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Evolution of a silicic magma reservoir in the upper crust

Reyðarártindur pluton, Southeast Iceland

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Abstract
Rhodes, E. 2022. Evolution of a silicic magma reservoir in the upper crust. Reyðarártindur pluton, Southeast Iceland. *Digital Comprehensive Summaries of Uppsala Dissertations from the Faculty of Science and Technology* 2151. 67 pp. Uppsala: Acta Universitatis Upsaliensis. ISBN 978-91-513-1505-8.

Field observations of extinct and exposed magma reservoirs shed light on processes operating in the roots of presently active volcanoes. The Reyðarártindur pluton, Southeast Iceland is an example of a fossil shallow magma reservoir that fed eruptions. The different chapters in this thesis examine the accumulation of magma, and processes occurring during the development and evolution of the magma reservoir from different methodological perspectives. A final model for the evolution of the Reyðarártindur pluton is then presented.

The majority of the pluton consists of one voluminous rock unit, the Main Granite, that formed by rapid magma emplacement. However, a local zone of geochemically distinct, but related further Granite Enclaves and Quartz Monzonite Enclaves attest to variations in the composition of the underlying source reservoir. Space for the ca. 2.5 km^3 of magma in the pluton was made by piecemeal floor subsidence, which began with multiple dykes that then propagated laterally to form flat-roofed intrusions at different depths. During the first stages of magma emplacement, shattering, sintering and sanidinite-facies contact metamorphism affected a ca. 10 m thick zone of the basalt host rock at the magma reservoir roof. The resulting hornfels was stronger than the original altered basalt, and contained zero porosity and permeability. It thus formed a ‘cap-rock’ to the magma reservoir, limiting heat, volatile and fluid transfer until fractured and faulted at a later stage.

The magma reservoir erupted at least once, causing local subsidence of the roof, which would have been observable at the Earth’s surface. Recharge of the magma reservoir by the same Quartz Monzonite and further Granite as exposed in the Reyðará River led to overpressure and eruption. We envisage that cooling and sealing of the piecemeal subsidence network preceded eruption, causing overpressure with magma recharge. The eruptive lifetime of the magma reservoir was limited to ca. 1000 years. This timeframe is much less than the duration of silicic magmatism in a typical Icelandic central volcano, or at other rhyolite-erupting volcanoes worldwide, which is in the order of hundreds of thousands to millions of years. Hence, the Reyðarártindur pluton likely represents a small, ephemeral part of a wider magmatic plumbing system that feeds a central volcano.

The results from these studies can provide volcano-monitoring personnel with scenarios for magma emplacement, and processes leading to eruption, which they can then use as a framework for interpreting detectable signals of magma movement and volcanic unrest.

Keywords: magma, magma emplacement, magma reservoir, floor subsidence, piecemeal subsidence, granite, pluton, cap-rock, magma mingling, magma plumbing systems, volcanic and igneous plumbing systems, eruption

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To the unknown unknowns

Plain Language Summary

Scientists who monitor active volcanoes are interested in the pathways that magma takes on its journey through the Earth's crust to eruption. In Iceland it is known that magma often moves from chambers deep in the crust where the magma is generated (>20 km) to shallow levels (<5 km) where the magma is stored for a time prior to eruption or solidification. Scientists cannot physically 'observe' active magma chambers in the crust. Therefore, we turn to older, solidified 'fossil' magma chambers exposed at the Earth's surface to examine what processes occurred during their development, and what might have led to eruption. Eruption forecasters can use these results to help align the surface deformation and seismic signals recorded at active volcanoes with what could be happening in the subsurface. This thesis examines Reyðarártindur, a shallow fossil magma chamber in Iceland that was active 7 million years ago, and deciphered its history from formation to eruption. This required extensive geological mapping of the rocks within and surrounding the magma chamber, and systematic sampling of the magma chamber rocks for geochemical analysis.

