementary Table 1). In fact, the only conclusive, documented examples relate to the 1980 eruption of Mount St Helens, which destroyed Loowit and Leschi Glaciers (Fig. 5). Though rare, these instances are interesting because they potentially allow the initiation, style, and timing of glacier (re)growth to be directly observed (Section 3.1.3.).

2.3.4. Glacier advance/acceleration

As with enhanced subglacial heating (Section 2.1.), subglacial volcanic eruptions can result in meltwater accumulation at the ice-bed interface, with the potential to promote subglacial sliding, glacier advance and/or acceleration. For example, following the 1953 eruption of Klyuchevskoy Volcano, Sopochny Glacier advanced 1–2 km; following the 1966–68 eruption, Vlodavtsa Glacier advanced 2.2 km; and following the 1977–80 eruption, Shmidta Glacier advanced until 1987, when part of the glacier tongue was destroyed by a second eruption, before advancing again following the 2005–10 eruption (Muraviev and Muraviev, 2016). Similarly, a fissure eruption at Tolbachik in 1975–76 caused the advance of Cheremoshny Glacier (Muraviev et al., 2011), and several effusive events and low-intensity explosive activity at Mt Belinda in 2001–07 (from a pyroclastic cone within an ice-filled caldera) apparently caused an adjacent valley glacier to advance (‘surge’) a few hundred metres into the sea (Smellie and Edwards, 2016).

2.3.5. Overall glaciological impacts of subglacial eruptions

The overall glaciological impact of subglacial eruptions is typically destructive, often involving considerable ice loss. The most common outcome is the formation of ice craters, melt pits and fractures (crevasses/ice-fissures). Despite their prevalence, smaller surface fractures and/or melt pits likely have little notable or long-term impact on glacier behaviour. By contrast, large ice surface cauldrons likely persist for a considerable time and re-open during repeated periods of activity. Larger surface cauldrons and fractures might also have an indirect impact on glacier behaviour, since they are a potential route for meltwater drainage to the bed, and can cause local reversals in ice flow direction (Fig. 2a, point 3). Examples of glacier beheading due to subglacial eruptions are comparatively common, but since beheading tends to coincide with supraglacial tephra/debris deposition, glaciers that might otherwise retreat (due to the loss of their accumulation areas) tend to stagnate (Section 2.6.2.). Complete glacier destruction is rare, as is the destruction of parts of glacier ablation zones (tongues).

2.4. Lava flows

Lava is produced during both effusive and explosive volcanic eruptions (Fig. 2d, points 19 and 23), and can flow supraglacially and/or subglacially (Fig. 2d, points 21 and 24). Supraglacial lava flows (and their glaciological consequences) are often conspicuous (i.e., they stand-out against the ice/snow over which they flow), whereas subglacial flows are extremely difficult to observe. The latter are often inferred from either lava flows disappearing into (or emerging from) glaciers, or from their impact at the ice surface (e.g., where they form supraglacial channels or depressions) (Fig. 2d, points 21 & 22). The primary glaciological impact of lava flows is to cause ice melt, and documented examples of both supraglacial and subglacial flows with glaciological consequences are discussed below.

2.4.1. Supraglacial lava flows

Documented supraglacial lava flows are comparatively common and often result from lava fountaining (e.g., at Beerenberg 1970–72; Volcán Villarrica 1971; Mount Westdahl 1991–92; Volcán Llaima 2008; Mount Pavlof 2013) or emanate from summit lava lakes (e.g., Volcán Llaima 2008) (Supplementary Table 1). They can extend for hundreds or thousands of meters, and produce notable supraglacial melt (e.g., Beerenberg 1970–72; Volcán Llaima 1979, 2008; Klyuchevskoy 1984–85, 1985–86, 1986–90; Volcán Hudson 1991; Mount Westdahl 1991–92; Mount Shishaldin 1999; Mt Belinda 2001–07; Mount Pavlof 2013). However, where lava effusion rates are low and/or surface debris is thick/extensive this melt is not always rapid (Section 3.2.1.). One common consequence of supraglacial lava flows is the formation of ice surface channels (sometimes tens of metres deep) (e.g., Volcán Villarrica 1971, 1984–85; Klyuchevskoy 1985–86; Mount Belinda 2001–07; Mawson Peak 2006–08; Volcán Llaima 2008) (Fig. 2d, point 21). These channels form as a direct consequence of melting, but can also develop through resulting hydrological erosion (e.g., Mount Shishaldin 1999). Where supraglacial lava flows begin to pond (Fig. 2d, point 22), crevasse-bounded depressions (melt pits) can form (e.g., Mount Veniaminof 1983–84, 1993–95, 2013; Mawson Peak 2006–08), into which supraglacial channels may extend. One example comes from Mawson Peak, where, in 2007, supraglacial lava flows (Fig 7a) and associated ponding appear to have melted supraglacial channels and a crevasse-bounded depression (Fig. 7b) (Patrick and Smellie, 2013).

Direct observations of the glaciological impacts of supraglacial lava flows are hindered by the snow often covering the higher altitude sectors of glaciers, i.e., where lava typically flows. Interactions between lava and snow are thus better observed and understood than interactions between lava and glaciers (e.g., Edwards et al., 2012, 2013, 2014, 2015). Much of this research indicates that lava’s impact on snow varies according to differences in the velocity and style of lava flows, and the thickness and distribution of supra-snowpack debris (including tephra) (Edwards et al., 2014, 2015). In many cases, lava can flow across snow without causing substantial melt (Edwards et al., 2014, 2015), and snow typically protects underlying glacial ice (e.g., Hekla 1947; Eyjafjallajökull/Fimmvörðuháls 2010; Tolbachik 2012–13) (Edwards at al., 2012, 2014). However, since snowmelt is partly inhibited by trapped air (which limits heat transfer), the rate of melting can notably increase when lava comes into direct contact with ice (e.g., Villarrica 1984–85) due to the reduced pore space relative to snow (Naranjo et al., 1993; Edwards et al., 2013, 2015).

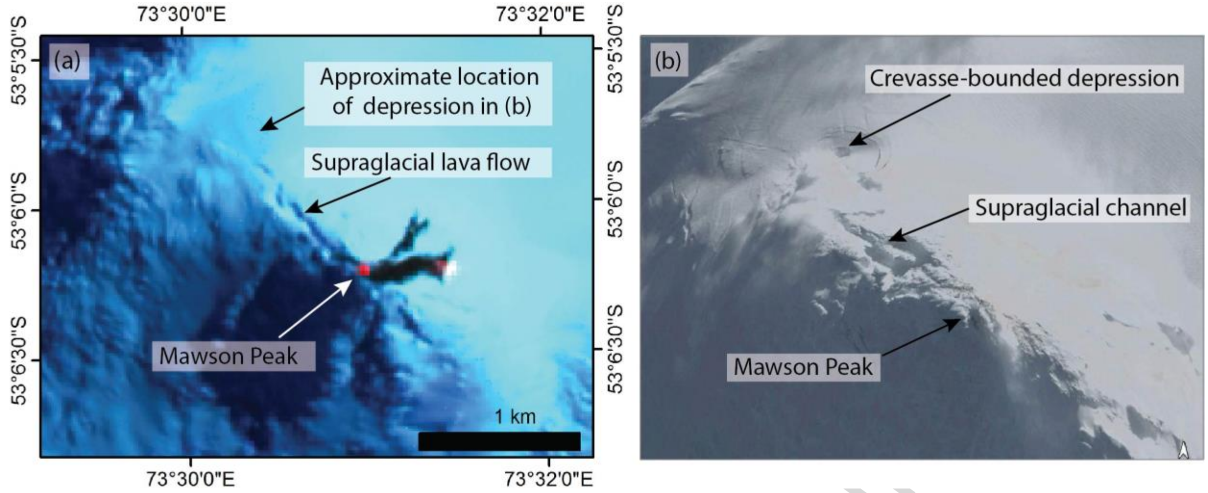

Fig. 7. (a) Supraglacial lava flows emanating from the summit of Mawson Peak on 17 February, 2007 (Landsat 7 ETM+ image). (b) Image from 13 May 2007 (DigitalGlobe™ image viewed obliquely in GoogleEarth™), showing a supraglacial channel and crevasse-bounded depression (melt pit), presumed to have been formed by the earlier supraglacial lava flow (shown in 'a') and subsequent ponding (not shown in 'a').

