How will melting of ice affect volcanic hazards in the 21st century?

Hugh Tuffen

Lancaster Environment Centre, Lancaster University, Lancaster LA1 4YQ, UK.

h.tuffen@lancaster.ac.uk

Abstract

Glaciers and ice sheets on many active volcanoes are rapidly receding. There is compelling evidence that melting of ice during the last deglaciation triggered a dramatic acceleration in volcanic activity. Will melting of ice this century, which is associated with climate change, similarly affect volcanic activity and associated hazards?

This paper provides a critical overview of the evidence that current melting of ice will increase the frequency or size of hazardous volcanic eruptions. Many aspects of the link between ice recession and accelerated volcanic activity remain poorly understood. Key questions include how rapidly volcanic systems react to melting of ice, whether volcanoes are sensitive to small changes in ice thickness, and how recession of ice affects the generation, storage and eruption of magma at stratovolcanoes. A greater frequency of collapse events at glaciated stratovolcanoes can be expected in the near future, and there is strong potential for positive feedbacks between melting of ice and enhanced volcanism. Nonetheless, much further research is required to remove current uncertainties about the implications of climate change for volcanic hazards in the 21st century.

Key index words or phrases

Volcanic hazards, climate change, volcano-ice interaction, ice sheets, glaciers, lahars

1.1. Introduction

There is growing evidence that past changes in the thickness of ice covering volcanoes has affected their eruptive activity. Dating of Icelandic lavas has shown that the rate of volcanic activity in Iceland accelerated by a factor of 30-50 following the last deglaciation at ~12 ka (Maclennan et al. 2002). Analyses of local and global eruption databases have identified a statistically significant correlation between periods of climatic warming associated with recession of ice and an increase in the frequency of eruptions (Jellinek *et al.* 2004, Nowell *et al.* 2006, Huybers and Langmuir 2009). Today the bodies of ice found on many volcanoes are rapidly thinning and receding. These bodies range from extensive ice sheets to small tropical glaciers and thinning is thought to be triggered by contemporary climate change (e.g. Rivera *et al.* 2006, Vuille *et al.* 2008, Björnsson and Pálsson 2008).

This leads to the following question: will the current ice recession provoke increased volcanic activity and lead to increased exposure to volcanic hazards? In this paper I analyse our current knowledge of how ice thickness variations influence volcanism and identify several unresolved issues that currently prevent quantitative assessment of whether activity is likely to accelerate in the coming century. These include the poorly-constrained response time of volcanic systems to unloading of ice, uncertainty about how acceleration in volcanic activity scales to the rate and total amount of melting, and the lack of models to simulate how melting of ice on stratovolcanoes may affect magma storage and eruption to the surface. In conclusion I highlight some of the future research needed for better understanding of how melting of ice may force volcanic activity.

1.2. What are hazards at ice and snow-covered volcanoes and where are they found?

Many volcanoes are mantled by ice and snow, especially those located at high latitudes or that reach over 4000 m in altitude. Notable examples occur in the Andes, the Cascades, the Aleutian-Kamchatkan arc, Iceland, Antarctica, Japan and New Zealand (Fig. 1, Fig. 2). The nature of ice and snow cover spans a broad spectrum from seasonal snow (Mee *et al.* 2006), small bodies of ice and firn in summit regions (Houghton *et al.* 1987, Julio-Miranda *et al.* 2008), larger alpine glaciers on volcano flanks (Fig. 1b, Fig. 2a, e.g. Rivera *et al.* 2006, Vuille *et al.* 2008), thick ice accumulations within summit craters and calderas (e.g. Gilbert *et al.* 1996, Huggel *et al.* 2007a), to substantial ice sheets that completely cover volcanic systems (Fig. 2b, e.g. Guðmundsson *et al.* 1997, Corr and Vaughan 2008).

Historical eruptions at more than 40 volcanoes worldwide have involved disruption of ice and snow (Major and Newhall 1989), whereas numerous geological studies have enabled the recognition of interactions between volcanoes and ice or snow in ancient eruptions (e.g. Noe-Nygaard 1940, Mathews 1951, Gilbert *et al.* 1996, Smellie 1999, Lescinsky and Fink 2000, Mee *et al.* 2006). Volcanic deposits provide an invaluable record of palaeo-environmental change, such as fluctuations in ice thickness and extent (Smellie *et al.* 2008, Smellie 2008, Tuffen *et al.* 2010), as well as the processes and hazards associated with various types of volcano-ice interaction (e.g. Smellie and Skilling 1994, Smellie 1999, Lescinsky and Fink 2000, Tuffen and Castro 2009, Carrivick 2007).

1.3. Hazards at ice- and snow-covered volcanoes

The presence of ice and snow on volcanoes can greatly magnify hazards, principally because perturbation of ice and snow during eruptions can rapidly generate large volumes of meltwater that are released in destructive lahars and floods. Major and Newhall (1989) compiled a comprehensive global review of historical eruptions at more than 40 volcanoes during which ice and snow were perturbed and lahars or floods generated. Major loss of life occurred in several eruptions, including Nevados de Ruiz (Columbia, 1985), Villarrica (Chile, 1971), Tokachi-dake (Japan, 1926) and Cotopaxi (Ecuador, 1877).

Pertubation of ice and snow by volcanic activity. Major and Newhall identified five distinct mechanisms that can cause perturbation of snow and ice or volcanoes: (1) mechanical erosion and melting by flowing pyroclastic debris or blasts of hot gases (e.g. Walder 2000), (2) melting of the ice or snow surface by lava flows (e.g. Mee *et al.* 2006), (3) basal melting by subglacial eruptions or geothermal activity (e.g. Guðmundsson *et al.* 1997), (4) ejection of water by eruptions through a crater lake, and (5) deposition of tephra onto ice and snow (e.g. Capra *et al.* 2004).

Subsequently, observations of volcanic activity in Columbia, Iceland, USA, New Zealand and Alaska (Waitt 1989, Pierson *et al.*, 1990, Guðmundsson *et al.* 1997, 2004, 2008; Carrivick *et al.* 2009*a*) have highlighted how rapidly meltwater may be generated during melting of the base of ice sheets and glaciers, and when pyroclastic debris move over ice and snow. Melting rates may exceed 0.5 km^3 per day during powerful subglacial eruptions (e.g. Guðmundsson *et al.* 2004). The hazards associated with meltwater production are exacerbated when transient accumulation occurs with craters or calderas, as this can lead to even higher release rates of meltwater when catastrophic drainage is triggered by dam collapse or floating of an ice barrier that allows rapid subglacial drainage (Pierson *et al.* 1990, Guðmundsson *et al.* 1997, Carrivick *et al.* 2004, 2009*a*).

The magnitude of meltwater floods (jökulhlaups and lahars) can exceed 40 000 $m^3 s^{-1}$ (Major and Newhall 1989, Pierson *et al.* 1990, Guðmundsson *et al.* 2004, 2008), creating significant hazards in river valleys and on outwash plains many tens of kilometres from the site of melting (Fig. 1c,d; Fig. 2b; e.g. Pierson *et al.* 1990, Eliasson *et al.* 2006, Huggel *et al.* 2007*a*). The total volume of meltwater floods may be restricted by either the amount of pyroclastic material or lava available to cause melting, or the volume of ice and snow that can be melted.

Explosive eruptions. The hazards posed by explosive eruptions at ice- and snow-covered volcanoes are typical of those at other volcanoes, with the following important modifications: 1) Interactions between magma and meltwater may trigger phreatomagmatic activity (Fig. 1a), even during basaltic eruptions that would not otherwise be explosive (e.g. Smellie and Skilling 1994, Guðmundsson *et al.* 1997). 2) When ice is thick the explosive phase of eruptions may partly or entirely take place beneath the ice surface (Tuffen 2007), reducing the hazards associated with ashfall and pyroclastic debris. 3) If explosive eruptions do occur then widespread perturbation of ice and snow by pyroclastic material may be important, both at the vent area and in more distal areas.

Edifice instability and collapse. Ice-and snow-covered volcanic edifices are especially prone to collapse, creating hazardous debris avalanches that may convert to lahars (e.g. Huggel *et al.* 2007*a,b*) and reach many tens of kilometres from their source. Collapse is favoured by 1) constraint by ice, which may encourage the development of structurally unstable, oversteepened edifices, 2) melting of ice, which may create weak zones at ice-bedrock interfaces (Huggel 2009) and 3) shallow hydrothermal alteration driven by snow and ice melt, which can greatly weaken volcanic edifices (e.g. Carrasco-Núñez *et al*., 1993, Huggel 2009). Ice avalanches are a newly-recognised phenomenon that may occur at ice-covered volcanoes (Fig. 2a; Huggel *et al.* 2007*b*). Ice avalanches ranging from 0.1 to 20×10^6 m^3 in volume originate from steep areas near the summit of Iliama volcano, Alaska, where the geothermal flux is high. These avalanches travel up to 10 km down the volcano flanks at speeds of 20-70 m s^{-1} (Huggel 2009). The thermal perturbations that can trigger slope failure include volcanic/geothermal, glacier-permafrost and climatically-induced warming.

