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## RESEARCH ARTICLE

10.1002/2017JB014067

## Key Points:

- Clinopyroxene crystallization of Grímsvötn basalt indicates mid-to-lower crustal magma storage above the Iceland mantle plume
- Different eruption regimes are due to variable recharge of deeper-seated and gas-rich magma recharge
- Sulfur concentration of matrix glass and exsolution budget correlate with known magma volumes

## Supporting Information:

- Supporting Information S1
- Table S1
- Table S2
- Table S3

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# Magma storage beneath Grímsvötn volcano, Iceland, constrained by clinopyroxene-melt thermobarometry and volatiles in melt inclusions and groundmass glass

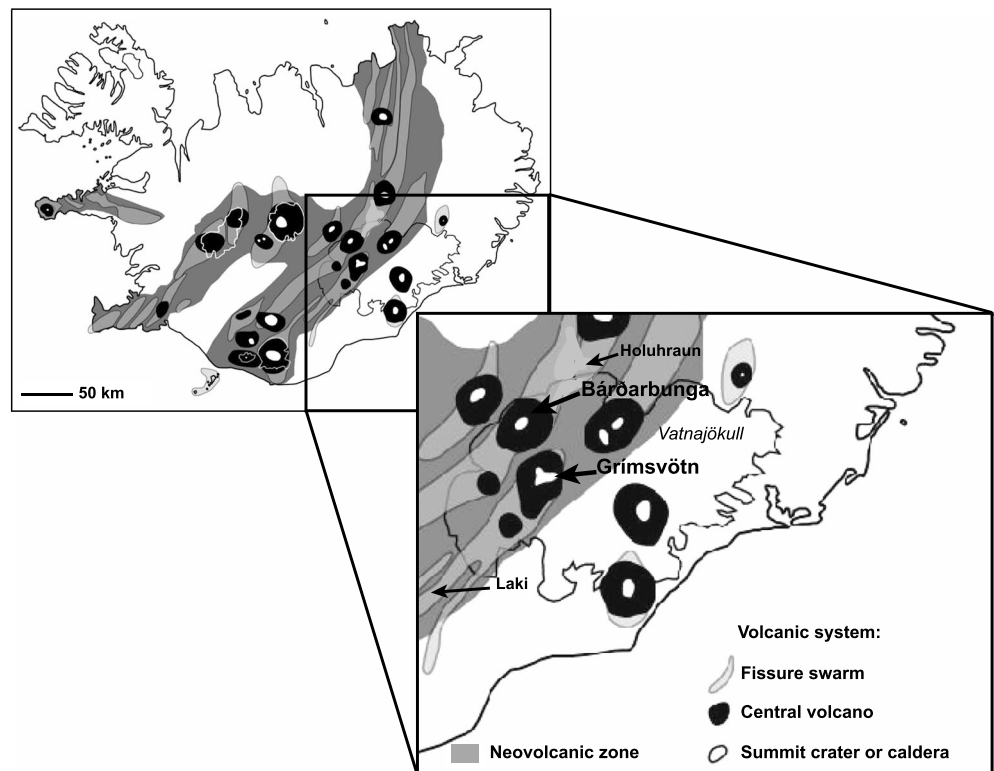
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**Abstract** Basalt eruptions at Grímsvötn volcano, Iceland, are generally of low intensity; however, occasionally, an order of magnitude larger eruptions occur. In order to discuss the reasons for this difference, the degassing budget of S and Cl and crystallization conditions of the eruptive magma were determined from volatile concentration measured in melt inclusion (MI) and groundmass glass and thermobarometry, respectively. Tephra of two of the largest historical eruptions (2011 and 1873) and two much smaller eruptions (2004 and 1823) were investigated. Sulfur and Cl concentrations are higher in groundmass glass of the smaller eruptions due to incomplete outgassing caused by melt quenching in contact with glacial water. Sulfur concentration and degassing budget correlate with erupted magma volumes. Higher volatile concentrations of MI from the larger eruptions reflect important recharge of gas-rich magma from depth. The recharge causes a high-magnitude eruption followed by increased eruption frequency over the following decades. Pressure and temperature estimates of crystallization are obtained through equilibrium clinopyroxene-glass pairs, where crystals adjacent to, and in textural equilibrium with, both groundmass glass and that of MI were measured. An average crystallization pressure of  $4 \pm 1$  kbar corresponding to approximately  $15 \pm 5$  km depth was obtained together with a considerable temperature range, 1065–1175°C. Similar crystallization depths are obtained for the basalt of the 2014–2015 Bárðarbunga rifting event and to a low resistivity layer revealed by magnetotelluric surveys. Therefore, an important magma storage depth is inferred at lower crustal depth above the center of the Iceland mantle plume.

## 1. Introduction

Knowing the depth at which magma is stored beneath active volcanoes is important for eruption mitigation and corresponding hazard assessment. The configuration of magma plumbing systems beneath active basaltic volcanoes is principally inferred from geophysical observations. The most densely equipped volcanoes, such as Kilauea (Hawaii), Etna (Italy), and Piton de la Fournaise (Réunion Island), appear to be characterized by a plexus of dikes and sills or “vertically stacked magma reservoirs” [Poland *et al.*, 2014, and references therein]. At Kilauea, basalt of different nature allows identifying the relative depth of magma storage, with evolved composition reflecting stagnant or residual magma stored at high-level and more mafic basalt deriving from further depth [e.g., Garcia *et al.*, 2000]. Time series of variable isotope and trace element ratios suggest decadal-long periods of magma supply from a single shallow magma body, which alternates with those of two laterally distinct bodies [Pietruszka *et al.*, 2015]. Magma recharge from depth can be identified at Kilauea by the eruption of high-MgO basalts or CO<sub>2</sub>-rich olivine-hosted melt inclusions [e.g., Anderson and Brown, 1993].

Grímsvötn volcanic system, Iceland, being mostly subglacial is scarcely equipped with geophysical instruments, but nevertheless, a high-level magma chamber has been inferred from combined magnetic, gravity, seismic, and deformation studies [e.g., Gudmundsson and Milsom, 1997; Alfaro *et al.*, 2007; Hreinsdóttir *et al.*, 2014]. A low-velocity body with a volume of 10–20 km<sup>3</sup> was inferred at approximately 3–4 km depth by Alfaro *et al.* [2007], beneath which a dense mass of higher velocity is observed. Measured deformation before and during the 2011 eruption was interpreted by a Mogi model to give a point pressure source depth of  $1.7 \pm 0.2$  km [Hreinsdóttir *et al.*, 2014], whereas the deformation pattern of Grímsvötn central volcano over a longer time period suggests at least two magma reservoirs [Reverso *et al.*, 2014]. Only basalt of relatively homogeneous and evolved tholeiitic composition is produced as tephra from the subglacial central



**Figure 1.** Map of South East Iceland showing Grímsvötn and Bárðarbunga volcanic systems partially located under the Vatnajökull ice cap. The volcanic systems are composed of a central volcano, hosting a caldera, and fissure swarms, at which occurred the basaltic fissure eruptions of Laki 1783–1784 and Holuhraun 2014–2015, respectively. Modified after Jóhannesson and Sæmundsson [1998].

volcano [Óladóttir *et al.*, 2011a, 2011b] and lavas from the subaerial fissure swarm of similarly uniform composition, such as that of the Laki eruption [e.g., Sigurdsson and Sparks, 1978; Sigmarsson *et al.*, 1991]. In order to better constrain the magma system beneath Grímsvötn, depth of crystallization at which magma resided is estimated through thermobarometry and phase equilibria. The results are compared with those obtained on the recent Holuhraun basalt, erupted in 2014–2015 on the adjacent Bárðarbunga volcanic system. Then, the petrological depth estimate of crystallization at these two volcanic systems is compared with results from large-scale magnetotelluric survey in southern Iceland [Eysteinsson and Hermance, 1985] and also with the inferred seismic layering of the Icelandic crust [e.g., Bjarnason, 2008]. In addition, sulfur and chlorine concentrations were measured in melt inclusions and groundmass glass from four historical eruptions, two larger events, and two smaller eruptions with an order of magnitude lower mass eruption rate. Their concentrations and degassing budget suggest a link between eruption regime, degassing efficiency, and volume of magma produced. The combined results permit evaluation of volatile-rich magma recharge in controlling different eruption regimes at Grímsvötn volcano.