2.4.2. Subglacial lava flows

Subglacial lava flows are very difficult to observe, and, as a result, documented instances of their glaciological impact are far rarer than for supraglacial flows. Despite this, their impacts on glaciers have been observed in Chile and Iceland (Supplementary Table 1). As with supraglacial examples, subglacial lava flows can be kilometres long (e.g., Volcán Llaima in 1994; Eyjafjallajökull 2010) and usually result in ice melt (e.g., beneath Huemules Glacier during the 1971 eruption of Volcán Hudson) (Fig. 2d, point 24). Melting typically occurs above an advancing lava front (e.g., beneath Gígjökull during the 2010 eruption of Eyjafjallajökull). In some cases, subglacial melt can be violent, as ice and water rapidly vaporise when they interact with lava. One example comes from the western summit glacier at Volcán Llaima in 1994, when lava flow resulted in violent subglacial melt through the overlying glacier, and formed a subaerial ice channel up to ~ 150 m wide and ~ 2 km long (Moreno and Fuentealba, 1994). In some cases, subglacial lava flows have also resulted in doming, fracturing and subsidence of overlying ice (e.g., Volcán Calbuco in 1961; Volcán Llaima 1994) (Fig. 2d, point 26), though the exact mechanisms involved remain unclear (Klohn, 1963).

2.4.3. Overall glaciological impacts of lava flows

The overall glaciological impact of lava flows is to cause ice melt, leading to the formation of supraglacial and/or subglacial channels and melt pits (Fig 2d, points 21, 22, 24 & 26). However, this melting is often localised, and likely has limited impact on overall glacier behaviour. In addition, the extent of melt is partly determined by lava effusion rates (Section 3.2.1.1.), and melt does not always occur, particularly if the glacier is protected by considerable surface snow or debris (Section 2.8.2.).

2.5. Supraglacial pyroclastic density currents

Pyroclastic density currents are hot, gravity-driven mixtures of volcanic debris and gas that emanate from volcanoes (Druitt, 1998). They are direct products of eruptions, or occur following dome growth and subsequent collapse (Section 2., Fig. 2c, point 18). Dilute pyroclastic density currents are often referred to as surges, and more concentrated examples as flows (Burgisser and Bergantz, 2002). The primary glaciological impact of pyroclastic density currents is to cause ice loss through melting and abrasion/erosion (Julio-Miranda et al., 2005; Waythomas et al., 2013), and documented examples are discussed below.

2.5.1. Melt, erosion/abrasion

Pyroclastic density currents have resulted in observed glacier mass loss in the USA, Chile, Columbia, Ecuador, and Mexico (Supplementary Table 1). They often melt and entrain snow and ice (with ice blocks up to metres in diameter), and can transition from hot-dry surges to cold-wet flows (i.e., forming lahars—Section 2.7.) as they progress down-glacier (e.g., Mount Redoubt 1989–90). Because of melt and entrainment, supraglacial pyroclastic density currents often cut ice channels/gullies (Fig. 2c, point 18), up to tens of metres deep, sometimes with associated snow and ice levees (e.g., Cotopaxi 1877; Nevado del Ruiz 1985; Popocatépetl 1994–01; Mount Redoubt 2009). An example comes from Nevado del Ruiz during the 1985 eruption, when pyroclastic density currents cut surface channels 2–4 m deep and up to 100 m wide into Nereidas, Azufrado and Lagunillas Glaciers (Pierson et al., 1990).

Pyroclastic density currents are particularly destructive if funneled through topographically confined sections of ice, or cross steep and fractured icefalls (e.g., Drift Glacier during the 1966–68, 1989–90 and 2009 eruptions of Mount Redoubt; Kidazgeni Glacier during the 1992 eruption of Mount Spurr; and Nereidas, Azufrado and Lagunillas Glaciers during the 1985 eruption of Nevado del Ruiz), when crevasses can be mechanically abraded and seracs planed smooth (e.g., Nevado del Ruiz 1985). In extreme cases, sections of ice can be scoured to bedrock, effectively beheading glaciers by separating their accumulation and ablation zones. A notable example comes from the eruptions of Mount Redoubt in 1966–68 and 1989–90, when a ~100 m thick gorge section of Drift Glacier was scoured to bedrock by supraglacial pyroclastic density currents (Trabant et al., 1994) (Fig. 8). This separated the glacier's accumulation and ablation zones, and reduced ice flux on the lower, piedmont section of Drift Glacier by more than 50% (Sturm et al., 1986).

Where pyroclastic density currents emerge from steeper, confined sections of glaciers onto shallower piedmont lobes (or other parts of the ablation area), they can still incise channels into the glacier surface (sometimes exploiting pre-existing longitudinal crevasses). For example, at Mount Redoubt in 1989–90 and 2009, channels in Drift Glacier's piedmont lobe were 10–100 m deep and wide, and formed a deeply incised ice-canyon system, which extended to the glacier bed (Trabant and Meyer, 1992) (Fig. 9). However, the extent of scouring tends to diminish down-slope and is minimal at glacier termini (e.g., Nevado del Ruiz in 1985). In fact, the lower sections of glaciers are often more likely to be covered by debris derived from pyroclastic density currents (e.g., Mount Spurr 1992) (Section 2.8.), protecting the surface from further incision (Section 2.8.2).

In general, dilute, fast-moving pyroclastic surges have limited glaciological impact since they are unable to produce much melting (i.e., they do not have enough thermal mass), but higher-density pyroclastic flows efficiently melt and entrain glacial ice. This was exemplified during the 1985 eruption of Nevado del Ruiz, when both dilute pyroclastic surges and concentrated pyroclastic flows were produced. The former caused no significant melting, while the latter eroded and melted into the underlying glaciers (Pierson et al., 1990). During the 1985 eruption, pyroclastic density currents also promoted ice and rock avalanches that led to major ice losses from glaciers in the Azufrado, Lagunillas and Farallon-Guali basins—destroying the 10–15 m thick crevassed terminus of Lagunillas Glacier and the hanging glaciers on the headwall of the Azufrado valley (Pierson et al., 1990).

2.5.2. Overall glaciological impacts of pyroclastic density currents

Pyroclastic density currents are some of the most glaciologically destructive volcanic events, as they rapidly melt and entrain ice, particularly on steep and crevassed sections of glaciers (e.g., at icefalls). In extreme cases, they can scour glacier ice to bedrock (effectively causing glacier beheading), and are known to produce voluminous lahars (McNutt et al., 1991) (Section 2.7). However, the glaciological impact of pyroclastic density currents is partly determined by the concentration of clasts present, with concentrated flows more destructive than dilute surges, and is limited where pre-existing surface debris is extensive/thick (Section 2.8.2.).