The distribution of hazards at active ice and snow-covered volcanoes such as Citlaltepetl, Mexico (lahars; Hubbard *et al.* 2007), Nevado de Ruiz, Columbia (lahars, avalanches; Huggel *et al.* 2007*a*), Mt Rainier, Washington (lahars; Hoblitt *et al.* 1998), Ruapehu, New Zealand (lahars; Houghton *et al.* 1987), Iliama, Alaska (lahars, avalanches; Waythomas and Miller 1999, Huggel *et al.* 2007*b*) and Katla, Iceland (jökulhlaups, Björnsson *et al.* 2000) reflect these different sources of volcanic hazard, principally meltwater floods, which potentially affect millions of people living close to these volcanoes.

2. How is ice thickness on volcanoes currently changing?

Rapid thinning and recession of ice has been noted on many active and potentially-active volcanoes, including Popocatepetl and other Mexican volcanoes (Julio-Miranda *et al.* 2008), Columbian stratovolcanoes (Huggel *et al.* 2007*a*), Villarrica and other Chilean volcanoes (Rivera *et al.* 2006) and Kilimanjaro, Tanzania (Fig. 3; Thompson *et al.* 2009). Ice sheets covering volcanic systems are also rapidly thinning, including Vatnajökull in Iceland (Björnsson and Pálsson 2008) and parts of the

West Antarctic Ice Sheet (Wingham *et al.* 2009). Selected measured or estimated rates of ice thinning and recession are provided in Table 1.

Whereas the changing mass balance of thick ice sheets is predominantly manifested in a reduction in ice surface elevation and therefore in ice thickness (e.g. Wingham *et al.* 2009), the surface area of smaller glaciers on many volcanoes is rapidly reducing, along with a rapid decrease in ice volume. Rates of thinning vary from 0.54 m a^{-1} on Kilimanjaro (Thompson *et al.* 2009) to 1.6 m a^{-1} (Pine Island Glacier, West Antarctic Ice Sheet; Wingham *et al.* 2009). Assuming that current rates of ice loss continue over the coming century, ice bodies on numerous volcanoes may therefore thin by ~50 to 150 metres by 2100. Stratovolcanoes hosting thin glaciers, such as Kilimanjaro, may therefore become completely ice-free in the coming century (Thompson *et al.* 2009). At some volcanoes this has already occurred, such as at Popocatepetl, Mexico where dramatic extinction of summit ice over the last 50 years reached completion in 2004 (Julio-Miranda *et al.* 2008).

Volcano	Last eruption	Area or volume of ice	Rate of thinning	Reference
Vatnajökull, Iceland	2004, 1998, 1996	A = 8100 km^2 V = 3100 km^3 (in 2000)	0.8 m a^{-1} average (1995-2008) Geothermal melting and eruptions melted 0.55 km^3 a^{-1}, annual surface ablation 13 km^3 a^{-1}.	Pagli and Sigmundsson 2008, Björnsson and Pálsson 2008
Volcán Villarrica, Chile	2005, 2007, 2008	A = 30.3 km^2	0.81 ±0.45 m a^{-1} (1961-2004)	Rivera *et al.* 2006
Popocatepetl, Mexico	1994-2001	Was 0.729 km^2 in 1958, now 0 km^2	1996 ~0.2 m a^{-1} 1999 ~4 m a^{-1}	Julio-Miranda et al 2008
Nevado del Ruiz, Columbia	1991	A = 19-25 km^2 (1985), 10.3 km^2 (2002-2003)	Not known	Ceballos *et al.* 2006, Huggel *et al.* 2007a
Cotopaxi, Ecuador	1940	A = 19.2 km^2 (1976), 13.4 km^2 (1997)	3-4 m a^{-1} on snouts	Jordan *et al.* 2005
Kilimanjaro, Kenya/Tanzania	150-200 ka	A = 2.5 km^2 (2000), 1.85 km^2 (2007)	0.54 m a^{-1}	Thompson *et al.* 2009
West Antarctic Ice Sheet	~2 ka?	V = 2.2 x 10^6 km^3	Pine Island Glacier ~1.6 m/a, accelerating over 1995-2006.	Lythe and Vaughan 2001, Shepherd et al 2001, Corr and Vaughan 2008, Wingham et al 2009.

Table 1. Current estimated rates of ice thinning and recession at selected volcanoes and ice sheets.

2.1. Ice thinning due to climate change

Much of the current recession and thinning of glaciers and ice sheets covering volcanoes is attributed to the effects of global climate change, with increasing mean temperature and in some cases decreasing precipitation leading to negative glacier mass balance changes (e.g. Rivera *et al.* 2006, Bown and Rivera 2007, Vuille *et al.* 2008). The equilibrium line altitude (ELA) is the altitude on a glacier where the annual accumulation and ablation rates are exactly balanced. The ELA of glaciers on Chilean stratovolcanoes such as Villarrica has migrated upwards by ~100 m between 1976 and 2004/2005 (Rivera *et al.* 2006), partly due to a mean temperature increase at 2000 m elevation of

0.023 °C a^{-1} (Bown and Rivera 2007). Ice thinning on Popocatepetl between 1958 and 1994 is likewise thought to be related to climatic change (Julio-Miranda *et al.* 2008), as is dramatic thinning of tropical mountain glaciers on Ecuadorian volcanoes such as Antizana and Cotopaxi (Vuille *et al.* 2008)

Thinning of Vatnajökull ice sheet in Iceland is also pronounced, with an average thinning of 0.8 m a^{-1} between 1995 and 2008 (Björnsson and Pálsson 2008, Pagli and Sigmundsson 2008). Future prediction of mass balance changes at Vatnajökull in the coming century, which incorporate glacier dynamic models with predicted increases in mean temperature (2.8 °C) and precipitation (6 %) project a 25 % volume loss by 2060 (Björnsson and Pálsson 2008). The effects of climate change on the mass balance of glaciers and ice sheets on volcanoes is likely to be strongly location-specific. This is because changes in temperature and precipitation are spatially heterogeneous and glacier dynamics highly variable. This local sensitivity is illustrated by the contrasting recession rates of glaciers on neighbouring Chilean volcanoes less than 50 km apart (e.g. Rivera *et al.* 2006, Bown and Rivera 2007) and highlights the importance of studying individual ice-covered volcanoes, rather than applying a regional or global climate change model (Huybers and Langmuir 2009) to predict local changes in ice thickness on specific volcanoes.

2.2. Ice thinning due to volcanic and geothermal activity

Volcanic and geothermal activity can strongly influence the mass balance of ice bodies on volcanoes both during eruptions and periods of quiescence. The “background” ablation and accumulation rates determine the overall effects of volcanic and geothermal activity on glacier mass balance and dynamics (Magnusson *et al.* 2005, Guðmundsson *et al.* 2009). Mechanisms of ice loss include basal melting and ice disruption during and after subglacial eruptive activity (Fig. 4a; e.g. Guðmundsson *et al.* 1997, Jarosch and Guðmundsson 2007), melting of the ice and snow surface by the heat of erupted debris (Julio-Miranda *et al.* 2008), changes to surface albedo due to tephra cover (Fig. 4b; Rivera *et al.* 2006), and lubrication by sustained basal melting due to geothermal heat (e.g. Bell 2008). Rapid melting, fracturing and mechanical erosion during eruptions can cause dramatic, localised thinning of ice above vents (Fig. 4a) and meltwater drainage pathways (e.g. Guðmundsson *et al.* 1997), with removal of tens or hundreds of metres of ice in hours. Perturbations to the ice surface may be transient, however, as depressions formed may swiftly fill due to increased snow deposition and inward deformation of surrounding ice (Aðalgeirsdóttir *et al.* 2000).

There is strong evidence that volcanic and geothermal activity is hastening the demise of ice bodies on some volcanoes. Eruptive activity at Popocatepetl, Mexico from 1994 to 2001 led to the complete extinction of its small (<1 km^2) summit glaciers (Fig. 4b; Julio-Miranda *et al.* 2008). This extinction reflects the negligible accumulation at a volcano located in an intertropical zone, which makes its glacier mass balance extremely sensitive to eruption-triggered ablation. It is speculated that the disappearance of ice on Popocatepetl was inevitable due to climate change, but greatly hastened by eruptive activity (Julio-Miranda *et al.* 2008). Recent changes in the mass balance of glaciers on Villarrica volcano, Chile reflect the effects of tephra cover on the ice surface (Rivera *et al.* 2006). At Villarrica mass balance is also strongly influenced by basal geothermal fluxes. Tephra cover drives enhanced melting when tephra is thin due to enhanced heat absorption, but thicker layers may insulate ice and snow and reduce melting (Rivera *et al.* 2006, Brock *et al.* 2007).

Melting and mechanical removal of ice during the 1985 eruption of Nevado del Ruiz, Columbia removed approximately 10 % of the volume of ice on the volcano (Ceballos *et al.* 2006, Huggel *et al.* 2007*a*), which totalled 0.48 km^3 in 2003. This demonstrates that a single moderately large volcanic eruption (VEI 3) can have an appreciable impact on the mass balance of ice on

Andean stratovolcanoes, due to the low ice accumulation rates on tropical glaciers (Ceballos *et al.* 2006).