## 2. Geological Setting

Grímsvötn and Bárðarbunga volcanic systems are located above the center of the Iceland mantle plume (Figure 1). For the former, most eruptions occur inside the subglacial Grímsvötn caldera, with a few notable exceptions such as the 1783–84 Laki and 1996 Gjalp events [e.g., Thordarson and Self, 1993; Sigmarsson *et al.*, 2000]. Grímsvötn central volcano is the most active Icelandic volcano, erupting on average once per decade over the last eight centuries [Larsen *et al.*, 1998]. The so-called Saksunarvatn ash, an important tephrochronological horizon from the very early Holocene, most likely originated at Grímsvötn [e.g., Neave *et al.*, 2015]. Eruptions within the ice-covered part of the volcanic system are (i) predominantly phreatomagmatic in nature or (ii) alternating between phreatomagmatic activity and magmatic phases of subplinian/plinian intensity, termed here as plinian for simplification. Eruptions on the ice-free part are

predominantly effusive and able to produce flood basalts, such as the Laki lava fields. The order of magnitude smaller Holuhraun 2014–2015 eruption on the Bárðarbunga volcanic system produced more primitive basalt than Grímsvötn or olivine normative tholeiite [e.g., *Gudmundsson et al.*, 2016]. Two of the four Grímsvötn eruptions studied here (1873 and 2011) are among the largest during the historical period (last thousand years approximately), whereas the remaining two (1823 and 2004) are typical of the more frequent low-intensity events (see further details in the supporting information [*Petersen et al.*, 2012; *Thjóðólfur*, 1873; *Thorarinsson*, 1974; *Thoroddsen*, 1924]).

### 3. Sample Description and Methods

#### 3.1. Samples

The G1823 and G1873 tephra were collected in August 2005, approximately 25 km south of the Grímsvötn caldera, from tephra layers recently exposed in the ablation region of one of the outwash glaciers of Vatnajökull ice-cap (see sampling methods in *Larsen et al.* [1998]). Two decimetric bombs (G-2004-T2 and G-2004-T4) were selected from the 2004 eruption products collected from the crater inside the caldera wall. The near-vent lapilli-size tephra from the first eruption day in 2011 (sample G2011-D) and fine-grained tephra collected on the lowland, south of Vatnajökull (Gv4), represent the 2011 tephra. For comparison, results from a tephra produced early in the 2014–2015 Bárðarbunga rifting event were included in this study.

#### 3.2. Melt Inclusion Preparation

Olivine, clinopyroxene, and plagioclase (hereafter abbreviated ol, cpx, and plag) crystals were handpicked under a binocular microscope from the 100–250 and 250–600  $\mu\text{m}$  grain size fractions of crushed tephra. To facilitate the handpicking, dense liquids (bromoforn and diiodomethane) were used to separate crystals from glass and plag from ferromagnesian minerals. Crystals with melt inclusions (MI) were washed with acetone, embedded in epoxy, and polished individually in order to generate adequate exposure of the MI for electron probe microanalysis (EPMA). The MI are spherical to oblate in shape and range in size from 5 to 190  $\mu\text{m}$  in diameter. Most MI contain shrinkage bubbles but lack minerals and are primary melt inclusions [*Roedder*, 1984]. In total, 103 crystals containing 179 MI were selected and prepared for analysis. The groundmass of the 2011 tephra is composed of glass patches with variable microlite contents ranging from those that are completely free of microlites to heavily crystallized groundmass. From the G2011 tephra, a total of 49 cpx hosted 76 MI, and one olivine-hosted MI were obtained in addition to those of an earlier study [*Sigmarsson et al.*, 2013] (Table S1 in the supporting information). From the G2004 tephra, 12 cpx containing 33 MIs and six ol with seven MI were collected. In that of G1873, 10 cpx with 23 MI and two MI in two ol were analyzed and four cpx hosting nine MI from the G1823 tephra. *Sigmarsson et al.* [2013] showed that most MI in plag from the 2011 eruption had composition modified by host crystallization and O isotope ratios measured in plag from Grímsvötn products [*Bindeman et al.*, 2006] are generally out of equilibrium with the groundmass glass. Therefore, plag was not studied in same details here as the ferromagnesian minerals. The groundmass glass around euhedral crystals was analyzed both adjacent to the crystal rims and elsewhere in each glass patch.

#### 3.3. Microprobe Data Acquisition

Major and volatile (S, Cl, and F) element concentrations were measured on Cameca SX-100 at Laboratoire Magmas et Volcans in Clermont-Ferrand, France. A 15 keV accelerating voltage and a 15 nA current were used for the mineral analyses. The larger MI and groundmass glass were analyzed with a spot diameter of 20  $\mu\text{m}$  and a current of 8 nA, whereas the smallest MI were analyzed with a 5  $\mu\text{m}$  beam and a current as low as 4 nA. The secondary glass standard A99 was analyzed with a 5  $\mu\text{m}$  beam. All results are given in Table S1. Reproducibility and accuracy were established by replicate analyses of VG-A99 glass standard for major element concentrations, Alvin glass standard for S concentrations, and Ke12 glass standard for Cl and F concentrations. Measured values of  $0.100 \pm 0.003$  wt % for S in Alvin,  $0.33 \pm 0.01$  wt % for Cl, and  $0.46 \pm 0.04$  wt % for F in Ke12 are in good agreement with the published accepted values of these standards [*Óladóttir et al.*, 2011a; *Moune et al.*, 2007, and references therein]. For the concentration levels of Grímsvötn glass, the 1 SD ranges from 2 to 10% and 10 to 15%, respectively, on S and Cl concentrations. The detection limit for F concentration is between 200 and 300 ppm. The  $K\alpha$  ray position of sulfur was used to estimate  $\text{fO}_2$  following *Carroll and Rutherford* [1988].

### 3.4. Test of Mineral-Melt Equilibrium and Geothermobarometry

Phase equilibrium is a prerequisite for application of thermobarometers based on mineral-liquid compositions and thermodynamic equilibrium. Three approaches are used here to assess the phase equilibrium, namely, textural criteria, calculated normalized mineral components, and equilibrium partition coefficients. Back-scattered electron (BSE) images revealed a few anhedral to subhedral crystals that were rejected from the  $P$ - $T$  calculations. Most crystals selected for analysis are euhedral, suggesting textural equilibrium with their surrounding liquid (Table S1 and Figure S1 in the supporting information). In the case of zoned crystals, only rims and neighboring glass were used as representative of the last state of equilibrium between a given crystal and melt. The special case of sector-zoned cpx is discussed further below. The state of equilibrium was assessed from the ideal cpx end-member components (DiHd, EnFs, etc. [Putirka, 1999]; Figure S2). Final test of equilibrium was the criterion of Fe-Mg partitioning between ferromagnesian minerals and liquids:  $K_D^{\text{min-liq}}(\text{Fe-Mg}) = [\text{MgO}_{\text{liq}} * \text{FeO}_{\text{min}}] / [\text{MgO}_{\text{min}} * \text{FeO}_{\text{liq}}]$ , where MgO and FeO are molar fractions. The equilibrium  $K_D$  value of  $0.28 \pm 0.08$  between cpx and melt is used and  $0.30 \pm 0.03$  for the ol-liquid equilibrium [e.g., Roeder and Emslie, 1970; Herzberg and O'Hara, 1998; Toplis, 2005; Putirka, 2008]. Clinopyroxene melt was used for  $P$ - $T$  estimates because plag-liquid equilibrium constant  $K_D^{\text{fsp-liq}}(\text{An-Ab})$  is not well constrained for tholeiitic magma, and the olivine crystals are rare.