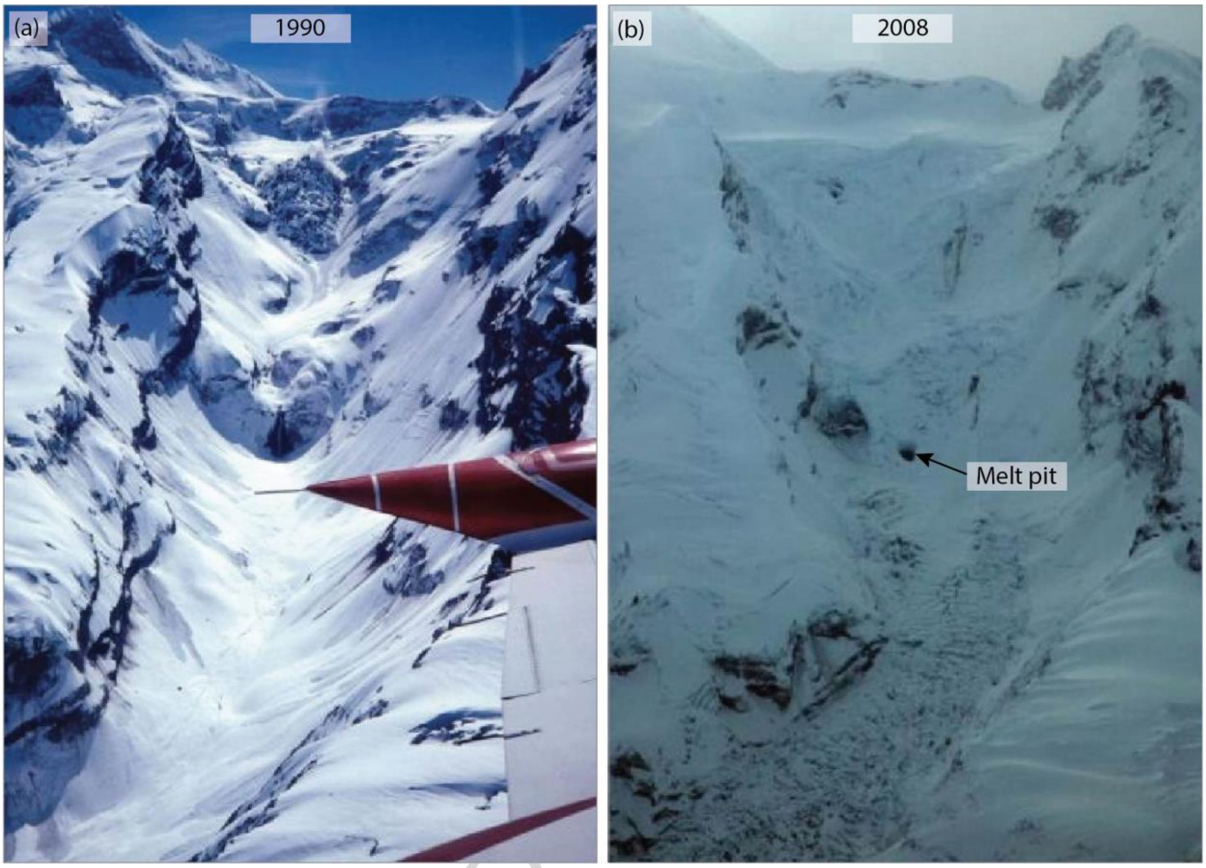

Fig. 8. The upper ‘gorge’ section of Drift Glacier (a) following the 1989–90 eruption of Mount Redoubt, and (b) immediately prior to the 2009 eruption. In (a), this section of the glacier has recently been scoured to bedrock (by pyroclastic density currents), effectively beheading the glacier. In (b) the glacier has recovered from beheading during the the 1989–90 eruption, but due to a period of unrest prior to the 2009 eruption, a prominent melt hole/pit is evident above a water fall shown in the 1990 image. Photographs taken by G. McGimsey (AVO/USGS), and obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=16578).

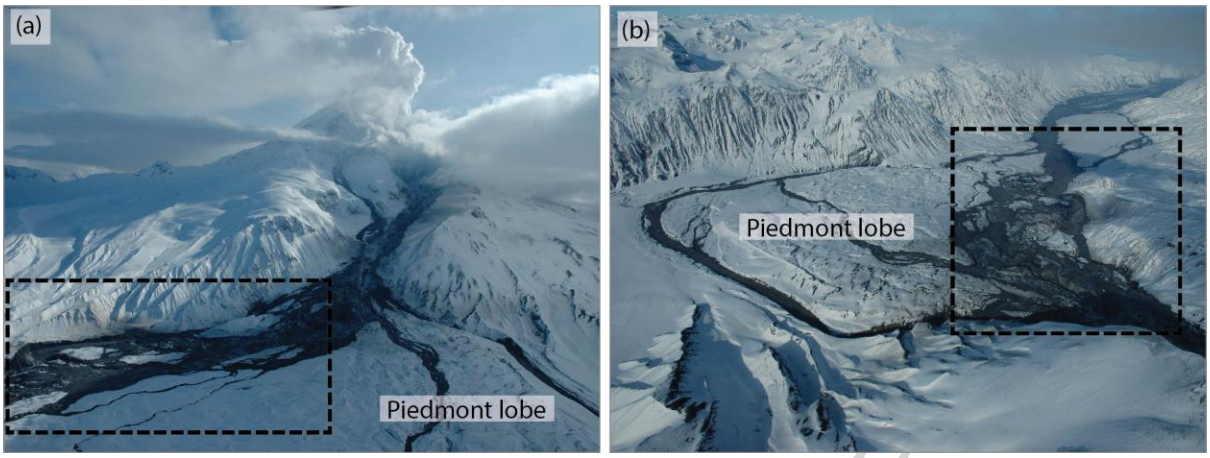

Fig. 9. Channels (in dashed boxes) and debris on the piedmont lobe section of Drift Glacier, formed by pyroclastic density currents and associated lahars following the 2009 eruption of Mount Redoubt. (a) View towards the south. (b) View towards the north. Photographs taken by G. McGimsey (AVO/USGS) on March 26, 2009. Images obtained from the AVO/USGS database (http://www.avo.alaska.edu/images/image.php?id=47241; http://www.avo.alaska.edu/images/image.php?id=47251). Descriptions based on McGimsey et al. (2014).

2.6. Supraglacial deposition of tephra

The deposition of tephra (ash, rock fragments and particles ejected by volcanic eruptions) can occur on glaciers occupying volcanoes or on glaciers down-wind of source eruptions. The glaciological impact depends on many factors including tephra temperature, thickness, spatial coverage, pre-existing surface debris, and weather conditions during and after deposition (Brook et al., 2011; Nield et al., 2013). For example, the impact of tephra cover is likely to be less important for glaciers that are debris- or snow-covered prior to tephra deposition (Rivera et al., 2012). In many cases, directly following deposition, supraglacial tephra causes increased melt due to its elevated temperature (though, except in a narrow zone close to the vent, many fall deposits are probably cold when they land), relative to the ice on which it lands. Once cooled, the impact might still be to promote melt, particularly if the tephra deposit is thin and/or discontinuous (Fig. 2d, point 28), as the albedo of the ice surface is reduced (Richardson and Brook, 2010). However, when tephra is more continuous, and particularly once it exceeds a threshold thickness, surface melt is likely to reduce (Fig. 2d, point 27). This effect is due, in large part, to the low thermal conductivity of the tephra, and (to a lesser degree) to its ability to shield ice from solar radiation (Brook et al., 2011; Rivera et al., 2012; Wilson et al., 2013). Thus, the glaciological impact of tephra cover is largely governed by its thickness, and whether or not the threshold thickness, which varies from glacier to glacier, is exceeded (Kirkbride and Dugmore, 2003). Documented examples of supraglacial tephra causing increased or decreased melt, and associated changes in glacier dimensions, are discussed below.

2.6.1. Increased melt

Examples of supraglacial tephra deposition resulting in increased melt and/or glacier recession come from Chile, the Sub-Antarctic Islands, Ecuador, Iceland, Mexico, and New Zealand (Supplementary Table 1). For example, on Deception Island, a short-lived eruption in 1969 deposited supraglacial tephra which lowered the ice surface albedo, and resulted in particularly negative mass balance for three subsequent years (up to 1973) (Orheim and Govorukha, 1982). In some cases, increased melt is caused by tephra deposition on glaciers that are kilometres away from source eruptions. For example, during the 2008 eruption of Volcán Chaitén, tephra deposition increased melt at glaciers occupying Volcán Michinmahuida, ~ 15 km to the east (Alfano et al., 2011; Rivera et al., 2012). Similarly, following the 1999–2001 eruption of Tungurahua Volcano, tephra deposition on glaciers occupying Chimborazo Volcano, ~ 40 km to the west, led to increased melt and small-scale glacier retreat (Morueta-Holme et al., 2015; La Frenierre and Mark, 2017).

Where glaciers are present on tephra-producing volcanoes, their upper reaches (i.e., in close proximity to vents) can become covered by thick tephra, promoting ice preservation, while lower sections are covered by comparatively thin tephra, promoting ice loss and glacier disintegration (Wilson et al., 2013). For example, during the 1994–2001 eruptive period at Popocatépetl, the upper part of the glacier occupying its summit was covered with thick tephra, whilst its lower reaches were covered with a thinner, discontinuous tephra layer. As a result, ice melt was suppressed in the upper part of the glacier, where a flat area began to form, and ice began to thicken. This region was separated from the lower section of the glacier, where rapid ice loss resulted in ‘stair-like’ morphology (Julio-Miranda et al., 2008). Because of this differential ablation, the glacier surface steepened, and ice was transmitted towards the terminus as a kinematic wave of ice thickening (Julio-Miranda et al., 2008). This did not cause glacier advance, but led to increased melt as ice was transferred into the ablation zone. As a result, the glacier front retreated dramatically in 2000, and much of the remainder began to fragment (by a combination of differential ablation and tephra remobilisation) (Julio-Miranda et al., 2008). By 2001, the glacier had fragmented into a set of ice blocks. Though these ice blocks were insulated on their upper surfaces, their tephra-free flanks were exposed to ablation. Ultimately, tephra deposition notably enhanced climate-related glacier recession, with ~ 53% of the glacier's surface area lost between 1996 and 2001 (Julio-Miranda et al., 2008). In 2004, the glacier disappeared completely (or at least the remaining ice was no longer flowing) (Julio-Miranda et al., 2008).