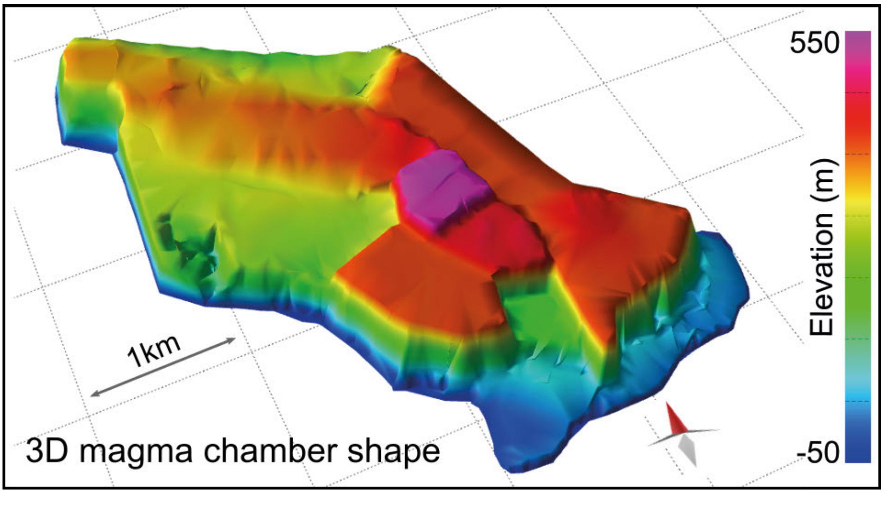

Figure 1: Left: Results of the shape reconstruction of the Reyðarártindur magma chamber. In 3D, the different steps or levels in the top of the magma chamber (the magma chambers 'roof') are clearly visible, giving the magma chamber a castle-like shape.

The studies in the thesis found that the magma chamber has a castle-like shape with a minimum volume of 2.5 km^3 (see Figure 1). The different magmas that make up the chamber are all high in silica content, which is important from an eruption hazard perspective, because high silica magmas have the potential to cause large, explosive eruptions. For example, the famous Eyjafjallajökull eruption that impacted airspace over Europe was produced by silicic magma.

Reyðarártindur was assembled rapidly by a mechanism that is called 'floor subsidence'. Floor subsidence occurs when space is made for the incoming magma by the 'floor' of the chamber dropping down, subsiding into the underlying magma reservoir that the new magma was coming from (see Figure 2). In the classic floor subsidence model, there is one subsiding floor block, which produces a flat magma chamber roof such as displayed in Figure 2. However, at Reyðarártindur the castle-like shape with steps in the roof indicates there must have been several different subsiding floor blocks, such as shown in Figure 3. This mechanism also implies that there must have been a wider, more voluminous magma chamber directly beneath the Reyðarártindur magma chamber.