Equilibrium crystallization temperatures ( $T$ ) between ol, plag, cpx, and matrix glass were calculated respectively with equations 22, 24a, 33, and 13 of Putirka [2008, and references therein], with associated standard error of estimate (SEE) being 29, 36, 42, and 71°C, respectively. The crystallization pressure ( $P$ ) estimates are based on clinopyroxene-liquid (cpx-liq) equilibrium using Putirka's equation 30 with a SEE of 1.6 kbar. The results are compared with  $P$  and  $T$  at which the phases olivine, plagioclase, augite, and melt are in equilibrium (OPAM; see the supporting information) [Yang *et al.*, 1996]. The OPAM phase equilibria was already applied to estimate crystallization pressure for several Icelandic basalts [Kelley and Barton, 2008], whereas Budd *et al.* [2016] and Geiger *et al.* [2016] applied the Putirka [2008] cpx-liq barometer by selecting what they term as "appropriate nominal melt composition." In the absence of measurable liquid composition adjacent to the cpx due to highly palagonized glasses and/or crystallized groundmass in the studies of Budd *et al.* [2016] and Geiger *et al.* [2016], the authors suggest adoption of a "plausible whole-rock composition." However, Óladóttir *et al.* [2008, 2011b] demonstrated significant variability in glass composition of Holocene tephra from subglacial basalt volcanoes in Iceland. Thus, identifying a single "appropriate nominal melt composition" remains highly subjective and can lead to erroneous results. Comparing the empirically calculated cpx components to those analyzed is not a very stringent tool in evaluating equilibrium condition and only eliminates obscure analyses (such as occasional mixed glass-crystal rim analyses identified by monitoring  $\text{K}_2\text{O}$  concentrations). The problem with these approaches is perhaps best illustrated by the case of sector-zoned cpx (see further discussion in the supporting information [Hammer *et al.*, 2016]). Two examples are given in Figure S3 illustrating that only analyses of euhedral mineral rims and the adjacent glass yield reasonable  $P$ - $T$  results, whereas other cpx analyses yield erroneous results despite  $K_D^{\text{min-liq}}(\text{Fe-Mg})$  (and cpx components) within accepted equilibrium range.

The approach preferred here was to calculate  $P$  and  $T$  only from analyses of crystal facets of the cpx phenocrysts in textural equilibrium with the adjacent glass, verified by BSE images, and the crystal cores enveloping glass inclusions. This ensures that only the relevant liquid and cpx compositions are utilized for the estimation of cpx crystallization pressure and temperature. Depth of cpx crystallization before two of the last three eruptions in Iceland, Eyjafjallajökull 2010 and Bárðarbunga 2014–2015, was estimated with this approach by Keiding and Sigmarsson [2012] and Bali *et al.* [2015]. In all three cases a strong agreement was found with depth estimates based on seismicity, minimum pressure derived from  $\text{H}_2\text{O}$ - $\text{CO}_2$  solubility, and GPS and interferometric synthetic aperture radar deformation data [e.g., Tarasewicz *et al.*, 2012; Gudmundsson *et al.*, 2016]. Recently, Neave and Putirka [2017] proposed a different calibration for calculating pressure of cpx crystallization with increased emphasis on results from 1 atm melting experiments. As discussed below, the application of their new calibration on two products from Grímsvötn volcanic system, the Laki basalt and the Saksunarvatn's tephra, yields results within error of those presented here on the four historical Grímsvötn tephra. However, their preferred calibration yields much lower  $P$  for experimental products at 5 and 10 kbar [Haddadi, 2016], whereas that of Putirka [2008] match the experimental results within error of that calibration. In addition,  $P$  estimates based on the OPAM phase equilibria are within error of the calibration of Putirka

[2008] that is therefore preferred in this paper. Nevertheless, it should be emphasized that only further experiments at  $P$  ranging from 1 atm to 5 kbar will permit refinement of the cpx-glass barometer.

### 3.5. Petrological Estimation of S Degassing

The petrological method for inferring S degassing is based on the difference ( $\Delta S$ ) between sulfur concentrations measured in phenocryst-hosted melt inclusions (MI), taken to indicate the preeruptive sulfur concentration dissolved in the parental melt, and the degassed groundmass glass representing the residual sulfur concentrations not emitted to the atmosphere [Devine *et al.*, 1998]. Scaled to the mass of magma erupted  $\Delta S$  yields the exsolved mass of sulfur, which can both form a gas phase ( $H_2S$  and  $SO_2$ ) and precipitate as sulfide [Sigmarsson *et al.*, 2013]. The petrological method estimates the mass of sulfur exsolved from the initial magma as follows:  $\text{mass } SO_2 = \alpha \rho V_{DRE} (S_{MI, \max} - S_{\text{Groundmass glass, mean}})$ , where  $\alpha = MW(SO_2)/MW(S) = 64.06/32.06$ ,  $\rho = 2750 \text{ kg m}^{-3}$  for Grímsvötn magmas and the dense rock equivalent volume ( $V_{DRE}$ ) from the eruption of interest.

## 4. Results

### 4.1. Petrography, Mineral, and Glass Compositions

The four Grímsvötn tephra are of quartz normative tholeiitic composition as are other Holocene Grímsvötn tephra [e.g., Grönvold and Jóhannesson, 1984; Óladóttir *et al.*, 2011a, 2011b]. All four tephra have average MgO concentrations close to 5 wt %, in both the groundmass glass and the MI. Largest variability is observed for the 2011 tephra with a range from approximately 3 to 7 wt % with a bimodal distribution. Groundmass glass analyzed adjacent to euhedral crystals is of identical composition to that analyzed further away from the crystal rim. The tephra are composed of ~95% glass with only ~5% of visible crystals, namely, plag, cpx, and ol in decreasing abundances. Sulfide globules and FeTi-oxides were only observed in the G2011 tephra [Sigmarsson *et al.*, 2013]. Representative mineral compositions are listed in Table S2. Plagioclase of the Grímsvötn 2011 tephra, which has the largest mineral compositional range of the four eruptions, varies from  $An_{55}$  to  $An_{76}$ , within the range found for the Saksunarvatn ash [Neave *et al.*, 2015]. Still larger compositional range is observed in Laki products with macrocrysts ranging from  $An_{53}$  to  $An_{89}$  [Neave *et al.*, 2013], extending the range ( $An_{75}$ – $An_{90}$ ) observed by Bindeman *et al.* [2006]. Macrocryst zonation is variable often with plag core, rich in Ca ( $An_{76}$ ), frequently mantled by thick rim of lower Ca composition ( $An_{73}$ ), whereas cpx occasionally shows complex sector zoning with a surrounding rim lower in Fe (Figure S3). These two principal mineral phases are both found as individual crystals but also forming glomerocrysts. Clinopyroxene is of augite composition with Mg # in the range of 73–84, within the range of those from the Saksunarvatn ash (Mg #: 71–87 [Neave *et al.*, 2015]) and those of the Laki eruption (Mg #: 50–83 [Neave *et al.*, 2013]). Olivines of the 2011 eruption have Fo values ranging from 68 to 77, which again is approximately within the range found in Saksunarvatn tephra (70–76 and 84–87 [Neave *et al.*, 2015]). Bindeman *et al.* [2006] observed a Fo range of 72–86 in Laki macrocrysts, while Neave *et al.* [2013] measured Fo as low as 60. Larger variations are thus observed in mineral compositions of the larger Grímsvötn eruptions. Results on S and Cl concentrations and cpx crystallization pressure and temperature will first be presented for the two larger eruptions (in 2011 and 1873) and subsequently for the two smaller eruptions (those of 2004 and 1823).

Water and carbon dioxide concentrations were measured in a single melt inclusion by the ion probe (IMS 1270 instrument at Nancy, France). The results are 0.66 wt %  $H_2O$  and 960 ppm  $CO_2$  (Table S2), which are comparable to 0.68 wt %  $H_2O$  measured in MI of olivines and plagioclases from the Saksunarvatn ash [Neave *et al.*, 2015].