Despite such instances, making clear links between tephra deposition and glacier retreat can sometimes be difficult. For example, at Volcán Lautaro, eruptions during the 20th century deposited tephra on some adjacent glaciers (many of these were outlets of the South Patagonian Ice Field). O'Higgins Glacier experienced rapid retreat (~ 14.6 km between 1945 and 1986). However, whether this was the result of increased calving or supraglacial tephra deposition (and perhaps geothermal heating) is unclear (Lopez et al., 2010).

In addition to tephra, other volcanic materials (including 'bombs') can be ejected from volcanoes, land supraglacially, and cause ice-melt. For example, during the 1994–2001 eruptive period at Popocatépetl, incandescent material landed supraglacially, and formed melt holes and impact craters (via physical impact and subsequent melting) (Julio-Miranda et al., 2008). The parts of glaciers where ejected material typically lands (i.e., in a glacier's upper reaches, in proximity to vents) are often snow covered, hence melt pits have been observed in supraglacial snow (e.g., Nevado del Ruiz, 1985; Belinda 2001–07) (Pierson et al., 1990; Smellie and Edwards, 2016), but impacts on underlying ice often remain unclear. In addition, melt due to the supraglacial deposition of ejecta is often very localised, and is likely to have little (if any) impact on overall glacier behaviour.

2.6.2. Decreased melt

There are numerous documented instances where supraglacial tephra deposition has resulted in decreased melt and/or glacier stagnation (preservation). This includes examples from the USA, Chile, Iceland, Jan Mayen Island, and Kamchatka (Supplementary Table 1). For example, tephra deposited on the surface of Svínafellsjökull during the 2011 eruption of Grímsvötn is estimated to have reduced ablation rates by up to 59% (Nield et al., 2013). Kozelsky Glacier stagnated over much of the 20th century in response to supraglacial tephra deposited during the 1945 eruption of Avachinsky (Vinogradov and Muraviev, 1982; Muraviev et al., 2011). The same happened to Knife Creek Glacier and two of the Mount Griggs Glaciers following the 1912 eruption of Novarupta (Hildreth and Fierstein, 2012).

A small number of glaciers are thought to have advanced in response to tephra-related decreases in melt. For example, Gígjökull advanced following the deposition of Hekla tephra in 1947 (Kirkbride and Dugmore, 2003), and Knife Creek Glaciers 1, 2, 4, 5 and one of the Mount Griggs Glaciers advanced following the 1912 eruption of Novarupta (Hildreth and Fierstein, 2012). However, given overall climatically driven glacier mass loss during the late 20th century, in many cases ice insulation beneath supraglacial tephra has simply slowed the rate of recession. For example, following the 1970–72 eruption of Beerenberg, supraglacial tephra reduced surface ablation and thus slowed the rate of retreat at Sørbreen (this continued up to 1978) (Anda et al., 1985). Similarly, due to a mantle of supraglacial tephra, the late 20th century retreat of Pichillancahue-Turbio Glacier at Volcán Villarrica has been slower than for other glaciers in the region (Masiokas et al., 2009).

2.6.3. Overall glaciological impacts of supraglacial tephra deposition

The overall glaciological impacts of supraglacial tephra deposition are complex, with documented examples of both increased and decreased melt. The former are often instances where tephra cover is thin and/or discontinuous (Fig 2d. point 28), while the latter reflect thick and/or continuous coverage (Fig 2d, point 27). There are also documented examples of glacier advance or retreat in response to supraglacial tephra deposition, but making conclusive causal links between changes in glacier dimensions and periods of deposition is difficult. In many cases, tephra distributes heterogeneously across a glacier's surface and therefore enhances melt on some portions, and restricts it on others.

This differential melting can result in an undulating ice surface, and in some cases has a notable impact on glacier dynamics and mass balance.

2.7. Floods and lahars

Floods (including jökulhlaups—glacier outburst floods) and lahars (mixed meltwater and debris) are a common consequence of glacio-volcanic activity, extensively studied in the wider geo-hazards framework (Major and Newhall, 1989). Floods are distinguishable from lahars by their debris content (i.e., floods are dilute, while lahars are concentrated with debris). However, they are grouped together here, since initially dilute floods can transition into lahars as they accumulate debris (e.g., Mount Shishaldin 1999) (Fig. 2a, point 4), and their triggers and glaciological impacts are often similar. Floods and lahars are caused by sudden ice melt linked to many of the processes described in previous sections of this paper, including: melt caused by enhanced subglacial heat flow (e.g., Mount Kazbek various 20th and 21st century); melt during subglacial eruptions (e.g., Hekla 1947); hot subglacial water migrated to the glacier surface through ice fissures (e.g., Deception Island 1969; Mount Westdahl 1991–92; Nevado del Huila 2007–12); melt caused by supra- and sub-glacial lava flows (e.g., Beerenberg 1970–72; Volcán Llaima 1979, 1994, 2008; Klyuchevskoy 1984–85, 1985–86, 1986–90; Volcán Hudson 1991; Mount Westdahl 1991–92; Mount Pavlof 2013); melt caused by pyroclastic density currents (e.g., Cotopaxi 1877; Volcán Calbuco 1961; Mount St Helens 1980; Mount Spurr 1992; Popocatépetl 1994–01; Mount Redoubt 2009; Mount Pavlof 2013); and melt generated by the supraglacial deposition of tephra (e.g., Volcán Hudson 1991). The glaciological consequences of floods and lahars (i.e., the focus of interest here) include supraglacial and subglacial melt/erosion (including ice break-off) (Fig. 2a, point 13) and glacier advance/acceleration. Documented examples of these different outcomes are discussed below.

2.7.1. Supraglacial melt/erosion

Examples of floods and/or lahars causing supraglacial melt/erosion are common and are reported in Alaska, Chile, Columbia, the Sub-Antarctic Islands, Ecuador, Iceland, Kamchatka, and New Zealand (Supplementary Table 1). These (often warm) floods and lahars can cut supraglacial channels/canyons (with vertical ice walls) as they melt and entrain ice (e.g., Mount Westdahl 1978 and 1991–92; Mount Shishaldin 1999; Fourpeaked Mountain 2006) (Fig. 10; Fig. 2a, point 5). Supraglacial channels can be kilometres long (e.g., channels cut into the surface of Vatnajökull during the 1996 eruption of Gjálp; and on the surface of the glacier occupying the NE flank of Mount Pavlof following its 2013 eruption) and terminate in newly developed, or pre-existing, moulins or cauldron-shaped collapse features (e.g., on Drift Glacier during the 1966–68 eruption of Mount Redoubt; and on Vatnajökull during the 1996 eruption of Gjálp) (Fig. 2a, point 8).