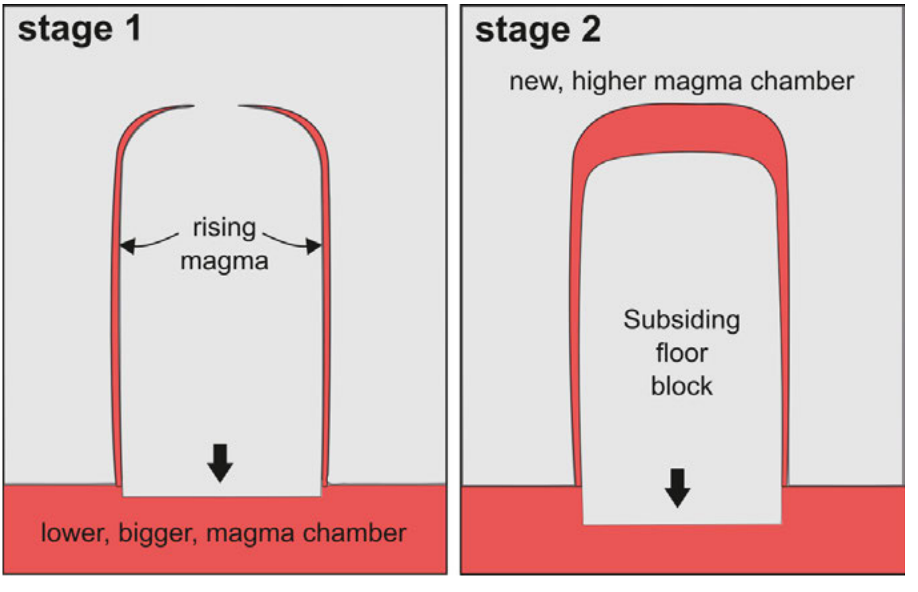

Figure 2: An illustration of how the floor subsidence mechanism functions to make space for a new magma chamber. The volume of magma in the upper magma chamber roughly corresponds to the volume of magma removed from the lower magma chamber.

As Reyðarártindur was forming, the intruding magma shattered the overlying layer of the roof rocks. Following this, the high temperatures metamorphosed the rocks and sealed them together, forming a lid-like cap-rock (Figure 3). The new cap-rock was stronger than the original rock, and limited heat loss from the magma, insulating it from rapid cooling and solidification. Additionally, the high strength and extremely low permeability of the cap-rock inhibited degassing, allowing gas pressure to build up within the magma chamber. As the gas pressure is one of the main parameters that influence eruption initiation and size, the formation of the cap-rock may have meant that an eruption was more likely to initiate, and once initiated, would be more explosive.

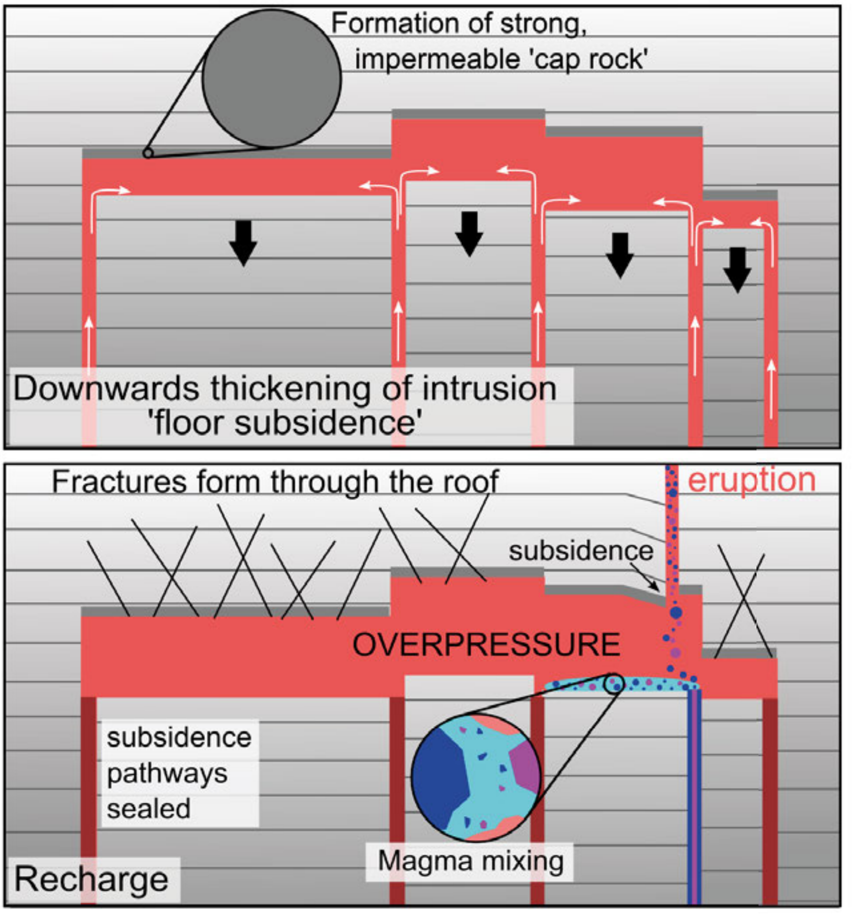

Figure 3: Model presented for the evolution of the Reyðarártindur magma chamber, from growth to eruption. In the top diagram, the magma chamber is grown by floor subsidence, where multiple blocks of the floor subside separately, which gives the roof the 'stepped' shape in 3D. In the bottom diagram, the magma feeder pathways that accommodated subsidence of the blocks sealed. When magma recharge occurred, forming the magma mixing patterns observed in the field, there was overpressure in the magma chamber. The overpressure fractured the magma chamber cap-rock and roof, eventually leading to an eruption.