### 4.2. Sulfur and Chlorine

Sulfur and Cl are of concentration high enough to be measured by EMP, whereas the F concentrations are below the EMP detection limit (~300 ppm) and will not be further discussed. In the MI, the highest S and Cl concentrations, which range from 1103 to 1974 ppm and 112 to 339 ppm, respectively, are observed in MI of the larger eruptions (G1873 and G2011; Table 1 and Figure 2). Lower maxima S and Cl concentrations were obtained in MI of minerals of smaller eruptions (G1823 and G2004). Groundmass glass volatile concentrations vary significantly with S ranging from 407 to 1758 ppm and Cl from 142 to 307 ppm. Significantly, the weighted mean S concentrations of the combined groundmass glass of the larger two eruptions is significantly lower (~880 ppm) than that of the combined groundmass glass for the smaller eruptions

**Table 1.** Petrological Method Applied to Sulfur and Chlorine<sup>a</sup>

|            | Sulfur |               |       |               |       |               |       |               | Chlorine |               |       |               |       |               |       |               |
|------------|--------|---------------|-------|---------------|-------|---------------|-------|---------------|----------|---------------|-------|---------------|-------|---------------|-------|---------------|
|            | G1823  |               | G1873 |               | G2004 |               | G2011 |               | G1823    |               | G1873 |               | G2004 |               | G2011 |               |
|            | ppm    | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ | ppm      | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ | ppm   | $\pm 1\sigma$ |
| Min(MI)    | 1386   |               | 1410  |               | 1220  |               | 1103  |               | 164      |               | 122   |               | 114   |               | 112   |               |
| Max(MI)    | 1731   | 65            | 1789  | 73            | 1667  | 75            | 1974  | 70            | 205      | 34            | 266   | 36            | 226   | 42            | 339   | 37            |
| Mean(MI)   | 1571   |               | 1594  |               | 1443  |               | 1578  |               | 183      |               | 175   |               | 161   |               | 180   |               |
| Min(GG)    | 407    |               | 513   |               | 671   |               | 465   |               | 166      |               | 142   |               | 173   |               | 153   |               |
| Max(GG)    | 1758   |               | 1047  |               | 1586  |               | 1749  |               | 182      |               | 209   |               | 307   |               | 252   |               |
| Mean(GG)   | 1050   | 56            | 766   | 47            | 1138  | 62            | 908   | 55            | 210      | 13            | 180   | 27            | 227   | 39            | 208   | 34            |
| $\Delta X$ | 681    | 121           | 1023  | 120           | 539   | 137           | 1066  | 125           | −5       | 47            | 86    | 63            | −1    | 81            | 131   | 71            |
| $D_X$ (%)  | 39     | 8             | 57    | 9             | 32    | 10            | 54    | 8             |          |               | 32    | 28            |       |               | 39    | 25            |

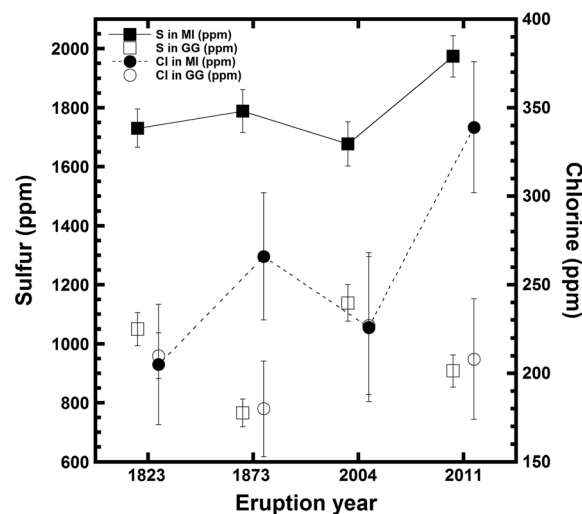
<sup>a</sup>It is based on the difference between the maximum content in the melt inclusions minus the groundmass glass (GG) mean volatile concentration of the same element (i.e.,  $X_{\text{max MI}} - X_{\text{mean GG}} = \Delta X$ ). Percentage of initial  $X$  available ( $D_X$ ), which exsolved from the melt is  $D_X = 100 * \Delta X / (X_{\text{max MI}})$ . Multiplying  $\Delta X$  by the dense rock equivalent volume of an eruption and by the mean density leads to the quantification of the mass emission of a given volatile ( $X$ ). All analyses are given in Table S1

(~1100 ppm). Consequently, the difference between the maximum S concentrations in the MI minus mean S concentrations ( $\Delta S$ ) of the groundmass glass is significantly larger for the two larger eruptions (~1000 ppm) compared to  $\Delta S$  of the smaller eruptions (~600 ppm). This observation is reinforced by higher  $\Delta \text{Cl}$  for the larger eruptions and negligible Cl degassing in tephra from the smaller two eruptions. Both chlorine and sulfur degassing thus appears hindered in the phreatomagmatic glass of the two smaller eruptions.

### 4.3. Pressure and Temperature of Partial Crystallization

#### 4.3.1. Grímsvötn 2011

All the  $P$ - $T$  calculations are performed assuming 0.6 wt %  $\text{H}_2\text{O}$ . Olivine-liquid temperature is estimated from a single MI-host-crystal pair giving  $1212 \pm 29^\circ\text{C}$ , whereas the groundmass glass composition gives  $T$  in the range of  $1100$ – $1149^\circ\text{C}$ . The cpx-liq thermometer for MI in cpx gives results between  $1065$  and  $1175^\circ\text{C}$  (Table 2), which display a near Gaussian distribution with a median centered on  $1135^\circ\text{C}$  (Figure 3). The lowest cpx-liquid temperature is close to the 1 atm experimental determined temperature for the appearance of FeTi-oxides on the liquidus, namely,  $1070^\circ\text{C}$  [Bell and Humphries, 1972], which were only observed in the G2011 tephra.



**Figure 2.** Concentrations of S (squares) and Cl (circles) in melt inclusions (filled symbols) and groundmass glass (open symbols) of four Grímsvötn eruptions. Chlorine did not form a volatile phase during the smaller 1823 and 2004 eruptions, resulting in identical concentrations in groundmass glass and clinopyroxene-hosted glass inclusions. See Table 1 for further information.

All MI pressure estimates from cpx-liquid equilibria are within the range of 0.6–6.4 kbar (mean 4.3 kbar). The frequency distribution in Figure 3 suggests that roughly 95% of the cpx crystallized between 2 and 6.4 kbar (MI mean: 4.3 kbar), while the rest originate from a shallower level corresponding to 0.6–1.5 kbar. Final  $P$ - $T$  equilibrium of cpx crystallization is obtained from euhedral crystal rims and adjacent glass.

#### 4.3.2. Grímsvötn 1873

Groundmass glass composition yields  $T$  range of  $1125$ – $1137^\circ\text{C}$ , consistent with two olivine-liquid temperatures ( $1126$  and  $1129^\circ\text{C}$ ) for two MI in olivines having  $\text{Fo}_{73}$  and  $\text{Fo}_{74}$ , respectively. The MI in cpx formed at  $T$  between  $1092$  and  $1150^\circ\text{C}$ , with a mean of  $1130^\circ\text{C}$  ( $n = 21$ ), which is indistinguishable (Table 2) to that obtained from the