By contrast, even considerable volcanically-triggered melting probably has a negligible effect on the mass balance of Iceland's Vatnajökull ice sheet over a decadal timescale. This is because Icelandic glaciers and ice sheets are characterised by high annual accumulation and ablation due to temperate conditions and extremely high precipitation (Björnsson and Pálsson 2008). At Vatnajökull on average 0.55 km^3 a^{-1} was melted by volcanic eruptions during the period 1995-2008, but this amounted to only 4 % of the total surface ablation from the ice sheet during this period (13 km^3, Björnsson and Pálsson 2008). However, the effects of geothermal heat fluxes may have significant effects on ice dynamics and mass balance over both long and short timescales: models of the volume of Vatnajökull at the last glacial maximum are highly sensitive to basal geothermal heat fluxes (Hubbard 2006) and eruption-triggered jökulhlaups may also trigger surging, which affects glacier mass balance (Björnsson 1998, Björnsson and Pálsson 2008).

3. How has ice recession affected volcanic activity in the past?

3.1. Evidence for accelerated volcanism triggered by deglaciation

There is strengthening quantitative evidence linking periods of deglaciation with increased volcanic activity in many different volcanic settings. The best established and most dramatic acceleration in activity occurred in Iceland, where vigorous volcanism is strongly affected by a temperate ice sheet that may almost completely cover the island during glacial periods and almost completely disappear during interglacials (Björnsson and Pálsson 2008). Unloading of hundreds of metres to 2 km of ice during deglaciation in Iceland causes decompression that, according to current models, leads to a greater degree and depth range of mantle melting (Jull and McKenzie 1996, Maclennan *et al.* 2002). This is reflected in a 30- to 50-fold increase in the rate of magma eruption on individual volcanic systems in the 1.5 ka after the deglaciation of each area, inferred from the volume of erupted deposits (Fig. 5; Maclennan *et al.* 2002). The short time delay between inferred ice unloading and enhanced volcanism shows that the “extra” magma generated is rapidly transported from source to surface without prolonged storage in magma chambers, so that Icelandic volcanism responds swiftly to changes in ice thickness. In most other volcanic settings magma accumulation in chambers is the norm (e.g. volcanic arcs), in which case the mechanism for enhanced volcanism may differ. It may reflect the response of magma chambers to unloading, rather than the eruption of primitive melts directly to the surface.

Statistical analyses of eruption databases have shown quantitatively that patterns of volcanic activity elsewhere are also influenced by changes in ice thickness: both globally (Huybers and Langmuir 2009, Fig. 6a), in Eastern California (Jellinek *et al.* 2004, Fig. 6b) and in western Europe (Nowell *et al.* 2006). It is important to note that most statistical studies use a global climate proxy from marine $\delta^{18}O$ records as an indication of ice thickness changes, rather than local ice thickness changes on volcanoes themselves (which are poorly constrained). Further, only the number of eruptions is considered in analyses, rather the volume of eruptions. Huybers and Langmuir (2009) used a database of global eruptions in the last 40 ka (Siebert and Simkin 2002) to calculate the change in frequency of eruptions with VEI>2 prior to, during and after the last deglaciation. The increase in volcanic activity during deglaciation above modern values was found to be statistically highly significant ($p < 0.01$) and activity during deglaciation (18-7 ka) was significantly higher than glacial rates between 40-20 ka. Although there are doubts about the completeness of the eruption record, interesting trends emerge from the data. The timing of enhanced volcanism differs between localities (e.g. a global increase occurred at ~18 ka, but occurred later in Iceland, at ~12 ka). This may reflect differing regional deglaciation histories, although other factors such as the delay between deglaciation and magma reaching the surface may also differ and depend upon the plumbing system of individual volcanic complexes. There is currently no discussion in the literature about whether the magnitude of volcanic eruptions increases during deglaciation, or whether it is only the frequency of eruptions that is affected.

Qualitative evidence for accelerated volcanism at individual volcanic complexes during deglaciation includes studies at Mt Mazama, western USA (Bacon and Lamphere 2006) and three Chilean volcanoes: Lascar, Puyehue and Nevados de Chillan (Gardeweg *et al.* 1998, Singer *et al.* 2008, Mee *et al.* 2009). However confidence about whether glacial-interglacial cycles truly influence eruptive activity is generally low, as there are insufficient dated eruptions at individual volcanoes to adequately test statistical significances. In some cases there is no obvious increase in activity during the last deglaciation (e.g. Torfajökull, Iceland; McGarvie *et al.* 2006).

3.2. Edifice collapse triggered by ice recession

A mechanistic link between deglaciation and collapse of ice-covered stratovolcanoes has been proposed by Capra (2008), who noted the coincidence between major edifice collapses and periods of rapid ice recession in the last 30 ka for 24 volcanoes, predominantly located in Chile, Mexico and the USA. Capra proposed that abrupt climate change resulting in rapid ice melting may trigger edifice collapses through glacial debuttressing and an increase in fluid circulation and humidity. However, more data is required to quantitatively test whether periods of rapid ice decline do indeed correlate with acceleration in the incidence of edifice collapse.

4. How does the rate and extent of current ice melting compare with past changes?

In order to assess whether the current changes in ice thickness and extent on many volcanoes are likely to trigger accelerated volcanic activity the current rate of melting must be compared with inferred rates of melting during the last deglaciation. Precisely reconstructing rates of ice thinning during the last deglaciation is problematic, due to the limits of resolution provided by proxies for changing ice extent and thickness. Furthermore, the history of deglaciation was complex, with major stepwise advances and retreats including the Younger Dryas event at 11-10 ka and the Preboreal Oscillation at 9.9-9.7 ka (Geirsdóttir *et al.* 2000).

Quoted "average" rates of deglaciation for Iceland, as used in mantle melting models, are 2 m a^{-1} (2 km in 1 ka; Jull and McKenzie 1996, Pagli and Sigmundsson 2008). Similarly, the mean rate of surface elevation change of the Laurentide ice sheet during early Holocene deglaciation is estimated at 2.6 m a^{-1} (Carlson *et al.* 2008). However, it is inappropriate to assume a constant rate of ice unloading, as phases of dramatic warming, such as the end of the Younger Dryas event, are likely to have involved much more rapid recession over shorter time intervals. Indeed, there is geological evidence for bursts of considerably faster deglaciation during abrupt warming events (e.g. 100 m a^{-1} in Denmark between 18-17 ka, Humlum and Houmark-Nielsen, 1994). Rapid deglaciation in volcanically active areas could be further driven by positive feedback, with eruption-triggered jökulhlaups potentially playing an important role in glacier break-up (Geirsdóttir *et al.* 2000, Carrivick *et al.* 2009b).

Nonetheless, it is informative to compare data: the current rates are mostly about 20-40 % of the mean estimated deglaciation rates for the Icelandic and Laurentide ice sheets (Fig. 7). The extent of ice unloading is, however, very different, as rapid unloading has only occurred since the end of the Little Ice Age. For example, Vatnajökull in Iceland has only shrunk since 1890 (Björnsson and Pálsson 2008, Pagli and Sigmundsson 2008). This means that total thinning of only ~60 m has occurred at Vatnajökull since 1890, compared with ~2 km during the last deglaciation (Fig. 7). Tropical glaciers in the Andes reached their maximum extents of the last millennium at between 1630 and 1730 AD (Jomelli *et al.* 2009) and have only rapidly retreated since the middle of the 19th century. Therefore current thinning has only been sustained over a 100-200 year period, which is considerably shorter than major deglaciation events. As a consequence the total reduction in ice thickness to date during current warming is probably less than 10 % of that during major past deglaciation events.

However, there are marked local and regional discrepancies in how rapidly ice sheets and glaciers have receded during current (post-Little Ice Age) warming. Over 3030 km^3 if ice has been lost from Glacier Bay, Alaska since 1770 (Larsen *et al.* 2005), with local thinning of up to 1.5 km at a mean rate of up to 6.5 m a^{-1}. This value is more comparable to changes during the main phases of deglaciation, but has not occurred in an active volcanic region.

5. How might hazards be affected by melting of ice and snow?

5.1. Ice unloading may encourage more explosive eruptions

The explosivity of eruptions beneath ice sheets is restrained by thick ice, as high glaciostatic pressures (>5 MPa) inhibit volatile exsolution (Tuffen *et al.* 2010) and rapid ice deformation can close cavities melted at the base of the ice, encouraging intrusive rather than explosive activity (Fig. 8; Tuffen *et al.* 2007). Thinning of ice covering a volcano may therefore encourage more explosive eruptions, which generate meltwater more rapidly than intrusive eruptions (Guðmundsson 2003) and, if the ice surface were breached, create hazards associated with tephra. Where ice is thin (<150 m) there is generally comparatively little interaction between magma and meltwater, as thin ice fractures readily, offers little constraint to the force of eruptions and is inefficient at collecting meltwater around the vent (Smellie and Skilling 1994, Smellie 1999). Thinning of ice may therefore generally lead to more explosive eruptions at volcanoes that are currently covered by substantial thicknesses of ice (>300 m), especially those with deep ice-filled summit calderas such as Sollipulli, Chile (Gilbert *et al.* 1996) and Katla, Iceland (Björnsson *et al.* 2000). It is important to note, however, that there is currently no quantitative relationship between eruption explosivity and ice thickness. The models quoted only simulate a small part of the coupled volcano-ice system and thus are essentially qualitative; they do not incorporate feedbacks between the dynamics of magma storage, ascent and the response of the overlying ice.