In fact, I found evidence that the Reyðarártindur magma chamber erupted at least once. The magmatic rocks preserved in the conduits have typical textures of explosive volcanic rocks, in contrast to the coherent magmatic rocks inside the chamber. Hence, Reyðarártindur fed at least one explosive eruption, like the one that occurred in 2010 at Eyjafjallajökull volcano, rather than an effusive eruption, like the 2021 Fagradalfjall eruption. New magma coming into the already established magma chamber (a process known as magma recharge) triggered the eruption. The new magma was also high-silica in composition and was preserved while mixing with the magma from the Reyðarártindur magma chamber in the conduits. Magma mixing textures were also observed near the bottom of the chamber (Figure 4). The high-silica composition of the incoming magma is also notable, because it is considered difficult to erupt high-silica magma without recharge by mafic magma. Mafic magma is hotter, more liquid and contains an abundance of gasses, all which can help trigger a cooler, 'thicker' silicic magma to erupt. Instead, a new conceptual model for eruption is presented from these results, as follows (see Figure 3): the magma chamber floor stopped subsiding with the incoming magma. Hence, pressure built up in the Reyðarártindur magma chamber until the roof 'broke' and magma could travel to, and erupt at, the Earth's surface. The pressure build-up caused lots of smaller fractures and faults in the magma chamber roof/cap-rock before eruption, the formation of which would have been detectable via volcano monitoring equipment. The eruption of magma caused the roof surrounding the conduit to subside, and a depression likely formed at the site on the Earth's surface. The depression would have been filled with the erupted magma as ash, and/or pyroclastic deposits.

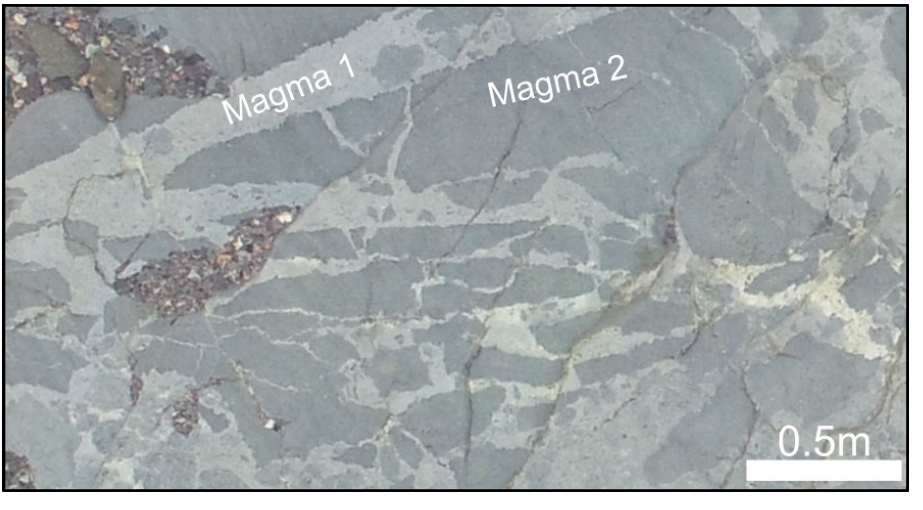

Figure 4: Photo of two magmas preserved mixing in the pluton.

Modelling of the cooling of the magma chamber determined that it could have been active and fed eruptions for up to 1000 years. A typical volcano in Iceland is active for 1 million years, so this magma chamber was likely only a short-lived part of a wider system of many magma chambers that fed a volcano. This anatomical study of how a magma chamber of silicic magma assembled and eventually erupted will help volcano monitoring teams interpret ground deformation and seismic signals at active volcanoes today.