**Table 2.** Result Summary of Temperature and Pressure Calculation for Four Historical Grímsvötn Eruptions<sup>a</sup>

|       |        | T Liq | T Ol-Liq | T Ol-MI | T Pl-MI | T Cpx-Liq | T Cpx-MI | P Cpx-Liq | P Cpx-MI |
|-------|--------|-------|----------|---------|---------|-----------|----------|-----------|----------|
| G1823 | min    | 1126  |          |         |         | 1099      | 1108     | −0.1      | 2.5      |
|       | max    | 1158  |          |         |         | 1163      | 1164     | 4.9       | 6.2      |
|       | median | 1136  |          |         |         | 1141      | 1133     | 4.1       | 3.8      |
|       | mean   | 1139  |          |         |         | 1139      | 1133     | 3.5       | 3.9      |
|       | N      | 11    |          |         |         | 9         | 14       | 9         | 14       |
| G1873 | min    | 1125  | 1140     | 1126    |         | 1116      | 1092     | 2.6       | 2.0      |
|       | max    | 1137  |          | 1129    |         | 1157      | 1150     | 5.9       | 6.0      |
|       | median | 1129  |          | 1128    |         | 1137      | 1129     | 3.9       | 4.0      |
|       | mean   | 1130  |          | 1128    |         | 1136      | 1130     | 4.0       | 4.1      |
|       | N      | 6     | 1        | 2       |         | 7         | 21       | 7         | 21       |
| G2004 | min    | 1073  | 1125     |         |         | 1082      | 1075     | 0.0       | 1.8      |
|       | max    | 1147  |          |         |         | 1148      | 1159     | 5.6       | 7.0      |
|       | median | 1113  |          |         |         | 1129      | 1120     | 4.3       | 3.8      |
|       | mean   | 1116  |          |         |         | 1124      | 1120     | 3.7       | 3.8      |
|       | N      | 15    | 1        |         |         | 13        | 40       | 13        | 40       |
| G2011 | min    | 1100  |          | 1212    | 117     | 1082      | 1065     | 0.1       | 0.6      |
|       | max    | 1149  |          |         | 1165    | 1150      | 1175     | 4.5       | 6.4      |
|       | median | 1128  |          |         | 1151    | 1116      | 1125     | 3.2       | 4.5      |
|       | mean   | 1129  |          |         | 1149    | 1118      | 1122     | 2.8       | 4.3      |
|       | N      | 21    |          | 1       | 21      | 10        | 82       | 10        | 82       |

<sup>a</sup>See Table S1 for the detailed results.

2011 tephra (given the associated error). Groundmass glass-cpx pairs  $T$  range is slightly wider (1075–1157°C) but their mean, 1129°C, is not different ( $n = 7$ ).

The melt inclusions in cpx formed at  $P$  of 2.0–6.0 kbar with a mean  $P$  of 4.1 kbar ( $n = 21$ ), again identical to what is observed for the products of the other large eruption (2011). Similar  $P$  is obtained for the final rim crystallization, or 2.6–5.9 kbar, with a mean of 4 kbar ( $n = 7$ ).

#### 4.3.3. Grímsvötn 2004 and 1823

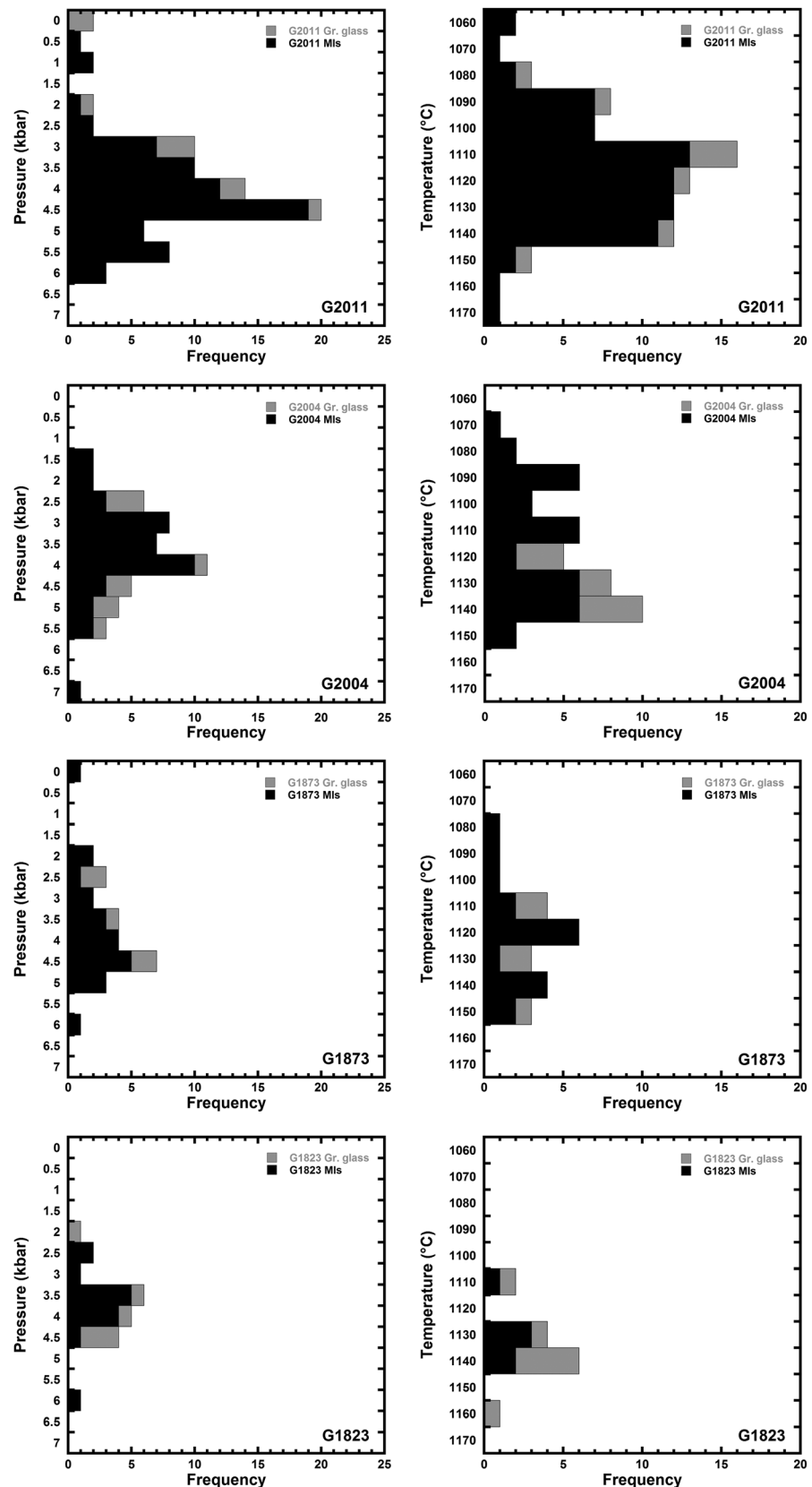
Crystallization  $T$ - $P$  conditions before the smaller eruptions are similar and thus discussed together. Olivine-hosted MIs were obtained for the 2004 eruption, with six crystals containing seven MI and two were surrounded by groundmass glass. Only a single pair has a  $K_D^{\text{ol-liq}}(\text{Fe-Mg})$  within the equilibrium range, giving  $T$  of  $1125 \pm 29^\circ\text{C}$ , identical to the mean groundmass glass-cpx  $T$  of  $1124^\circ\text{C}$  (range: 1082–1148°C;  $n = 13$ ) for the 2004 eruption. For that of 1823, the cpx-liq  $T$  range from 1099 to 1163°C (mean 1139°C;  $n = 9$ ). The groundmass glass-cpx barometry indicates crystallization from 0 to 5.6 kbar (mean 3.7 kbar,  $n = 13$ ; Table 2 and Figure 3) and between −0.1 and 4.9 kbar (mean 3.5 kbar,  $n = 9$ ) for the 2004 and 1823 products, respectively.

Similar  $P$ - $T$  conditions are obtained for MI formations in cpx or  $T$  in the range of 1075–1159°C (mean 1120°C;  $n = 40$ ) for the 2004 eruption, while fewer observations of the 1823 eruption product display a narrower range, 1108–1164°C (mean 1133°C;  $n = 14$ ), and  $P$  in the range 1.8 and 7.0 kbar (mean 3.8 kbar) and 2.5–6.2 kbar (mean 3.9 kbar) for the 2004 and 1823 eruptions, respectively.