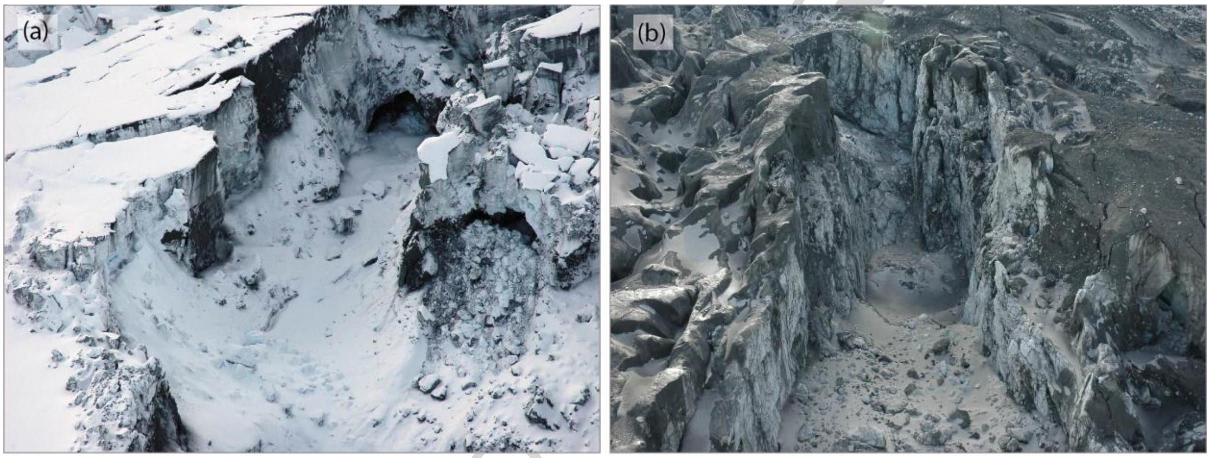

Fig. 10. Supraglacial channels cut by an outflow event (lahar/flood) during the 2006 eruption of Fourpeaked Mountain (channels are cut into a glacier on the mountain's NW flank). (a) The point of origin of the outflow event. (b) A deeply incised canyon (~ 100 m deep and wide). Both photographs taken by K.L. Wallace (AVO/USGS) on September 25, 2006. Images obtained from the AVO/USGS database (http://www.avo.alaska.edu/image.php?id=11837; http://www.avo.alaska.edu/image.php?id=11848). Descriptions based on Neal et al (2009).

2.7.2. Subglacial melt/erosion

Documented instances of floods or lahars causing subglacial melt/erosion are comparatively rare, likely reflecting difficulties with observing subglacial environments (Section 3.2.2.). Despite this, there are two notable examples from Iceland (Supplementary Table 1). The first occurred following the 1996 eruption of Gjálp, when warm (15–20°C) meltwater travelled 15 km along a narrow channel beneath Vatnajökull and into Grímsvötn subglacial lake, exiting it later as a jökulhlaup (Gudmundsson et al., 1997). Subglacial melting occurred along the meltwater flow path into Grímsvötn, inside the lake, and on the jökulhlaup path out of the lake, and resulted in supraglacial subsidence and the formation of a shallow linear depression in the ice surface (Gudmundsson et al., 1997). The second example occurred during the first days following the main explosive eruption at Eyjafjallajökull, in 2010, when meltwater drained beneath Gígjökull in several jökulhlaups (Magnússon et al., 2012). Ice-melt and mechanical erosion occurred along the subglacial flood path due to the thermal and frictional energy of floodwaters, forming a subglacial channel. However, because of the high water pressures, floodwaters destroyed the roof of the channel, and emerged supraglacially as a slurry flow (Magnússon et al., 2012). Thus, drainage was subglacial for the first 1–1.5 km, but then water emerged, and drained supraglacially down both sides of the glacier.

Documented examples of subglacial floods breaking blocks of ice from glacier termini (Fig. 2a, point 13) are rare and only reported in Iceland. For example, at Katla, in 1918, an eruption beneath a ~ 400 m thick section of the Mýrdalsjökull ice cap caused a major jökulhlaup, with water flowing both supraglacially and subglacially. The force of the subglacial meltwater is considered to have torn icebergs (50–60 m diameter) from the glacier terminus, where it also blasted a 1,460–1,830 m long, 366–550 m wide, and 145 m deep gorge (Russell et al., 2010). Similarly, at Grímsvötn in 1934, subglacial melt led to a large jökulhlaup that removed ice blocks from the terminus of Skeidarárjökull, resulting in 40-m-high fracture faces (Nielsen, 1937).

2.7.3. Glacier advance/acceleration

Documented examples of floods and/or lahars causing glacier advance/acceleration are uncommon, and the evidence connecting these events is rarely clear. Despite this, there are examples from Alaska, the Sub-Antarctic Islands, and Iceland (Supplementary Table 1). For example, at Deception Island, in 1969, a large jökulhlaup flowed across the summit ice cap. Downslope from the ice fissures from which this flood emanated, the ice experienced a short-lived surge-like advance (Smellie and Edwards, 2016). At Katla during the 1918 jökulhlaup, while icebergs were torn from the glacier front (Section 2.7.2.), the whole glacier terminus floated (Fig. 2a, point 12) and may have moved forward

(Smellie and Edwards, 2016). On Montagu Island in 2001–07, subglacial melt, triggered by the eruption of Mt Belinda caused an adjacent valley glacier to advance a few hundred metres into the sea (Smellie and Edwards, 2016). Finally, during the 2004 Grímsvötn eruption, a jökulhlaup (the onset of which preceded the eruption by four days) apparently caused the short-term (monthly) flow velocity of Skeidarárjökull (an outlet of Vatnajökull) to increase by up to 0.4 m d^{-1}, compared to annual values (Martinis et al., 2007; Sigurðsson et al., 2014) (Fig. 11). This acceleration occurred over the entire width of the glacier, and was potentially caused by increased subglacial sliding due to widespread basal lubrication (Martinis et al., 2007; Sigurðsson et al., 2014).

2.7.4. Overall glaciological impacts of floods and lahars

The overall glaciological impact of floods and lahars is typically destructive, casing ice melt, erosion, entrainment, and, in a small number of cases, ice-block break-off from glacier termini. There are examples where floods are presumed to have caused glacier advance/acceleration. However, this only relates to a small number of cases, and the evidence is rarely clear. Many of the better-documented examples of interactions between floods/lahars and glaciers come from Iceland, where glaciers are comparatively accessible, close to settlements, and easily observed.

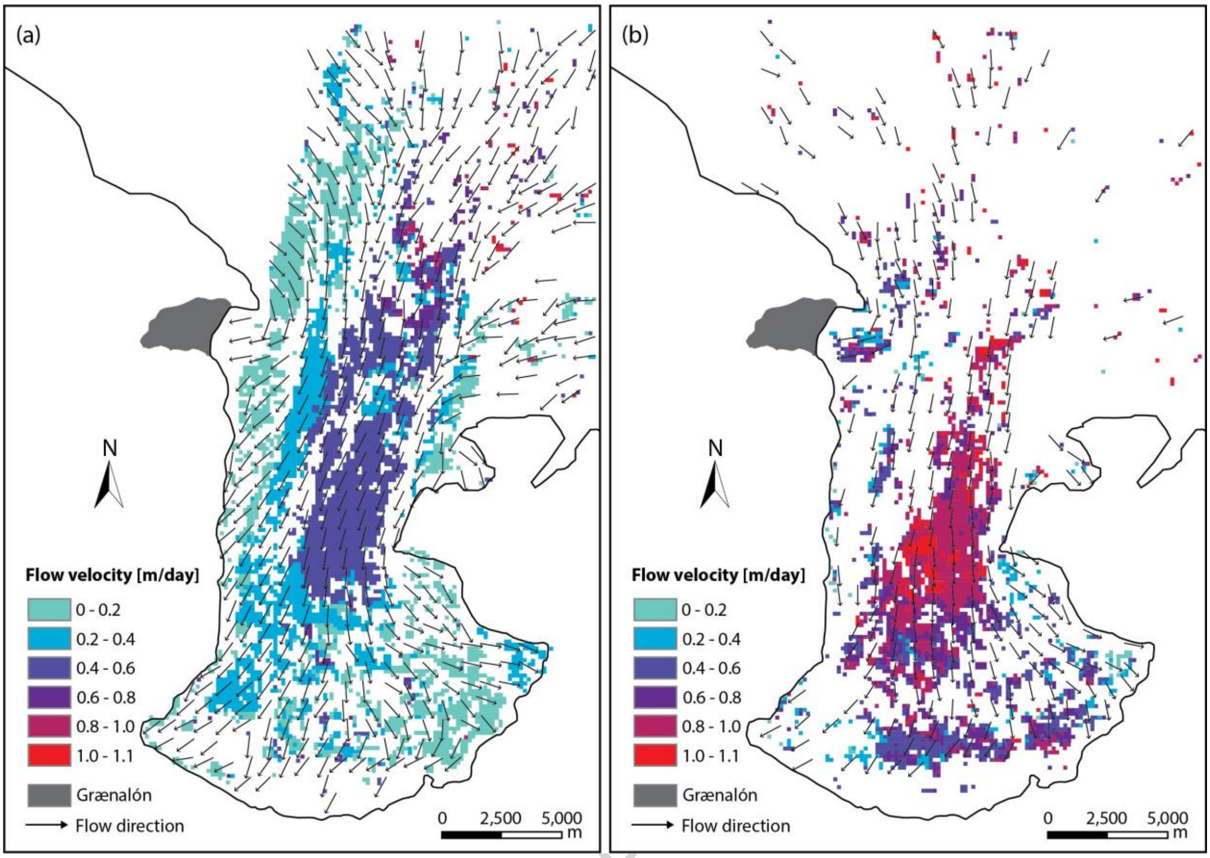

Fig. 11. Surface velocity fields at Skeidarárjökull (an outlet of Vatnajökull), derived from ASTER satellite images (Martinis et al., 2007). (a) Velocity between September 27, 2004 and July 28, 2005 (i.e., approximately annual velocity). (b) Velocities during a period (i.e., from September 27, 2004 to November 30, 2004) which coincides with the 2004 eruption of Grímsvötn (November 1–6). The accelerated flow in (b) is thought to result from increased glacier sliding, related to widespread basal lubrication caused by a subglacial jökulhlaup. Figure modified from Sigurðsson et al. (2014).