List of Papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals.

I. **Rhodes, E**., Barker, A. K., Burchardt, S., Hieronymus, C. F., Rousku, S. N., McGarvie, D. W., Mattsson, T., Schmiedel, T., Ronchin., E., Witcher, T. (2021) Rapid assembly and eruption of a shallow silicic magma reservoir, Reyðarártindur pluton, Southeast Iceland. *Geochemistry, Geophysics, Geosystems*, 1–26.

II. **Rhodes, E**., Burchardt, S., Greiner, S. M. H., Mattsson, T., Sigmundsson, F., Schmiedel, T., Barker, A. K., Ronchin, E., Witcher, T. (2022) Volcanic unrest as seen from the magmatic source- Reyðarártindur pluton, Iceland. *Manuscript submitted to Scientific Reports.*

III. **Rhodes, E**., Burchardt, S., Tuffen, H., Heap, M. J., Wadsworth, F. B., Barker, A. K., Chun, H. H., McGarvie, D. W., Witcher, T., Schmiedel, T., Unwin, H. E. (2022) Cap-rock formation above a magma reservoir, Reyðarártindur pluton, Iceland. *Manuscript in preparation.*

Additional Papers

In addition, the author has contributed to the following papers, which are not included in the thesis:

I. **Rhodes, E.,** Kennedy, B. M., Lavallée, Y., Hornby, A., Edwards, M., & Chigna, G. (2018). Textural insights into the evolving lava dome cycles at Santiaguito Lava Dome, Guatemala. *Frontiers in Earth Science*, 6, 1–18.

II. Lavallée, Y., Dingwell, D. B., Johnson, J. B., Cimarelli, C., Hornby, A. J., Kendrick, J. E., von Aulock, F W., Kennedy, B M., Wadsworth F B., **Rhodes, E**., Chigna, G. (2015). Thermal vesiculation during volcanic eruptions. *Nature* 528, 544–547.

III. Di Baldassarre, G., Nohrstedt, D., Mård, J., Burchardt, S., Albin, C., Bondesson, S., Breinl, K., Deegan, F. M., Fuentes, D., Girons-Lopez, M., Granberg, M., Nyberg, L., Rydstedt-Nyman, M., **Rhodes, E.,** et al. (2018). An Integrative Research Framework to Unravel the Interplay of Natural Hazards and Vulnerabilities. *Earth's Future*, 6(3), 305–310.

IV. Schipper, C.I., Castro, J.M., Kennedy, B.M. Tuffen, H., Whattam, J., Wadsworth, F., Paisley, R., Fitzgerald, R., **Rhodes, E.**, et al., (2021) Silicic conduits as supersized tuffisites: Clastogenic influences on shifting eruption styles at Cordón Caulle volcano (Chile). *Bulletin of Volcanology* 83, 11

Contributions

The papers included in this thesis are the result of collaboration with several authors. My contributions and those of the co-authors to each paper are summarised below:

I. 80% of the total effort. Steffi Burchardt initiated the study. I formulated the research question together with Steffi Burchardt and Abigail Barker. I undertook fieldwork and collected samples with Steffi Burchardt, Abigail Barker, Tobias Mattsson, Erika Ronchin, Tobias Schmiedel and Taylor Witcher. I processed the samples and the data under the supervision of Abigail Barker. Christoph Hieronymus conducted the numerical cooling model. Sabine Rousku traced the enclaves in photos and conducted the statistical analysis on their proportions and size. I wrote the first manuscript with input on data interpretation from Abigail Barker and Dave McGarvie. All other co-authors read and reviewed the manuscript and I implemented the comments.

II. 70% of the total effort. Steffi Burchardt initiated the study and formulated the research question. I undertook the fieldwork with assistance from Steffi Burchardt, Abigail Barker, Tobias Mattsson, Erika Ronchin, Tobias Schmiedel and Taylor Witcher. I processed the photogrammetry data with some assistance from Tobias Mattsson. Sonja Greiner constructed the numerical deformation model with oversight from Freysteinn Sigmundsson and Halldór Geirsson. Together Steffi Burchardt and I wrote the manuscript, which was then reviewed by the co-authors.