5.2. Ice unloading and increased melting may trigger edifice stability

It has been hypothesised that melting and recession of ice on volcanic edifices may lead to instability and edifice collapse due to two independent mechanisms: firstly, debuttressing and the withdrawal of mechanical support from ice (Capra 2008) and secondly, an increase in the pore fluid pressure within shallow hydrothermal systems, which may trigger movement on pre-existing weaknesses (Capra 2008). However, this hypothesis currently remains unproven due to insufficient data. A significant proportion of glacier meltwater may enter the hydrothermal system of volcanoes (e.g. Antizana volcano, Ecuador, Favier *et al.* 2008). Seasonal seismicity at volcanoes such as Mt Hood (USA) is consistent with seismic triggering by an increase in meltwater input (Saar and Manga 2003), illustrating that movement on pre-existing weaknesses is favoured by enhanced meltwater production.

5.3. Melting of ice and snow may decrease the likelihood and magnitude of meltwater floods

As the volume of ice and snow on a volcano decreases, the size of the reservoir of potential meltwater decreases. At volcanoes where a relatively small volume of ice and snow is present the total volume of lahars may be restricted by the volume of ice and snow available for melting (Huggel *et al.* 2007*a*). This leads to the following qualitative prediction: as this volume decreases the total volume and magnitude of meltwater floods should decrease for a given size of eruption, thus reducing the associated hazards. Björnsson and Pálsson (2008) have shown that meltwater discharge from thinning Icelandic glaciers is likely to peak in 2040-2050 as ablation rates rise, but thereafter recede, reflecting the diminished volume of ice available for melting. Futhermore, as meltwater floods are triggered when tephra falls onto ice and snow (Major and Newhall 1989, Walder 2000, Julio-Miranda *et al.* 2008), a reduction in the area of ice and snow will reduce the probability that this will occur, therefore reducing the incidence of lahar generation. However, if the size of eruptions were to increase then in some cases a dwindling ice volume would not prevent an increase in the magnitude of meltwater floods, as recognised by Huggel *et al.* (2007*a*).

6. What are the likely effects of 21st century climate change on hazards at ice-covered volcanoes?

Unloading as ice and snow melt may trigger increased volcanic activity. Vexed questions include how quickly volcanic systems respond to ice thickness changes, which baselines for rates of volcanic activity are appropriate for the Holocene, and how to scale past accelerations in volcanic activity to changes in the 21st century.

6.1. Increased magma production and eruption in Iceland?

Melting of Icelandic ice sheets leads to increased mantle melting and eruption of magma to the surface (Fig. 5; Jull and McKenzie 1996, Maclennan *et al.* 2002, Pagli and Sigmundsson 2008). It is estimated that melting of Vatnajökull between 1890 and 2003 (435 km^3 loss, with a thinning rate of ~0.5 m a^{-1}) led to a 1 % increase in the rate of magma production (Pagli and Sigmundsson 2008). If current melting rates continue throughout the 21st century a roughly similar additional rise in melt production may be anticipated. Any increase in the thinning rate would trigger a stronger acceleration in melt production. It is important to note, however, that the rate and amount of ice thinning are far lower than during the last deglaciation (Fig. 7), and the projected increases in the rate of melt production are far weaker (at most a few percent increase, as opposed to a 30-50 fold increase). Although studies have shown that additional melt was transported to the surface at a rate of over 50 m a^{-1} (Maclennan *et al.* 2002), this only constrains the timescale of melt extraction to being <2 ka.

There is incomplete evidence collected to date, but some preliminary data suggests that the timing of Icelandic volcanism during deglaciation may have coincided with rapid warming events, indicating a short delay between extra melting and eruption to the surface. The timing of large tuya-building eruptions in north Iceland appear to correspond with two most marked warming events during deglaciation – the Bolling warming and the end of the Younger Dryas (Licciardi *et al.* 2007).

We therefore have insufficient knowledge to predict whether the "extra" melt generated would be erupted to the surface in the 21st century and whether any statistically significant increase in activity should be anticipated.

6.2. Increased magma production and eruption globally?

The pioneering study by Huybers and Langmuir (2009) attempts to relate changes in global volcanic activity during deglaciation to estimates of the rate of ice unloading. In it they use a simple glacier mass balance model to estimate modern changes in ice thickness at a number of glaciers. This model considers only relative accumulation vs ablation rates and ignores the ice dynamical processes (e.g. Bell 2008, Wingham *et al.* 2009) and local variations in precipitation and temperature (e.g. Vuille *et al.* 2008) that strongly influence mass balance and ice sheet profiles (e.g. Hubbard 2006). The results of eruption datasets are used to calculate glacial/deglacial and deglacial/Holocene eruption frequency ratios (Fig. 6a). Volcanoes with a current strong negative ice volume balance are excluded from the analysis as they are assumed not to have been ice-covered during the late Pleistocene, and therefore insignificant ice unloading is assumed to have occurred during deglaciation. Analysis of the eruption frequency of volcanoes considered to have been ice-covered then produces an enhancement in the rate of volcanic activity by a factor of 2 and 6 between 12 and 7 ka. These figures were generated using a -6 m a^{-1} and a -9 m a^{-1} cut-off, respectively.

Estimates of the amount of increased melting and magma eruption to the surface are very approximate. Huybers and Langmuir assume that unloading 1 km of ice above a 60 km thick melting region triggers a 0.1% increase in the melt percentage, and that 10 % of the melt then reaches the surface. They then estimate that 15 % of the 1.8×10^6 km^3 of ice lost from mountain glaciers between the last glacial maximum and today influences magma production. The validity of this percentage needs to be checked against the distribution of global ice loss from mountain glaciers, which is itself very difficult to constrain due to a lack of data and the complexity of local climatic variations (e.g. Vuille *et al.* 2008). The melting model also ignores diversity in melt zone depths and does not take in account crustal storage in magma chambers.

Elsewhere, Jellinek *et al.* (2004) examine statistical correlations between changes in ice thickness (assumed to be related to the time derivative of the SPECMAP $\delta^{18}O$ record) and the frequency (rather than magnitude) of documented volcanic eruptions in Eastern California (Fig. 6b). They found a significant correlation, with increased frequency of volcanism following periods of inferred glacier unloading. Models indicated a delay between unloading and increased volcanism of 3.2 ± 4.2 kyr and 11.2 ± 2.3 ka for silicic and basaltic eruptions respectively. Although local ice thickness fluctuations are unlikely to relate consistently or linearly to the oxygen isotope record, this analysis does point to intriguing differences between the rate of response to unloading between different magma types and volcanic plumbing systems. Similarly, Nowell *et al.* (2006) found evidence for accelerated volcanism during deglaciation of western Europe.

These studies indicate that a statistically significant correlation exists between unloading of ice and increased volcanism. However, as is the case for Iceland, the timescale of the response of volcanic systems to ice unloading is not well constrained. Data from Eastern California suggest that the volcanic response may be delayed by thousands of years. If this were the case, volcanism in the coming century may reflect changing ice thicknesses in the mid-to-late Holocene, rather than melting of ice since the Little Ice Age. Scaling issues are also problematic. There is considerable uncertainty about how the magnitude of acceleration in melt production and magma eruption to the surface scale to the amount and rate of ice unloading. A simple linear relationship between melt production and unloading (e.g. Huybers and Langmuir 2009) is not appropriate as the rate of melt production also depends on the previous loading and unloading history (Jull and McKenzie 1996). Furthermore, magma residence in chambers may decouple the timing of melt production from that of magma eruption to the surface.

6.3. Potential effects on volcanic hazards

An increase in the rate of magma eruption to the surface would entail larger and/or more frequent eruptions, thus increasing exposure to hazards. Indeed, analysis of tephra in the Greenland ice core (Zielinski *et al.* 1996) has shown that the greatest frequency of volcanic events in the last 110 ka occurred between 15 and 8 ka, closely corresponding to the timing of northern hemisphere deglaciation. The largest eruptions also occurred during a similar, overlapping interval, between 13 and 7 ka. To date most studies have focussed solely on the frequency of eruptions (e.g. Jellinek *et al.* 2004, Nowell *et al.* 2006, Huybers and Langmuir 2009). Increased eruptive frequency at a given volcano will increase risk exposure. The intensity and explosivity index (VEI) of eruptions also scale to their total volume (e.g. Newhall and Self 1982, Pyle 1999). There is currently insufficient evidence to determine whether the size or frequency of eruptions will increase in the 21st century.

The explosivity of eruptions beneath ice is expected to generally increase as the ice thins (Fig. 8; Tuffen *et al.* 2007). Therefore, where ice is over 150 m in thickness and thinning of more than 100 m occurs, the probability of more hazardous explosive eruptions will increase. This will be most relevant to volcanoes with deep ice-covered calderas such as Sollipulli, Chile (Gilbert *et al.* 1996). However, it is not currently possible to quantify the increased probability of explosive eruptions and whether it is significant.