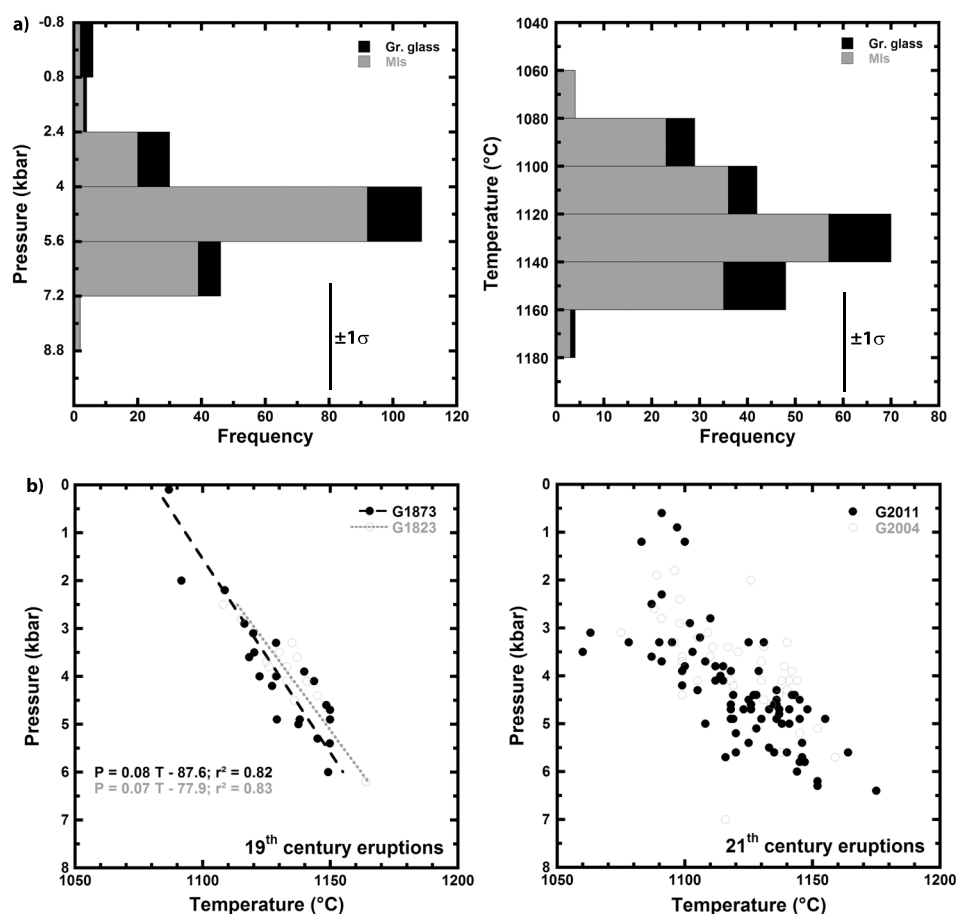
#### 4.3.4. Summary of Thermobarometric Results

The highest equilibrium pressure between cpx and MI exceeds 6 kbar for the four tephra studied. Crystallization at deeper depth cannot be excluded since only final equilibrium pressure is recorded in the crystal-melt composition during magma ascent. The average pressure of MI-cpx reaction for all four eruptions is  $4.1 \pm 1.1$  (SD; Figure 4a and Table S1). This pressure compares well with OPAM phase equilibrium calculated from the Grímsvötn MI glass composition giving  $P = 4.8 \text{ kbar} \pm 2.1$  (SD) on average and  $4.0 \text{ kbar} \pm 1.7$  (SD) from groundmass glass composition (Table S2).

The MI equilibrium temperature may suggest slow cooling of the Grímsvötn magma system with time. The youngest tephra (G2011 and G2004) record identical mean cpx-MI temperature, or 1120°C and 1122°C. Nominally higher temperature (1130°C) is calculated for the older G1873 and G1823 tephra, and highest preeruptive estimates of ~1140°C have been suggested for 1783–1784 Laki eruption and the ~10 ka Saksunarvatn tephra [Bindeman *et al.*, 2006; Neave *et al.*, 2015, and references therein]. Although these



**Figure 3.** Grímsvötn thermobarometry results based on clinopyroxene and melt equilibrium in the four tephra samples studied. Melt inclusions and groundmass glass are abbreviated MI and Gr. glass, respectively.



**Figure 4.** Pressure versus temperature from cpx-liq in Grímsvötn tephra. (a) Combined thermobarometry results from the four eruptions studied. (b) Linear correlation is observed only for the 19th century eruptions. The 21st century results suggest disconnected magma system at depth; see text for further discussion.

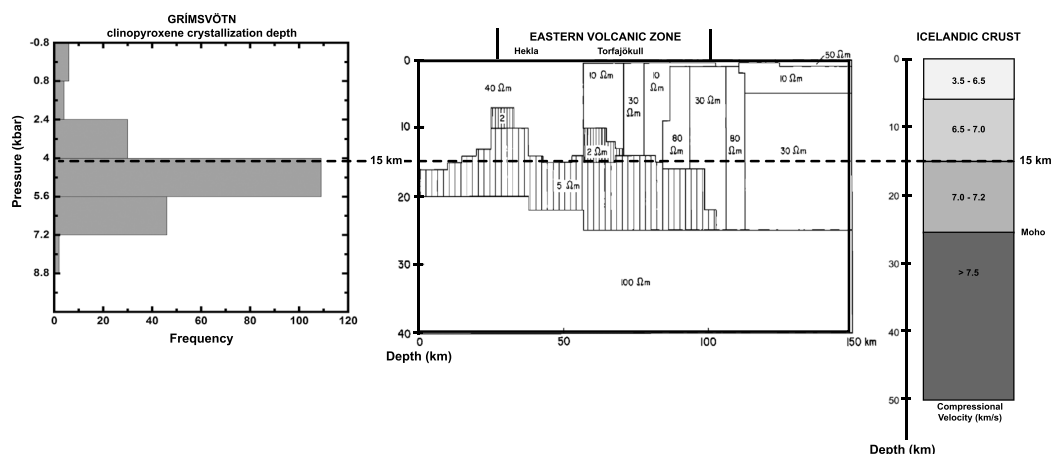
estimates are within the proposed thermometer errors [e.g., Putirka, 2008], similar pattern is observed in the final temperature recorded in cpx—groundmass glass thermometers (Table 2). A regular nominal decrease in calculated temperature with time is apparent (G1823 mean = 1139°C, G1873 mean = 1136°C, G2004 mean = 1124°C, and G2011 mean = 1118°C; Figure S4). Finally, the lowest recorded temperature, 1065°C, is from the G2011 tephra in which the presence of FeTi-oxides was identified for the first time in Grímsvötn products [Sigmarsson *et al.*, 2013].

## 5. Discussion

In this discussion both thermobarometry results and volatile concentrations measured in glasses of MI and groundmass are combined with the goal of putting constraints on the magma system behavior beneath Grímsvötn volcano.

### 5.1. Crystallization Depth and Magma Storage Beneath Grímsvötn

The average pressure recorded in the composition of the melt inclusions and clinopyroxene host crystals from the four Grímsvötn eruptions is approximately 4 kbar, whereas the spread in the results suggests crystallization within the entire crust, namely, cpx-hosted MI pressure between 0.6 and 7 kbar, i.e., 2 to 25 km. The results reveal the depth range of the magma system activated during each eruption and emphasize the frequency of cpx arising from approximately 15 km depth. Crystallization depth for two of the largest known or inferred eruptions at Grímsvötn, namely, the Laki and Saksunarvatn's eruptions, was studied by Neave *et al.* [2013, 2015], yielding a mean of 3.5–3.8 kbar ( $\pm 1.0$ –0.6; 1 SD) for magnesian cpx and lower pressure for more