2.8. Supraglacial deposition of other glacio-volcanic products

Many of the glacio-volcanic processes outlined in this paper not only have direct glaciological impacts, but also result in debris which, when deposited supraglacially, can impact glacier response to other forcing mechanisms (e.g., climate). In almost all cases, the glaciological consequence is that the debris acts to reduce ice ablation, and/or protects ice from further thermal and/or mechanical erosion. Documented examples of these scenarios are discussed below.

2.8.1. Reduced ablation

Debris derived from pyroclastic density currents (e.g., Novarupta 1912; Bezymianny 1955–57), avalanches/landslides (Klyuchevskoy 1944–45; Mount Redoubt 1989–90), and floods/lahars (Mount Redoubt 1966–68, 1989–90, 2009) has acted to insulate glacier ice (Supplementary Table 1). The main result is typically a slowing in the rate of climatically driven glacier retreat/mass-loss. However, in some cases, the insulating impact of surface debris is thought to have caused glacier advance. The most notable example is Erman Glacier, which has advanced by ~ 4 km since the 1944–45 eruption of nearby Klyuchevskoy volcano (from which the glacier partly emanates). This glacial advance is ongoing (Fig. 12), despite regional atmospheric warming, and is thought to reflect the impact of landslide debris which was deposited on the glacier's accumulation area during the 1944–45 eruption, and subsequently spread to cover the ablation area where it likely acted to insulate the underlying ice (Muraviev and Muraviev, 2016; Dokukin et al., 2017).

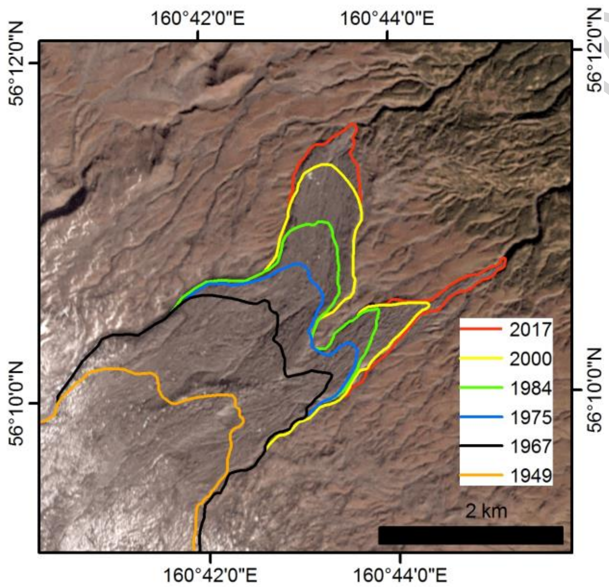

Fig. 12. Post-1949 advance of Erman Glacier (Kamchatka) following the supraglacial deposition of landslide debris during the 1944–45 eruption of Klyuchevskoy volcano (from which the glacier partly emanates). Image based on Muraviev and Muraviev (2016).

2.8.2. Glacier protection from erosion

Instances where supraglacial deposits derived from the glacio-volcanic processes outlined in this review have acted to protect ice from subsequent erosion/incision are best documented at Mount Redoubt (Supplementary Table 1). For example, floods and lahars during the 1989–90 and 2009 eruptions led to the formation of a supraglacial 'ice diamict', composed of gravel-sized clasts of glacier ice, rock, and pumice in a matrix of sand, ash, and ice (frozen pore water) (Waitt et al., 1994). On the piedmont section of Drift Glacier (Fig. 9), these deposits were 1–10 m thick, and protected the underlying ice from thermal and mechanical erosion by later supraglacial pyroclastic density currents and lahars (Gardner et al., 1994).

2.8.3. Overall glaciological impacts of the supraglacial deposition of other glacio-volcanic products

The primary glaciological consequence of supraglacial pyroclastic, avalanche, lahar and flood deposits is to insulate the underlying ice, and protect it from further thermal and/or mechanical erosion. This typically promotes glacier preservation/stagnation, and partly acts to counter the otherwise largely glaciologically destructive impacts of volcanic activity.

3. Present and future volcanic impacts on glaciers

Here, we use the information outlined in section 2 to address three general questions about volcanic impacts on glaciers. 1. What are the overall glaciological consequences of volcanic activity? 2. How many of Earth's glaciers are impacted by such activity? 3. What is the future importance of volcanic impacts on glaciers?

3.1. What are the overall glaciological consequences of volcanic activity?

The glaciological consequences of volcanic activity typically relate to local increases in meltwater and debris. However, the importance of different processes and interactions varies according to the timescale under consideration.

3.1.1. Short-term

Over the period of days-to-months, the glaciological impacts of volcanic activity are typically destructive, involving ice-melt, erosion and entrainment (e.g., Sections 2.3.2. and 2.3.3). For example, during the 1996 eruption at Gjálp, 3 km^3 of ice melted in just 13 days (when the eruption ended), with a further 1.2 km^3 melting over the following three months (Gudmundsson et al., 1997). One of the key reasons mass loss dominates over the short-term is that it can be caused by a number of processes that typically occur during the early stages of volcanic activity, including enhanced subglacial heat flow, subglacial volcanic eruptions, supra- and sub-glacial lava flows, pyroclastic density currents, floods/lahars, and the supraglacial deposition of hot tephra. In some cases, ice loss due to these processes results in supraglacial subsidence, deformation, and fracturing; in other cases, glaciers are partially or entirely destroyed.

3.1.2. Medium-term

Over the period of months-to-years, volcanic activity can act to either destroy or preserve glacial ice. For example, pyroclastic density currents, which are destructive over the short-term, may lead to ice preservation over the medium-term (and perhaps longer; Carey et al., 2010), as their deposits insulate and protect underlying ice (Section 2.8.). These medium-term impacts may also act to counter some of the short-term destruction. For example, following beheading, glaciers that might otherwise rapidly retreat (in response to partial or complete removal of their accumulation areas) often stagnate due to ice protection beneath supraglacial tephra/debris (Section 2.3.2.). Despite such instances, documented cases of volcanic activity causing glacier stagnation and/or advance/acceleration are certainly less common than instances of mass loss and/or glacier retreat. Clear evidence linking volcanic activity to periods of glacier advance or acceleration is particularly scarce. The medium-term impacts of volcanic activity also partly depend on factors such as the weather conditions following eruptions (which control how supraglacial material is re-distributed) (Nield et al., 2013), and glaciological characteristics such as the efficiency of subglacial drainage (Section 3.2.1.2.), and therefore vary from glacier to glacier.

3.1.3. Long-term

Glaciological impacts of volcanism can be observed years-to-decades after periods of activity (e.g., Carey et al., 2010). For example, the advance of Erman Glacier continues to this day, apparently in response to the supraglacial deposition of landslide debris during the 1944–45 eruption of Klyuchevskoy (Fig. 12). In some cases, glaciers that were destroyed or beheaded by volcanic activity fail to recover, and disappear entirely (e.g., Shoestring, Nelson, Forsyth, and Dryer Glaciers at Mount St Helens). In other cases, beheaded glaciers recover and new glaciers form in areas where they were previously destroyed. A notable example of recovery from beheading is Drift Glacier, which was beheaded by pyroclastic density currents during the 1966–68 and 1989–90 eruptions of Mount Redoubt (Section 2.5.1.), but in less than a decade had re-formed (e.g., Fig. 8). Following the 1966–68 beheading, the reconnection of the regenerated part of the glacier and the piedmont section below resulted in a kinematic wave of thickening (of > 70 m) and surface acceleration (by an order of magnitude) in the lower section of the glacier, whilst thinning (by ~ 70 m) occurred in the upper section. These processes were accompanied by surface crevassing, likely reflecting the glacier's return to its pre-eruption equilibrium condition (Sturm et al., 1986).