There is stronger evidence that current ice recession may considerably increase hazards related to edifice instability. Capra (2008) has proposed that that the incidence of major volcano collapses is strongly affected by ice recession during deglaciation. Huggel *et al.* (2008) have noted an upturn in the rate of large-volume avalanches, which corresponds with and is attributed to recent climate change. Similar predictions are made for mountain instabilities due to recession of alpine glaciers (Keiler et al., this volume). Melting and unloading of ice may have a much more rapid effect on edifice stability than on melt production and eruption. Modelling by Huggel (2009) shows that the thermal perturbations that may destabilize slopes are likely to occur over tens or hundreds of years (for conductive heat flow processes) and years to decades (for advective/convective heat flow processes). Perturbations triggered by volcanic activity may be effective over much shorter time scales.

Andean stratovolcanoes that host rapidly-diminishing tropical glaciers are likely to be particularly sensitive to climate warming. Many glaciers are completely out of equilibrium with current climate and may completely disappear within decades (Vuille *et al.* 2008). Model projections of future climate change in the tropical Andes indicate a continued warming of the tropical troposphere throughout the 21st century, with a temperature increase that is enhanced at higher elevations. By the end of the 21st century, following the SRES A2 emission scenario, the tropical Andes may experience a massive warming on the order of 4.5–5 °C (Vuille *et al.* 2008). This warming will drive edifice instability both by removing ice, increasing the amount of meltwater at high elevations on edifices and thawing ice-bedrock contacts, encouraging slippage.

Climate warming may in some incidences reduce lahar hazards, as the disappearance of small volumes of snow and ice from volcanoes such as Popocatepetl will reduce the volume of ice available for meltwater flood generation. Dwindling areas of ice and snow will also reduce the probability of lahar generation. This reduction in lahar hazards may only be notable in volcanoes undergoing almost complete glacier extinction (Huggel *et al.* 2007*a*).

7. Gaps in our knowledge and targets for future research

Important gaps in our knowledge of links between melting of ice and volcanic hazards remain, which include:

1) Uncertainty about the timescale of volcanic responses to ice unloading. We currently have only limited insight into the reasons for delayed volcanic responses (Maclennan *et al.* 2002) and the timescales involved (Jellinek *et al.* 2004); response times are likely to differ in different tectonic settings.

2) Poor constraint on how ice bodies on volcanoes will respond to 21st century climate change. The highly localised effects of topography, microclimates and local geothermal and eruption-related processes on volcanoes conspire to create considerable diversity in the response of individual glaciers and ice sheets to climate change (e.g. Geirsdóttir *et al*. 2006, Rivera *et al.* 2006, Bown and Rivera 2007, Brock *et al.* 2007).

3) The sensitivity of volcanoes to small changes in ice thickness or to recession of small glaciers on their flanks is unknown. Although there is strong evidence that wholesale ice removal during deglaciation can significant accelerate volcanic activity there is considerable uncertainty about how volcanic responses to unloading scale with the magnitude, rate and distribution of ice unloading. A simple linear relationship between the rates of ice melting and additional melt production is unlikely to be appropriate. The effects of recession of different scales of ice body need to be considered, from the largest ice sheets to the smallest summit glaciers.

4) Lack of data on how past changes in ice thickness have affected the style of volcanic eruptions and associated hazards. Most statistical studies of the effects of ice thickness changes on volcanism have focussed exclusively on the frequency of eruptions. It would be of great interest to know whether the sizes of eruptions or the probability of large caldera-forming events increase during periods of ice recession.

5) It is not known how localised ice withdrawal from stratovolcanoes will affect shallow crustal magma storage and eruption. Existing models for how loading by ice affects volcanism have focussed on large (>50 km diameter), near-horizontal ice sheets and mantle melting (e.g. Jull and McKenzie 1996, Pagli and Sigmundsson 2008). Stratovolcanoes, which constitute the vast majority of ice- and snow-covered volcanoes worldwide, are entirely different systems, being characterised by smaller, thinner ice bodies and the existence of crustal magma chambers.

6) Broader feedbacks between volcanism and climate change remain poorly understood. A number of potential positive feedbacks during volcano-ice interactions exist, which could potentially greatly magnify the rate of ice recession and effects on volcanic activity. Feedbacks include the increased CO_2 emissions from accelerated volcanism during ice unloading, which may act to further warm the climate (Huybers and Langmuir 2009). Enhanced basal melting may destabilise ice sheets, leading to more rapid ice recession (Bell 2008). More locally, tephra covering the ice surface may affect the mass balance of glaciers (Rivera *et al.* 2006, Brock *et al*. 2007). Currently little is known about the effects of these feedbacks and whether they will play an important role in the 21st century and beyond.

Future work required

In order to resolve these problems both new data and improved models are required. Existing databases of known volcanic eruptions need to be augmented by numerous detailed case studies of the Quaternary eruptive history of ice-covered volcanoes, especially in the Andes, to determine whether the frequency and size of their eruptions has been influenced by past changes in ice thickness. The volcanic response should be examined to both large-magnitude, long timescale climatic changes such as glacial-interglacial cycles and to smaller, briefer fluctuations in the last millennium such as the Little Ice Age. This will reveal the sensitivity and response time of volcanic systems to a range of forcing timescales and magnitudes.

The unique record of palaeo-ice thicknesses provided by subglacially erupted volcanic deposits (e.g. Mee *et al.* 2006, Licciardi *et al*. 2007, Smellie *et al.* 2008, Tuffen *et al.* 2010) must be exploited in order to precisely reconstruct fluctuating local ice thicknesses on volcanic edifices. In tandem high resolution dating techniques will be required, which stretch the limits of existing radiometric methods. Geochemical indicators of the residence time of magma in shallow magma chambers could reveal whether shallow magma storage is affected by ice thickness variations.

Improved physical models are required to test how magma generation, storage and eruption at stratovolcanoes is affected by stress perturbations related to the waxing and waning of small-volume ice bodies on what is commonly steep topography. Finally, feedbacks between the mass balance of ice sheets and glaciers and volcanic activity need to be incorporated into future Earth System Models.

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Figure captions

Figure 1. **(a)** Explosive phreatomagmatic activity at Grímsvötn, Iceland on 2nd November 2004. Photograph by Matthew Roberts, Icelandic Meteorological Office. **(b)** A small plume of ash and steam at the ice-covered summit of Mt Redoubt, Alaska in March 2009 (photograph by Alaska Volcano Observatory). **(c)** Lahar and flood deposits in the Drift River Valley following eruptions at Mt Redoubt in 2009. Photograph by Game McGimsey, AVO/USGS. **(d)** Aerial view of lahar deposits that destroyed the town of Armero in 1985 after the eruption of Nevado del Ruiz, Columbia. Photograph by R.J. Janda, USGS.

Figure 2. **(a)** Iliamna Volcano, Alaska, showing the path of an ice–rock avalanche that originated from a geothermally active zone high in the summit region. From Huggel 2009, photograph by R. Wessels. **(b)** Map of Myrdalsjökull ice cap, Iceland, showing potential drainage directions of jökulhlaups triggered by eruptions at the ice-covered Katla volcano (from Eliasson *et al.* 2006).

Figure 3. Dramatic loss of snow and ice from the summit of Kilimanjaro between 2000 and 2007, from Thompson *et al.* (2009).

Figure 4. **(a)** Disruption of ice at the site of the 1998 Grímsvötn eruption, Vatnajökull, Iceland. Photograph by Magnus Tumi Guðmundsson. **(b)** Tephra-covered blocks of ice were a last remnant of a now-extinct glacier on Popocatepetl in 2004. From Julio-Miranda *et al.* (2008).

Figure 5. Modelled acceleration in melting of the Icelandic mantle during the last deglaciation (from Maclennan *et al.* 2002). **(a)** Increased rate of melting vs depth in the mantle. The melting rate is the volume of melt produced from each unit volume of mantle per kyr. **(b)** Modelled rate of melt production with a “spike” between 12 and 11 ka.

Figure 6. a) The ratio of postglacial (18-7 ka) to glacial (40-20 ka) activity at volcanoes worldwide plotted against a proxy for the amount of ice unloading from ice mass balance models (Huybers and Langmuir 2009). Regions with a less negative ice volume balance are those that are most likely to have been glaciated, and thus have experienced significant unloading of ice during the last deglaciation. It is at these regions that the strongest acceleration in the rate of eruptions has occurred, suggesting a causal link between unloading of ice and enhanced volcanic activity. **b)** Data from Jellinek et al. (2004) showing the SPECMAP $\delta^{18}O$ curve (a proxy for global ice volume) and the time series of eruptions in the Long Valley and Owens Valley volcanic fields, California. This data is used to show statistically significant correlation between ice unloading and accelerated volcanism.

Figure 7. Some approximate rates and amounts of ice thinning since the Little Ice Age and during deglaciation, together with projections for the 21st century using current rates of ice melting. Note that total thinning may in many cases be limited by the complete extinction of ice (e.g. Popocatepetl, Julio-Miranda *et al.* 2008).