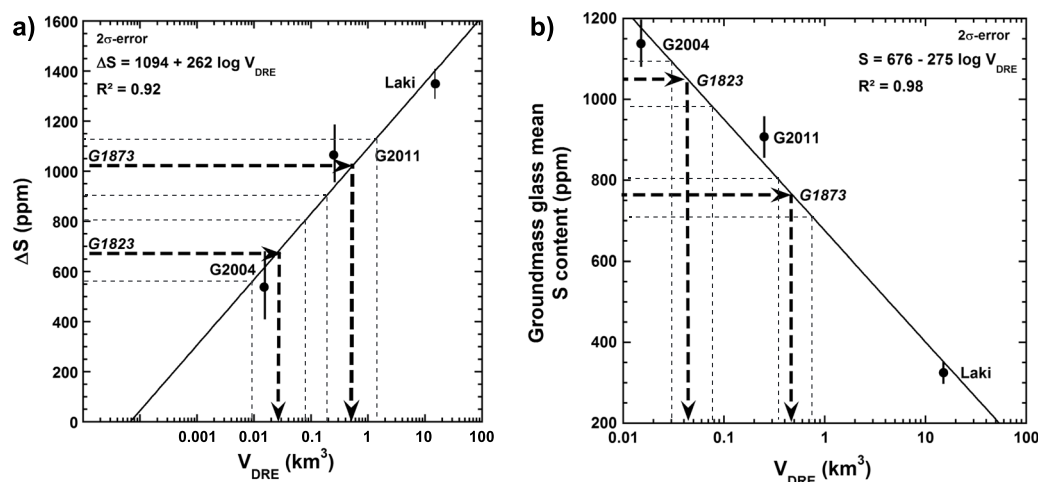


**Figure 5.** Clinopyroxene crystallization pressure estimated from equilibria with melt inclusions and groundmass glass from four Grímsvötn eruptions. The corresponding average depth estimate of  $15 \pm 5$  km correlates with a layer of low-resistivity proposed by Eysteinsson and Hermance [1985] and the limit of middle-deep crust inferred from compressional seismic wave velocities [Bjarnason, 2008].

evolved cpx [Neave and Putirka, 2017]. Combined MI and groundmass glass compositions yield OPAM phase equilibria at  $P = 4.6 \text{ kbar} \pm 2.0 \text{ (SD)}$ . These estimates thus give crystallization pressure within error of those obtained for the 1823, 1873, 2004, and 2011 eruptions at Grímsvötn. Consequently, it can be concluded that consistent results are obtained for the principal depth of Grímsvötn magma crystallization, namely, 13–17 km (3.5–4.8 kbar), and that different eruption magnitudes are not caused by difference in crystallization conditions.

In contrast to pressure, calculated equilibrium temperature appears to be time-dependent. Both  $\Delta T$  of cpx-liq calculated from host crystal and MI and the  $\Delta T$  calculated from the melt composition alone reveal a larger range for the 21st century eruption compared to that of the 19th century, or respectively 84–110°C and 56–58°C. This may suggest significant difference in thermal conditions between older and younger eruptions rather than between small and large eruptions at Grímsvötn. These differences could be explained by either different proportions and/or decreasing mixing efficiency of deep-derived basalts with more evolved high-level basalts today within the active magma system compared to the past. The lack of correlation between  $P$  and  $T$  for the 21st century eruptions may indeed suggest a magmatic system constituted of dykes and sills in which magma bodies are stored at different  $P$  and  $T$  conditions. The potential temperature decrease with time concurs with increasing concentrations of incompatible elements in historical Grímsvötn tephra caused by progressive fractional crystallization [Sigmarsson et al., 2016]. In Figure 4b the linear correlation between  $T$  and  $P$  for the 19th century eruptions is shown. Linear extrapolation of geothermal gradient in Iceland, as derived from geothermal exploration boreholes in Iceland, may indicate the depth to the 1200°C isotherm [Flóvenz and Sæmundsson, 1993]. Extrapolation of the cpx equilibrium crystallization  $P$ - $T$  correlation yields a lower crustal pressure of 9 kbar for the 1200°C isotherm. Assuming an average crustal density of  $2750 \text{ kg/m}^3$ , a hydrostatic pressure of 9 kbar corresponds to approximately 30 km crustal depth. This depth corresponds to the estimated crustal thickness from seismic tomography [e.g., Bjarnason, 2008]. A crust-mantle boundary is likely to act as a neutral buoyancy level at which mantle melt can accumulate and from where the Grímsvötn magma system is supplied.

Ascending basaltic magma is stored higher up in the crust where cpx dominantly crystallizes. That depth corresponds to the limit of middle and lower crust as inferred from geophysical observations (Figure 5). The magnetotelluric survey of Eysteinsson and Hermance [1985] suggests a layer (~5 km thick) of low resistivity at similar depth across the Eastern Volcanic Zone of Iceland. This layer corresponds to the lower crust as proposed from compressional seismic velocity by Bjarnason [2008]. Similar crystallization depth is also observed from cpx-melt relationship in the 2014–2015 Holuhraun lavas (Table S3) despite their more primitive olivine tholeiite composition, which concurs with earlier results by Bali et al. [2015], Gíslason et al. [2015], and Gudmundsson et al. [2016]. An important magma crystallization and storage at approximately  $15 \pm 5$  km depth is thus likely beneath the active volcanic systems of Iceland.



**Figure 6.** Relationship between the dense-rock-equivalent (DRE) volume and (a) the quantity of S exsolved ( $\Delta S = S_{\text{max MI}} - S_{\text{mean groundmass glass}}$ ; see text) and (b) the groundmass glass mean sulfur concentrations (in ppm) of different eruptions at the Grímsvötn volcanic system (volume estimates from Thordarson and Self [1993], Jude-Eton et al. [2012], and Gudmundsson et al. [2012]). The broken arrows highlight the  $V_{DRE}$  for the two 19th century eruptions estimated from the lognormal correlations.

## 5.2. Volatile Budget, Degassing, Eruption Volume, and Magma Recharge

Magma of the larger eruptions (1873 and 2011 tephra) erupted cpx-hosted MI with slightly higher S and Cl concentrations compared to those of the smaller phreatomagmatic eruptions (1823 and 2004 tephra; Figure 2). This suggests higher volatile concentrations in parental magma of the larger eruptions and/or higher proportions of magma recharge from depth compared to that of the smaller eruptions. Thus, the question whether different volatile concentrations contribute to the variable eruption regimes at Grímsvötn is worth exploring.

### 5.2.1. Sulfur and Eruption Volume

The maximum sulfur concentrations measured in the G2011 MI (max S = 1974 ppm; Table 1) and the average for the groundmass glass (908 ppm S) lead to 1.47 Tg of  $\text{SO}_2$  exsolved from the initial magma using the petrologic method, which is within error of earlier estimate ([Sigmarsson et al., 2013]  $\Delta S$  values of 1066 versus 1063 ppm, respectively). For the Grímsvötn 2004 eruption, magma volume ( $2.1 \pm 0.4 \times 10^{-2} \text{ km}^3$ ) estimated by Jude-Eton et al. [2012] gives 0.06 Tg of exsolved  $\text{SO}_2$ . How much of the sulfur was sequestered as sulfides and how much was emitted to the atmosphere cannot be evaluated due to lack of precise satellite observations during the small 2004 eruption. However, absence of sulfides in Grímsvötn tephra older than that of 2011 eruption suggests minimal S sequestration in sulfides of the older eruptions. The onset of titanomagnetite crystallization before the 2011 eruption corresponds to the progressive magma evolution at Grímsvötn during most of the historical period [Sigmarsson et al., 2016]. Oxygen fugacity ( $f_{\text{O}_2}$ ), which is close to the FMQ-buffer, decreases with the titanomagnetite precipitation and that will stabilize sulfur as a sulfide species [e.g., Jugo et al., 2010]. Relatively low  $f_{\text{O}_2}$  in the 2011 magma caused approximately 10% of the degassing sulfur as the reduced  $\text{H}_2\text{S}$  and important sulfur sequestration as sulfides [Sigmarsson et al., 2013]. Absence of volume estimates for the 19th century tephra hampers discussion of exsolved  $\text{SO}_2$ , although crude estimations could be given using the determined  $\Delta S$  (Figure 6a) for different eruptions on the Grímsvötn volcanic system.