Notable examples of new glacier formation following destruction by volcanic activity come from Mount St Helens and Mount Katmai. At Mount St Helens, some glaciers were beheaded and some destroyed during the 1980 eruption (Fig. 5), but, by 1999, a ~ 1 km^2 and ~ 200 m thick glacier had reformed in the initially ice-free summit crater (Schilling et al., 2004; Walder et al., 2007). This glacier was later displaced and deformed by subglacial dome extrusion (Section 2.2.). At Mount Katmai, summit collapse and glacier beheading during the 1912 eruption of Novarupta (Section 2.3.2.) generated a glacier-free caldera. Snow and ice then began to accumulate on inward sloping intra-caldera benches (300–400 m above the caldera floor). Snow patches had accumulated by 1917, modest snowfields by 1923, while the earliest confirmed reports of active glacial ice came in 1951. According to Muller and Coulter (1957), intra-caldera glaciers had effectively formed within 20 years of the summit collapse. These 'new' glaciers formed at the crater's northern and southern margins, with an ice tongue at the crater's SW margin extending from outside the caldera (Fig. 6). By 1953/54, the South glacier terminated in cliffs 50–80 m above the caldera lake, and the northern glacier reached halfway down to the lake. By 1987, both terminated just above lake level, but upon reaching the heated lake, melted rapidly. Thus, the volcanic lake (which has increased in depth since its inception) has acted to deter glacier growth/advance (Hildreth and Fierstein, 2012).

Overall, despite initial destruction or damage from which some glaciers fail to recover, the long-term glaciological impacts of volcanic activity often appear to be constructive, involving glacier re-growth or the formation on new ice masses.

3.2. How many of Earth's glaciers are impacted by volcanic activity?

The dataset presented here (Supplementary Table 1) suggests that observed volcanic impacts on the behaviour of modern glaciers are comparatively rare, and are only documented for ~ 150 glaciers (~ 0.08% of the global glacier population), during ~ 90 separate volcanic events or periods of activity. However, in considering the importance of these numbers, it is worth focusing on two questions. 1. What determines whether volcanic activity has a glaciological impact? 2. What determines whether volcanic impacts on glaciers are observed?

3.2.1. What determines whether volcanic activity has a glaciological impact?

There are numerous documented examples where, despite proven volcano-glacier interactions, volcanic activity has failed to produce an observable glaciological impact. In fact, the nature of volcanic impacts on glaciers, and whether or not volcanic activity has an observable glaciological impact, seems to partly depend on event size and duration, and glacier properties (including glacier size, thermal regime, and the nature of subglacial drainage).

3.2.1.1. Event size and duration

Large and/or long-lasting volcanic events are likely to have a greater glaciological impact than smaller/shorter equivalents. In addition, event size and duration may play a role in governing the nature of volcanic impacts (when they occur). For example, during the 1984–85 eruption of Volcán Villarrica, lava flows on the northern and northeastern slopes of the volcanic edifice melted the ice surface into numerous supraglacial channels and generated small floods (Section 2.4.1.), but lava effusion rates (20 m^3 s^{-1}) were too low to generate large floods/lahars (Moreno, 1993; Delgado

Granados et al., 2015). Similar conditions were observed during the early stages of the 1971 eruption, but effusion rates and lava volumes increased as a bedrock fissure opened across the summit crater (with lava fountains up to 400 m high and effusion rates up to 500 $m^3 s^{-1}$). This resulted in sufficient melting of the summit glaciers to generate lahars in five different drainage basins (Marangunic, 1974; Moreno, 1993), thus indicating that the effusion rates of lava flows have a strong control on glaciers, though factors such as the velocity and style of lava flows also play a role (Section 2.4.1.). Similarly, the size and duration of a volcanic eruption might determine its glaciological impact by controlling the thickness and extent of supraglacial tephra deposits—hence determining whether threshold thicknesses are exceeded, whether melt is enhanced or reduced, and whether underlying ice is protected (Section 2.6.). A further consideration here is where and when such materials are deposited. For example, material deposited in a glacier's accumulation area and/or during winter may become quickly snow covered, limiting its impact on surface albedo.

3.2.1.2. Glacier properties

Glacier size (horizontal extent and thickness) has some control over the nature of volcanic impacts. For example, dome extrusion has only been observed through small/thin glaciers (Section 2.2.2.), whilst surface craters surrounded by concentric crevasses are more likely to form on thick glaciers (Section 2.3.1.). It is also likely that ice sheets are less susceptible to many of the volcanic impacts described in this review because of their substantial thickness (km thick). In particular, subaerial processes (such as supraglacial tephra deposition and pyroclastic density currents) likely have limited overall glaciological impact. By contrast, widespread subglacial melt might have profound implications for ice-sheet stability (Blankenship et al. 1993; Vogel et al. 2006; Corr and Vaughan 2008; de Vries et al., 2017), though our understanding of volcanic impacts on ice sheet behaviour is limited by a dearth of observational information.

Another important property that might affect a glacier's response to volcanic activity is the basal thermal regime (i.e., whether cold-based, warm-based, or polythermal). For example, it is assumed that for temperate (warm-based) or polythermal glaciers, increases in subglacial meltwater might not necessarily result in advance/acceleration (as described in Sections 2.1.3. and 2.3.4.) since the bed is already wet. By contrast, cold-based glaciers are often frozen to their beds with minimal subglacial meltwater drainage, and increased subglacial melt is therefore likely to have a greater impact on glacier behaviour (i.e., resulting in acceleration and/or advance) (Rivera et al., 2012). In the case of cold-based ice sheets, the impact of subglacial melt is difficult to predict, as any meltwater generated might be confined by surrounding frozen-based ice, and therefore spatially limited. Similar difficulties exist in predicting the impact on polythermal glaciers, with their patchwork of cold- and wet-based ice (see Smellie et al., 2014). However, and regardless of the dominant thermal regime, the impact of increased subglacial melt varies from one ice mass to another, and may partly depend on the nature of subglacial meltwater routing. For example, during the 2004 eruption of Grímsvötn, a jökulhlaup apparently caused an increase in the flow velocity of Skeidarárjökull (Fig. 11) (Martinis et al., 2007; Sigurðsson et al., 2014). This acceleration occurred over the entire width of the glacier, and suggests that basal lubrication had a glacier-wide impact on ice dynamics (Sigurðsson et al., 2014). By contrast, following the 1996 Gjálp eruption (Section 2.7.2.), subglacial meltwater drainage and storage (in Grímsvötn subglacial lake) led to localised supraglacial subsidence, but elsewhere the glacier surface remained intact, suggesting that widespread basal sliding was not triggered (Gudmundsson et al., 1997). A possible explanation is that during the 2004 eruption of Grímsvötn, meltwater spread across the glacier bed via distributed, inefficient subglacial drainage, while following the 1996 Gjálp eruption water quickly formed, and drained through, a much more efficient (perhaps pre-existing) subglacial network (Fig. 2a, point 10). Gudmundsson et al., (1997) suggest that the formation of ice surface cauldrons over the eruptive bedrock fissure during the second event may have resulted in steep gradients in basal water pressure, towards the cauldrons (where overburden pressure was reduced), thus limiting the amount of meltwater able to reach the ablation area of the glacier. However, whether such gradients are sufficient to substantially reduce subglacial water drainage is open to question (see Smellie, 2009).