III. 80% of the total effort. The study was initiated in the field by myself, Steffi Burchardt, Hugh Tuffen, Fabian Wadsworth and Mike Heap. I processed the samples for geochemistry, thin section, and EMP analysis. Mike Heap measured the tensile strength and permeability of the rocks. The results were discussed between the coauthors before the first manuscript draft was written. I wrote the first draft of the manuscript, which was then reviewed by the coauthors. I implemented their comments for the final presented version.

Contents

Abbreviations

3D	Three-Dimensional
MAR	Mid-Atlantic Ridge
MOHO	Mohorovicic Discontinuity
SIIS	Southeast Iceland Intrusive Suite
UAV	Unmanned Aerial Vehicle
WDS	Wavelength Dispersive X-ray Spectroscopy
SEM	Scanning Electron Microscope
EDS	Energy Dispersive Spectroscopy
LOI	Loss on Ignition
REE	Rare Earth Element
HFSE	High Field Strength Element
LILE	Large Ion Lithophile Element
FFD	Fractures, Faults and Dykes
IDDP	Iceland Deep Drilling Project

Introduction

Volcanic eruptions are one of the most spectacular and powerful phenomena on Earth. There are more than 1000 active volcanoes on the planet, of which approximately 40 are erupting on any given day (Global Volcanism Program, 2013). Hazards from volcanoes can range from small scale and local lava flows to devastating pyroclastic flows, thick widespread ash and tephra fall-out, and secondary lahars (Cashman *et al.*, 2013). In particular, eruptions that are generated from magma of high-silica compositions are generally more explosive and are therefore associated with the highest risk for people and infrastructure (Sparks *et al.*, 1977). Therefore, forecasting eruptions, and their style and scale, is essential to respond with timely hazard mitigation response that will save lives (Fig. 1).

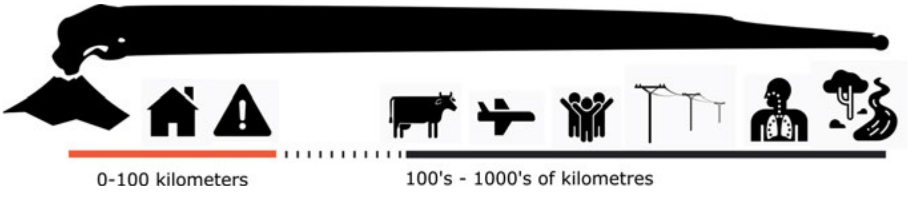

Figure 1: Hazards and impacts from volcanic eruptions are diverse and can be far reaching.

Eruption forecasting requires a detailed understanding of the fundamental processes driving volcanic activity, locations of magma storage, and processes occurring within the magmatic system (Sparks, 2003; National Academies of Sciences Engineering and Medicine, 2017). Volcano scientists and observatory personnel use this information to interpret the phenomena and scenarios behind detectable signals such as deformation, seismicity, and gas emissions (Sparks and Cashman, 2017; Sigmundsson et al., 2018). The information is then used to assess the eruption potential, eruptive style and eruption magnitude, which is communicated to civil authorities and emergency management, who can facilitate hazard response, such as evacuation, if need be (Sparks, 2003).

This thesis focuses on the development of a high-silica magma reservoir in the upper crust, often the last place for magma accumulation prior to eruption.

The central questions driving the research are: How do silicic magma reservoirs in the upper crust assemble and evolve? What is the eruptive potential and longevity of such a magma reservoir? What triggers them to produce eruptions? Moreover, how can the answers to the above questions help to improve the interpretation of volcanic unrest and hazards mitigation? The thesis uses the Reyðarártindur pluton, a 'fossil' magma reservoir exposed in Southeast Iceland, as a case study to explore the above questions, because it represents the remnants of a once active magma reservoir. The three manuscripts presented use a variety of methods to approach the above questions from different angles.