Figure 8. Results of modelling of rhyolitic eruptions under ice from a 1.5 km-long fissure. The evolution of subglacial cavities during melting and ice deformation is simulated and the combination of ice thickness and magma discharge rate likely to lead to explosive and intrusive eruptions is

indicated. Explosive eruptions (above the line) are favoured by thin ice and high magma discharge rates. They are more hazardous than intrusion eruptions since meltwater is produced far more quickly (Guðmundsson 2003) and eruptions may pierce the ice surface, producing tephra hazards. Modified from Tuffen *et al.* (2007).

Figure 1.

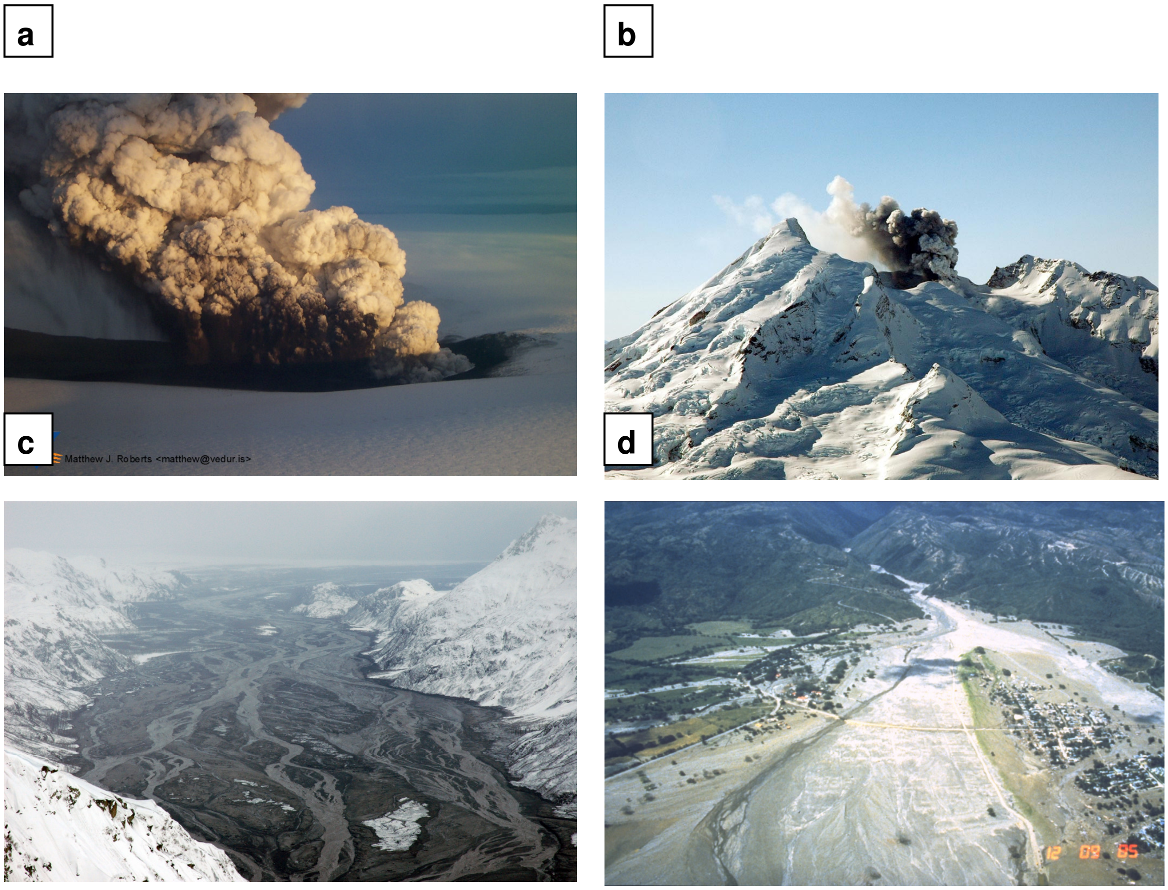


Figure 2.

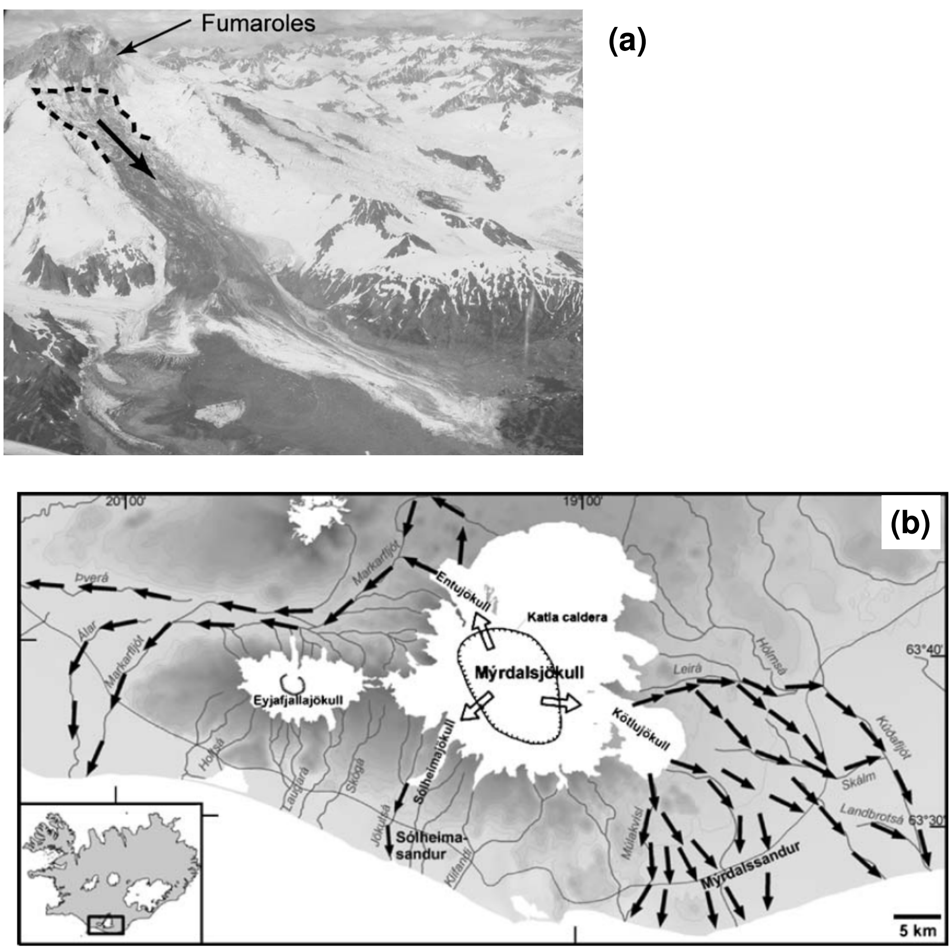

Fumaroles
(a)
(b)
20°00'
19°00'
63°40'
63°30'
Þverá
Markarfljót
Álar
Entujökull
Katla caldera
Mýrdalsjökull
Eyjafjallajökull
Kötlujökull
Leirá
Hólmsá
Kúðafljót
Skálm
Landbrotsá
Holtsá
Laugará
Skógá
Sólheimajökull
Jökulsá
Sólheima-
sandur
Klifandi
Múlakvísl
Mýrdalssandur
5 km

Figure 3.

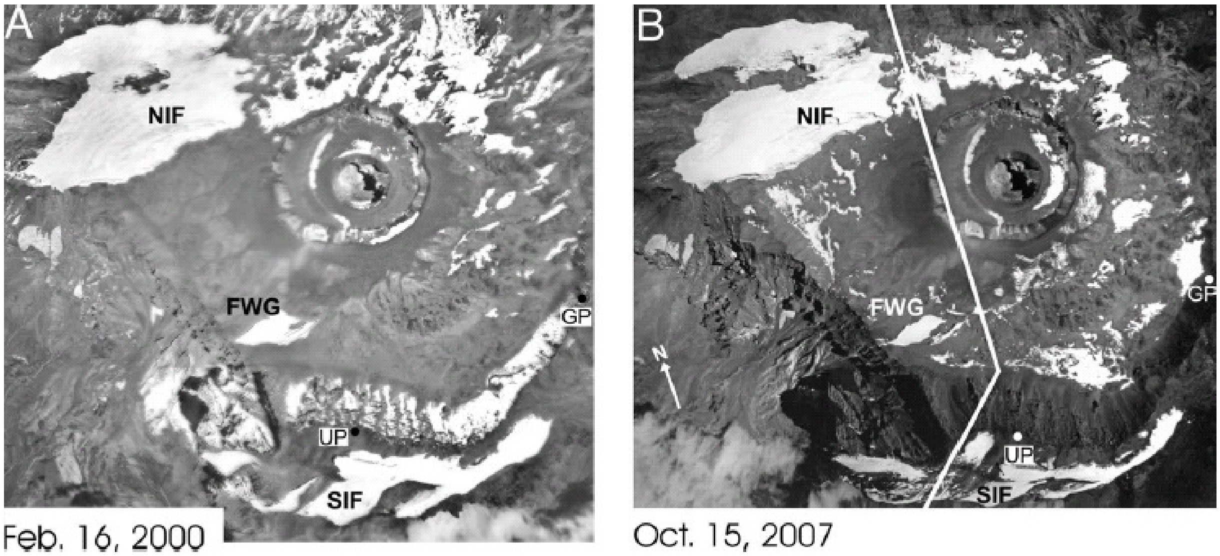


Figure 4.