3.2.2. What determines whether volcanic impacts on glaciers are observed?

Observations of volcano-glacier interaction are partly limited by event size and duration (Section 3.2.1.), as well as the weather conditions during periods of activity; the availability of aerial and satellite imagery; the accessibility of the sites; whether events occur supraglacially or subglacially; and luck (e.g., whether aeroplanes pass close to volcanoes during periods of activity). Thus, in this review, by emphasising 'observed'/'documented' interactions we undoubtedly underestimate the real importance and frequency of volcanic impacts on glaciers, and this is likely to be particularly true for certain types of impacts (e.g., those occurring subglacially). Direct observation was also more difficult before the widespread availability of satellite data, and probably means that events that occurred before the 1970s (when the Landsat satellites were first launched) are dramatically underrepresented. Also, events in comparatively accessible and populated regions (e.g., Iceland) are better represented, and documented in more detail, than in isolated areas (e.g., Kamchatka or the sub-Antarctic Islands). In fact, even with the widespread availability and use of remotely sensed data, some regions are still difficult to robustly and repeatedly observe, often because of limitations with obtaining repeated, cloud-free imagery (e.g., the sub-Antarctic Islands, Patrick and Smellie, 2013). It is also likely that during volcanic events, ash and steam further limit visibility. In all these instances, most of our understanding of the events and their impacts on glaciers is inferred from conditions following the event. The spatial resolution of available remotely sensed imagery also regulates whether events (and which events in particular) are observed. For example, ice surface channels cut by supraglacial lava flows might be too narrow (a few metres wide) to be observed from many satellite sources (e.g., Landsat).

In all, volcanic impacts on glaciers are likely dramatically underrepresented in the observed record, and, in some cases, volcanic impacts on glaciers, though observed, may not have been recognised as such. This is likely to be particularly true at the ice sheet scale, where recognising links between ice dynamics and subglacial volcanic activity is difficult (Section 3.3.). Despite these limitations, over coming years, the number and quality of observations will likely quickly increase, as further high-resolution remote sensing datasets become available (Section 4.).

3.3. What is the future importance of volcanic impacts on glaciers?

In the long-term, as ice masses globally continue to retreat in response to climate warming (Bliss et al., 2014; Radić et al., 2014), their interactions with volcanoes are likely to become less frequent. However, in the short-term, glacier retreat and associated unloading may trigger explosive volcanic activity (Huybers and Langmuir 2009; Watt et al., 2013; Praetorius et al. 2016), with considerable impacts on existing ice masses. In addition, while deglaciation may be widespread, hundreds of volcanoes will remain ice covered over the next decades to centuries (Curtis and Kyle, 2017). Thus, and despite intrinsic difficulties, understanding, predicting and quantifying future volcanic impacts on glaciers is extremely important.

For example, it has been suggested that future subglacial volcanic activity could lead to enhanced basal melt, increased ice flow, and overall instability of the West Antarctic Ice Sheet (Blankenship et al. 1993; Vogel et al. 2006; Corr and Vaughan 2008; de Vries et al., 2017), with global implications, including sea level rise. In addition, many settlements, particularly in places such as the Andes, are located at the foot of ice-covered volcanoes (Pierson et al., 1990; Thouret, 1990). Future eruptions, perhaps exacerbated by climatically driven glacier unloading, could trigger floods/lahars of extreme magnitude, with devastating impacts on these communities (e.g., during the 1985 eruption of Nevado del Ruiz, when a lahar killed > 23000 people, Pierson et al., 1990). Indeed, investigating future volcanic impacts on glaciers is vital if we are to better mitigate associated hazards (Blong, 1984; Tuffen, 2010; Iribarren Anacona et al, 2015; Carrivick and Tweed, 2016). In the longer-term, if interactions with volcanic activity facilitate complete glacier loss, freshwater availability to communities in and around ice-covered volcanoes is likely to be considerably reduced (particularly outside rainy seasons), with notable impacts on human health and wellbeing (Beniston, 2003).

4. Future research directions

Predicting future volcanic impacts on glaciers (Section 3.3.) requires an improved understanding of their interactions. One clear way of achieving this is simply to make more, and more detailed, observations. Part of this process will involve continued investigation of volcano-glacier interactions during the Quaternary (Smellie and Edwards, 2016). For events that occurred within recent decades (i.e., since the 1970s), there is further scope for systematically searching archival satellite and airborne imagery for undocumented instances of volcano-glacier interaction. In monitoring future events, though ground-based studies (including the use of ground penetrating radar) will be important for observing volcanic-glacier interactions, including subglacial environments, developments in satellite and airborne (including drones) remote sensing are likely to drive the greatest advances in our understanding (see Harris et al., 2016, and papers therein).

4.1. Satellite remote sensing

Recent improvements in the quality and availability of satellite data have opened up opportunities for remotely observing volcano-glacier interactions at unprecedented spatial and temporal scales. Improvements in spatial resolution allow smaller events and features to be observed, while better temporal resolution (i.e., shortening the time interval between image capture at a given location) potentially allows events to be documented as they occur (e.g., over hours to months), and increases the likelihood of obtaining cloud-free images (see Patrick et al., 2016). Recent improvements in satellite data are already allowing global databases of glaciers (e.g., Pfeffer et al., 2014) and volcanoes (e.g., Global Volcanism Program, 2013) to be compiled, and automated (or semi-automated) techniques for near-real-time volcano monitoring to be developed. For example, the MODIS Volcano Thermal Alert System (MODVOLC) is an algorithm that allows near-real-time detection of global lava flows (Wright et al., 2004, 2015). Similar systems for the automated detection, mapping and monitoring of volcanic impacts on glaciers are in their infancy (e.g., Curtis and Kyle, 2017), but will undoubtedly see notable developments over coming years. Despite this progress, observations based on satellite data will continue to be biased towards larger events and particularly those that occur (or are expressed) supraglacially. Smaller events will continue to necessitate the use of other means of observation (particularly airborne remote sensing—Section 4.2), and detailed regional/local observations will continue to represent an ideal source of information and a way to validate near-global analyses.

4.2. Airborne remote sensing

Airborne remote sensing has been a particularly useful means of documenting volcano-glacier interactions over the 20th and early 21st centuries. Historically many of these observations were fortuitous (e.g., pilots observing volcanic activity as they pass by) rather than occurring during flights targeted specifically at observing volcanoes/glaciers. For example, passing airline pilots reported the onset of eruptive activity on the glacier-occupied Mount Westdahl in 1991 (Doukas et al., 1995), and a local pilot reported heavily crevassed ice on the upper section of Chestnina Glacier following suspected volcanic activity at Mount Wrangell in 1999 (McGimsey et al., 2004). Similarly fortuitous observations are likely to continue in the future, but detailed descriptions of eruptions and their glaciological impacts will rely on targeted flights, flown for the express purpose of documenting events (e.g., Gudmundsson et al., 2007; Stewart et al., 2008; Magnússon et al., 2012). One notable direction for future progress is the use of unmanned aerial vehicles (UAVs), which allow centimetre-scale images to be captured in a quick and cost-effective way (Westoby et al., 2012). Such high-resolution data will allow some of the smaller (less conspicuous) events and features arising from volcano-glacier interactions to be better documented. UAVs are particularly useful in that they can be used to safely monitor otherwise dangerous/inaccessible sites (Corrales et al., 2012; Diaz et al., 2015), and potentially allow for near-continuous and repeat observations. This approach will flourish over coming years, but the use of UAVs requires an operator on the ground, and is therefore only likely during targeted periods of observation. In addition, difficulties with observing subglacial environments are likely to persist.

5. Conclusions

In this paper, we review volcanic impacts on modern glaciers (since AD 1800), supported by a global dataset of examples (Supplementary Table 1). The main findings can be summarised as follows:

1. Instances where volcanic activity has a documented impact on the behaviour of modern glaciers are comparatively rare. However, because of difficulties with observing these events, it is likely that their frequency and importance are underestimated.
2. Shorter-term (days-to-months) impacts are typically destructive, whilst longer-term (years-to-decades) impacts are likely to include reduced ablation, glacier stagnation and/or advance.

3. Predicting the future importance of volcanic impacts on glaciers is difficult because our understanding of their interactions is limited, and because the frequency and nature of volcano-glacier interactions is likely to change with time (e.g., future glacier retreat may lead to an increase in explosive volcanic activity). However, there is considerable interest in this area because volcanic activity may play a role in regulating the future stability of ice sheets (such as the West Antarctic Ice Sheet), and because there is a need to better mitigate future glacio-volcanic hazards (e.g., floods and lahars).
4. Fortunately, due to improvements in the availability and quality of remotely sensed data, future observations of volcanic impacts on glaciers are likely to be more frequent, and descriptions of these interactions more detailed. However, observations will continue to be biased towards larger events, and monitoring subglacial processes (in particular) is likely to remain challenging.

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