For the Laki (1783–1784) eruption, Thordarson et al. [1996] estimated  $\text{SO}_2$  emission of 122 Tg for a volume of  $15.1 \text{ km}^3$ , which corresponds to  $8.1 \text{ Tg/km}^3$ . The Grímsvötn 2011 eruption emitted  $1.47 (\pm 0.37) \text{ Tg}$  for a volume of  $0.25 (\pm 0.05) \text{ km}^3$  which is equivalent to  $5.9 \text{ Tg/km}^3$ . Hreinsdóttir et al. [2014] estimated that 95% of the volcanic deformation occurred during the first 24 h corresponding to emission of 1.4 Tg  $\text{SO}_2$  and a mean flux of 58 Gg of  $\text{SO}_2$  per hour. In comparison, the Laki event may have produced 60% of its volatile emissions during the first 45 days [Thordarson et al., 1996], which is equivalent to a mean flux of 68 Gg of  $\text{SO}_2$  per hour. Hence, these two very different eruptions of the same volcanic system appear to have produced similar  $\text{SO}_2$  fluxes. These fluxes are more than an order of magnitude higher than that observed

from the youngest flood basalt eruption in Iceland, namely, the Holuhraun 2014–2015 eruption (2.4 Gg/h on average [Gauthier *et al.*, 2016]). The different SO<sub>2</sub> fluxes from Grímsvötn and Bárðarbunga eruptions are readily explained by higher volatile concentrations, and their incompatible element behavior, in the more evolved tholeiitic basalt produced at Grímsvötn.

There is an apparent relationship between dense rock equivalent (DRE) volume of erupted magma and its associated  $\Delta S$  for eruptions at Grímsvötn volcanic system (Figure 6a). Magma volumes of the Laki, G2011, and G2004 eruptions correlate with  $\Delta S$ , with a correlation coefficient ( $r^2$ ) of 0.92. The Laki magma has the largest  $\Delta S$ , that of 2011 an intermediate value and the phreatomagmatic 2004 eruption the lowest. Using the  $\Delta S$  values measured (Table 1) for the 19th century eruptions, it is possible to evaluate their eruption volumes. Volume estimates of  $3 \times 10^{-2}$  km<sup>3</sup> for G1823 and 0.5 km<sup>3</sup> for G1873 are obtained, which are coherent with available written eruption descriptions (see Text S1 in supporting information). Although this relationship is only based on three observations, future Grímsvötn eruptions will serve to prove or reject this relationship.

Quantifying the S degassing, or rather the exsolution from the silicate liquid in case that part of the S is sequestered in sulfides, demonstrates the difference between the larger and the smaller eruptions. The relative exsolution is calculated as follows:  $D_S = 100 * \Delta S / S_{\text{initial}}$ , with  $\Delta S = S_{\text{initial}} - S_{\text{groundmass}}$  ( $S_{\text{initial}}$  = maximum S in MI), and leads to identical values for the plinian eruptions with  $D_S$  (G1873)  $\approx D_S$  (G2011)  $\approx 55\%$  (Table 1). Significantly lower values are obtained for the phreatomagmatic events, or  $D_S$  (G1823) = 39% and  $D_S$  (G2004) = 32%. In part, this is caused by variable  $\Delta S$ :  $\Delta S$  (G2011) = 1066 ppm, close to  $\Delta S$  (G1873) = 1023 ppm, whereas  $\Delta S$  (G2004) = 539 ppm is close to  $\Delta S$  (G1823) = 681 ppm. Therefore, the larger Grímsvötn eruptions not only had higher initial S concentrations but are also able to exsolve more sulfur than the smaller phreatomagmatic events. Obviously, the magma quenching effect due to efficient contact with glacial water arrests melt degassing of the less powerful eruptions. Higher mass eruption rate leads to lower water-melt contact and thus later melt quenching and enhanced outgassing. This is illustrated in Figure 6b, with increasing degree of degassing causing decreasing S concentrations in the groundmass glass in the order phreatomagmatic (1823 and 2004), plinian (1873 and 2011), and flood basalt eruptions (1783–1784 Laki). Similar phenomenon was observed for Holocene Katla tephra by Óladóttir *et al.* [2007]. If future eruptions from Grímsvötn confirm the proposed relationship between  $\Delta S$ , groundmass S concentrations and magma volume erupted, then the volcano's production rate can be quantified from analysis of its historical tephra.

### 5.2.2. Chlorine

The degassing pattern of S is reflected by that of Cl. For G2011 and G1873 eruptions, the  $\Delta \text{Cl}$  is 131 ppm and 86 ppm, but the tephra of both phreatomagmatic eruptions have indistinguishable Cl concentrations in groundmass and MI glasses (Table 1 and Figure 2). Chlorine degassing could thus have been prevented by early quenching due to important water-melt interaction during the small eruptions, similar to the arrested S degassing. Alternatively, the fact that Cl solubility in basaltic melt is high [e.g., Alletti *et al.*, 2009; Lesne *et al.*, 2011], magma fragmentation may have occurred deeper than Cl volatilization. This latter possibility would imply, for a similar fragmentation depth, higher volatile content of the magma causing the larger eruptions.

The  $D_{\text{Cl}}$  (G2011) = 39% and  $D_{\text{Cl}}$  (G1873) = 32%, which leads to the mobilization of 0.1 Tg of HCl during the Grímsvötn 2011 eruption, using tholeiite magma density (2750 kg/m<sup>3</sup>) and dense rock volume equivalent (0.27 km<sup>3</sup> [Hreinsdóttir *et al.*, 2014]) of magma emitted. This atmospheric mass loading is thousand times higher than that (0.1 Mt HCl) of the Holuhraun 2014–2015 fissure eruption (1.6 km<sup>3</sup> [Gíslason *et al.*, 2015]) despite an order of magnitude lower eruption volume. Chlorine concentrations are thus close to saturation in the quartz-normative tholeiitic melt under Grímsvötn. Taken together, the volatile results suggest degassing of sulfur before that of chlorine in the Grímsvötn tholeiites, as expected from the classical basaltic exsolution sequence of volatile elements, and higher volatile content in the magma of the larger eruptions.

### 5.2.3. Magma Recharge

Detailed tephrostratigraphy on the outlet glaciers of Vatnajökull (Figure 1) by Larsen *et al.* [1998] revealed a succession of high versus low eruption frequency phases at Grímsvötn volcano since at least the year 1200 Common Era (from the oldest tephra layer identified on the glaciers). A period of ~140 years and a similar duration of the two distinct phases, namely, 50–80 years, were identified. This conclusion is verified by the

138 years elapsing between the last two major Grímsvötn eruptions (2011 and 1873). These larger eruptions produce basalt with relatively high compatible element concentrations, such as Ni and Cr, attributed to magma recharge from depth [Sigmarsson *et al.*, 2016]. Therefore, the larger explosivity of eruptions with plinian activity is most likely due to important recharge of deep-derived and volatile-rich magmas, supported by higher maximum sulfur and chlorine concentrations in melt inclusions of the plinian tephra (Figure 2). The initial volatile concentrations and the relative proportions of recharge magma versus residing magma thus appear to control the eruption regime at Grímsvötn volcano.

## 6. Conclusions

The principal conclusions are as follows:

1. Average pressure estimates of clinopyroxene crystallization in large and small eruptions at Grímsvötn volcano give of  $4.1 \pm 1.1$  kbar, corresponding to approximately  $15 \pm 5$  km depth. Absence of good temperature-pressure correlation for the 21st century eruptions suggests that the erupted magma derive from separated magma bodies with different thermal evolution, possibly in a plexus of sills and dikes.
2. Similar average crystallization depths of magma emitted in the four eruptions studied cannot explain different eruption regimes at Grímsvötn.
3. Sulfur and chlorine concentrations being highest in the melt inclusions of the larger eruptions (up to 1974 ppm and 339 ppm, respectively) are lowest in their groundmass glass. This indicates important recharge of deep-seated gas-rich magma at the origin of the tropopause-penetrating eruptions.
4. Quenching with glacial water explains higher volatile concentrations in groundmass glass for the smaller phreatomagmatic eruptions. The differences in volatile concentrations measured in melt inclusions and groundmass glass (i.e.,  $\Delta S$ ) and the average concentration in the groundmass of each tephra appear to correlate with known eruptive volume for the 21st century eruptions and that of the Laki eruption, potentially permitting estimates of unknown volumes for older eruptions.

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