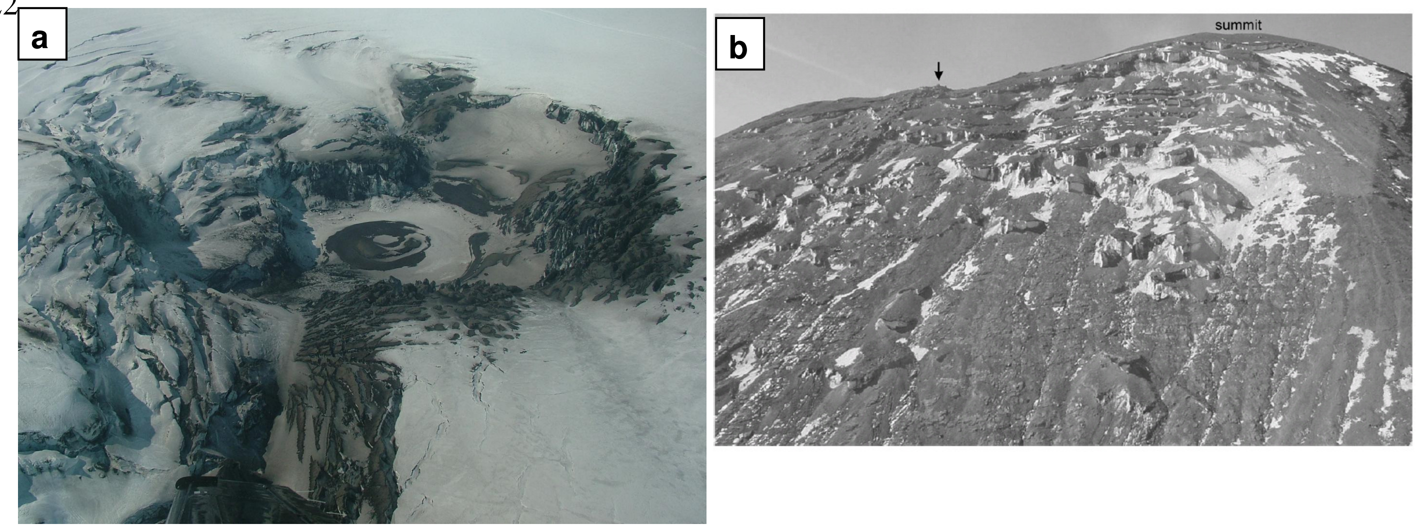
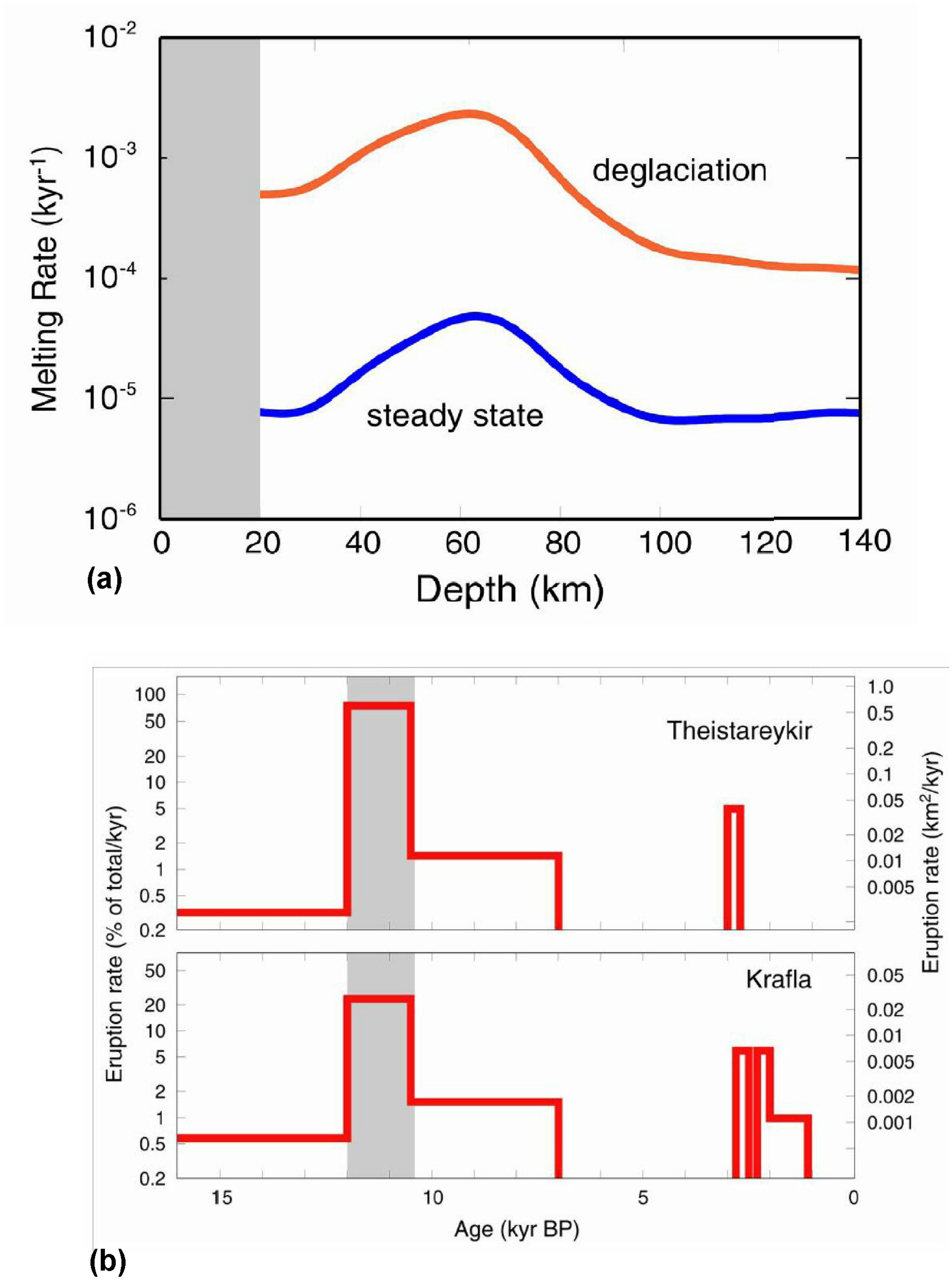
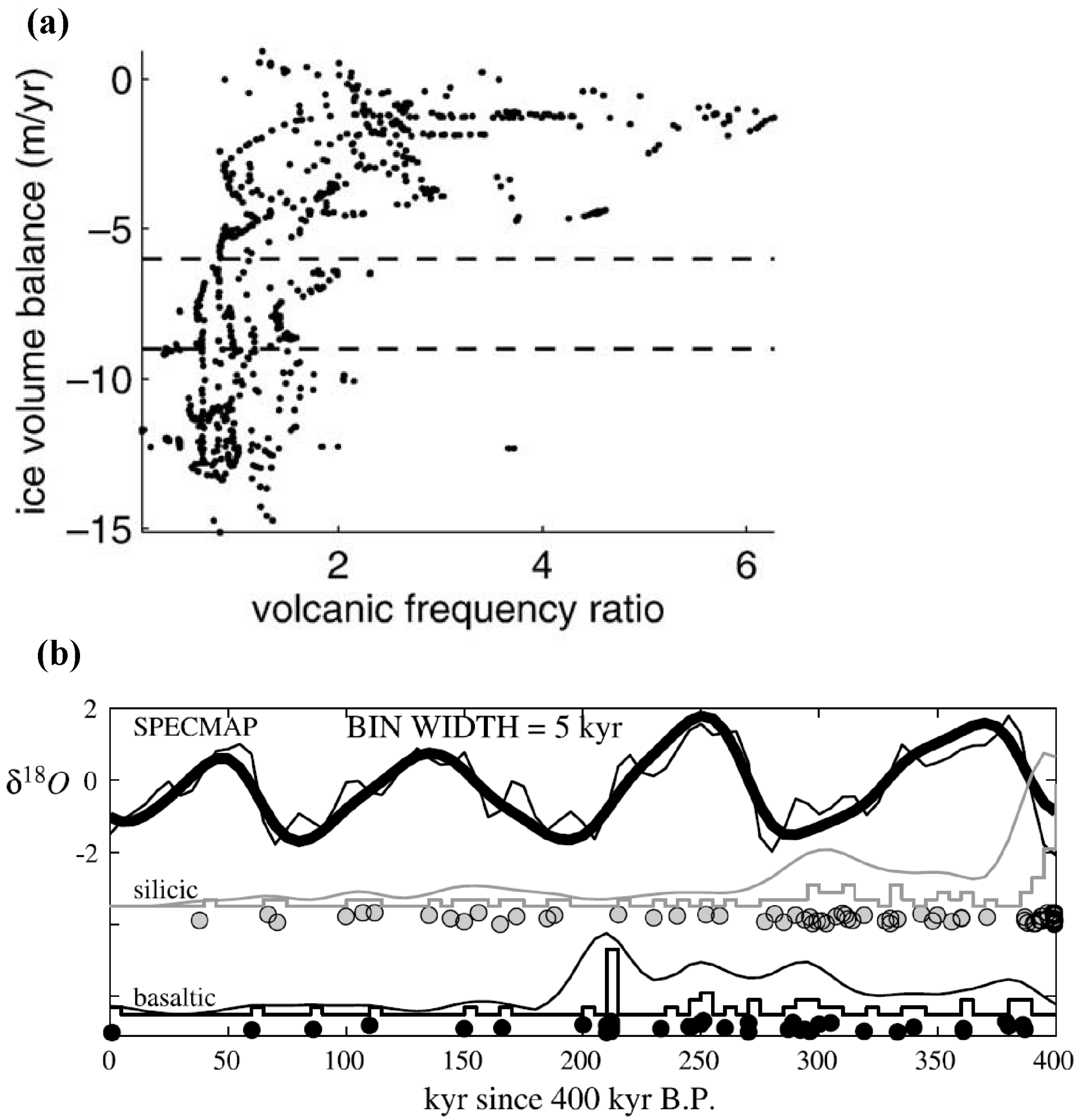


Figure 5.



Melting Rate (kyr^{-1})
10^{-2}
10^{-3}
10^{-4}
10^{-5}
10^{-6}
deglaciation
steady state
0
20
40
60
80
100
120
140
(a)
Depth (km)
Theistareykir
Krafla
100
50
20
10
5
2
1
0.5
0.2
1.0
0.5
0.2
0.1
0.05
0.02
0.01
0.005
0.002
0.001
Eruption rate (% of total/kyr)
Eruption rate (km^2/kyr)
15
10
5
0
Age (kyr BP)
(b)

Figure 6



(a)
ice volume balance (m/yr)
0
−5
−10
−15
2
4
6
volcanic frequency ratio
(b)
SPECMAP
BIN WIDTH = 5 kyr
$\delta^{18}O$
2
0
-2
silicic
basaltic
0
50
100
150
200
250
300
350
400
kyr since 400 kyr B.P.

Figure 7.

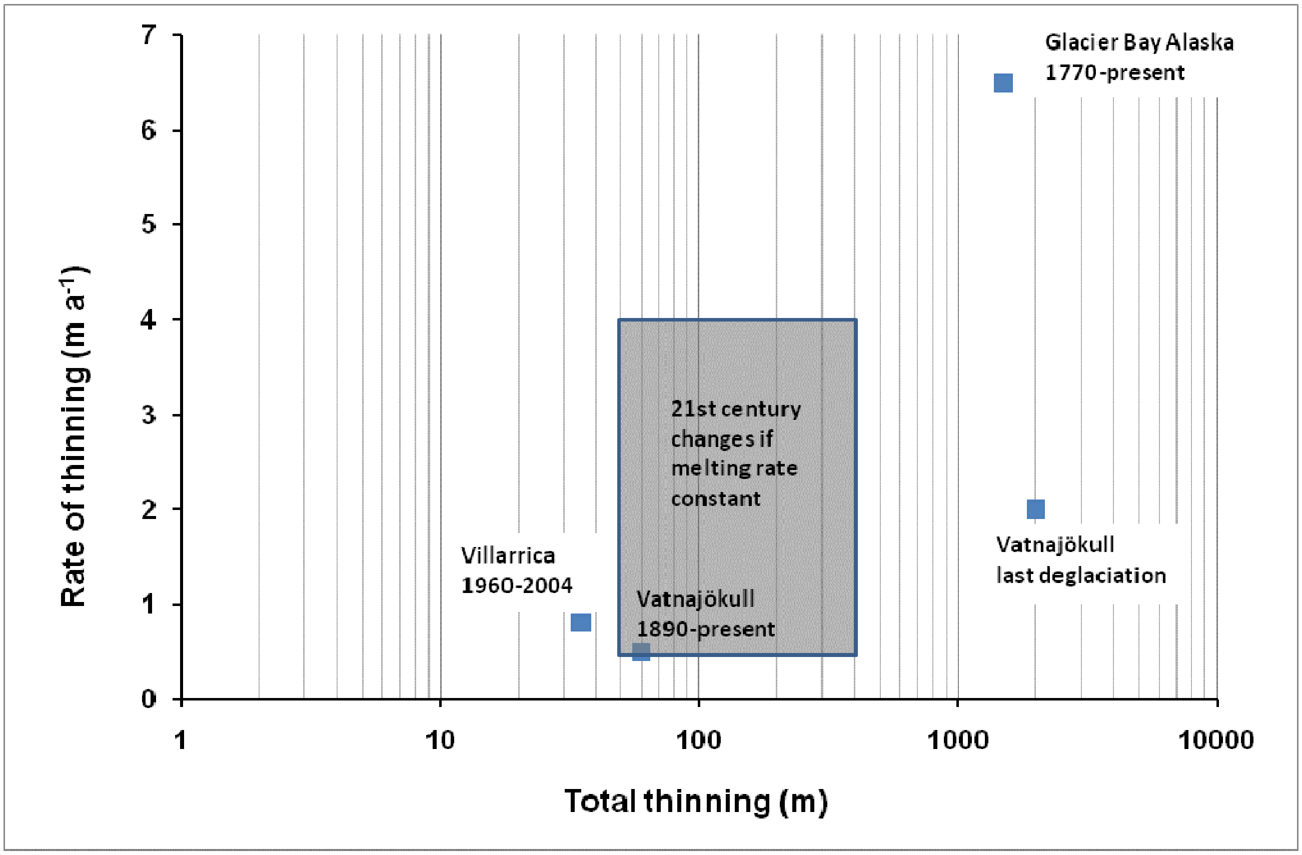
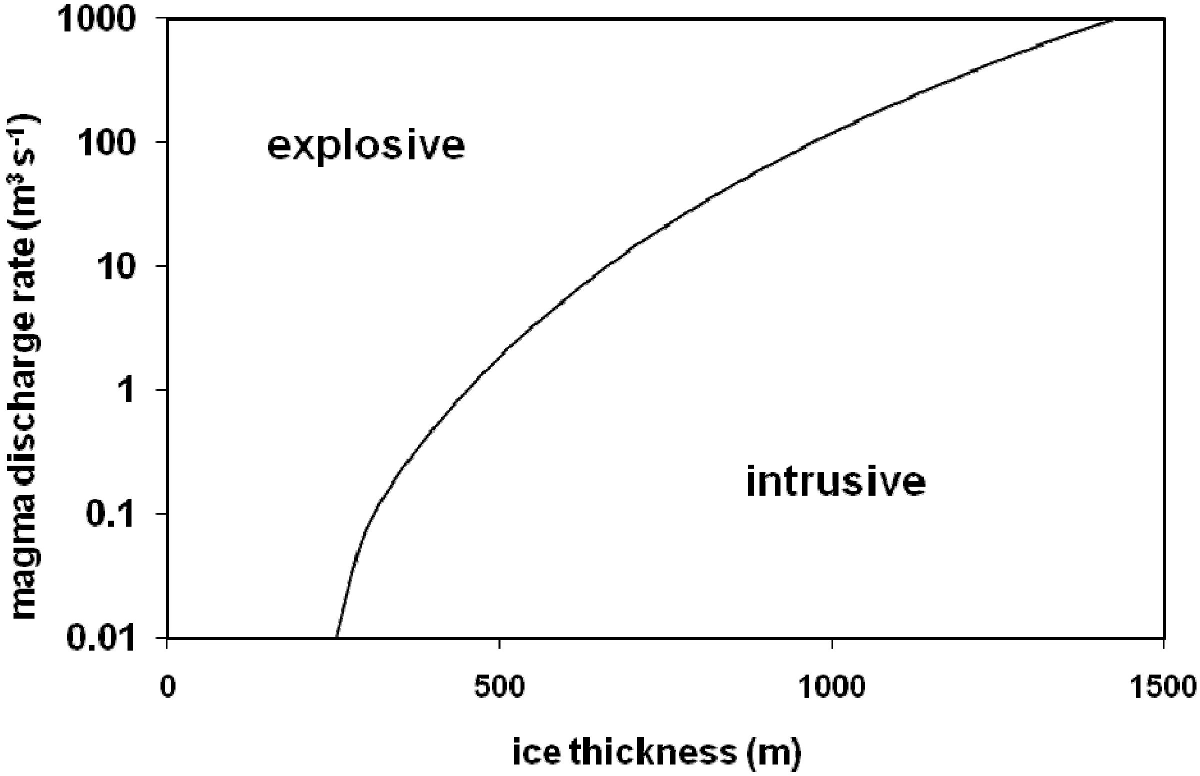

Glacier Bay Alaska
1770-present
21st century
changes if
melting rate
constant
Villarrica
1960-2004
Vatnajökull
1890-present
Vatnajökull
last deglaciation
Rate of thinning (m a-1)
Total thinning (m)
0
1
2
3
4
5
6
7
1
10
100
1000
10000

Figure 8.



explosive
intrusive
1000
100
10
1
0.1
0.01
0
500
1000
1500
magma discharge rate (m³ s⁻¹)
ice thickness (m)