Paper I focuses on determining the size and shape of the pluton, and establishing the composition, stratigraphy and relationship of the magmatic rock units. A thermal model is then developed to determine cooling timeframes, and therefore eruptive potential of the magma reservoir. The methods used are field mapping, petrology, geochemistry, 3D pluton reconstruction and a thermal numerical model. It was discovered that the former magma reservoir erupted from at least three locations. A first conceptual model for construction of the Reyðarártindur pluton is presented.

Paper II explores how space was created for the 2.5 km^3 of incoming silicic magma, i.e. the emplacement mechanisms. Quantification of deformation associated with eruption of magma was additionally conducted. The methods used to do this were field mapping, photogrammetry analysis of the host rock, and a finite element model that simulates the eruption-related deformation. The results are then used to discuss the likely detectable seismic and deformation signals related to the emplacement, growth, and eruption of the Reyðarártindur magma reservoir.

Paper III investigates the interface between the magma reservoir and the basaltic host rock, specifically the deformation and metamorphism of the host rock to form a 'cap-rock' above the magma reservoir. The methods employed to do this are field mapping, compositional analyses, geothermobarometry, rock strength testing and partial-melting modelling. A model for cap-rock formation was proposed from the results. The paper then considers what the formation and properties of the cap-rock mean for the failure of the magma reservoir roof, and for heat transfer between the magma and the host rock.

The results from **Papers I-III** are then combined to produce a full model for the evolution of the Reyðarártindur pluton, which is presented in the **Conclusions and Outlook** section of this thesis.

Background

Magma is generated by partial melting of the Earth's mantle and the lower crust at the mantle-crust boundary, either by the addition of heat, volatiles or by decompression driven partial melting (Grove and Till, 2015). From the mantle-crust boundary, buoyancy drives magma to travel towards the surface of the Earth, arresting in reservoirs at different levels in the crust for up to tens of thousands of years at a time (Kelley and Barton, 2008; Cruden and Weinberg, 2018; Sparks *et al.*, 2019). An individual magma reservoir can reach hundreds of cubic kilometres in volume and is filled incrementally over time (McNulty *et al.*, 2000; Cruden and McCaffrey, 2001; Glazner *et al.*, 2004; Annen *et al.*, 2015; Blundy and Annen, 2016). During its time in the different reservoirs, the magma composition may evolve via fractional crystallisation and/or crustal assimilation (Bunsen, 1851; Daly, 1925; Tuttle and Bowen, 1958; Carmichael, 1964; Sparks *et al.*, 1977; Patchett, 1980; Huppert and Sparks, 1988; Bindeman and Valley, 2001; Grove and Brown, 2018). The trans-crustal framework of magmatic bodies and feeder channels is referred to as the volcanic or magmatic plumbing system (Burchardt, 2018).

Nowadays, the magmatic plumbing system is viewed as a trans-crustal hot zone mainly comprised of 'crystal mush' (i.e. Edmonds *et al.*, 2019; Fig. 2). In the 'mush' paradigm, magma is stored in a semi-rigid framework of crystals (ca. >50%), and small pockets of almost pure melt (<5 vol.% crystals) may reside in the mush (Hildreth, 2004; Bachmann and Huber, 2016; Cashman *et al.*, 2017; Sparks *et al.*, 2019). The mush is considered uneruptible, unless remobilised by an external trigger, such as magma recharge (i.e Bachmann and Bergantz, 2003; Cooper and Kent, 2014). Eruptions may be fed directly from the mush zone, or melt may be extracted from the mush and reside in a shallow magma reservoir separate from the mush complex before erupting (or not erupting) (i.e. Kennedy *et al.*, 2018; Fig. 2). Notably, some studies have disputed the mush model for Iceland, in favour of a 'stacked' sill model where there are multiple storage zones of melt dominated magma at different crustal levels (Greenfield and White, 2015; Edmonds *et al.*, 2019; White *et al.*, 2019).

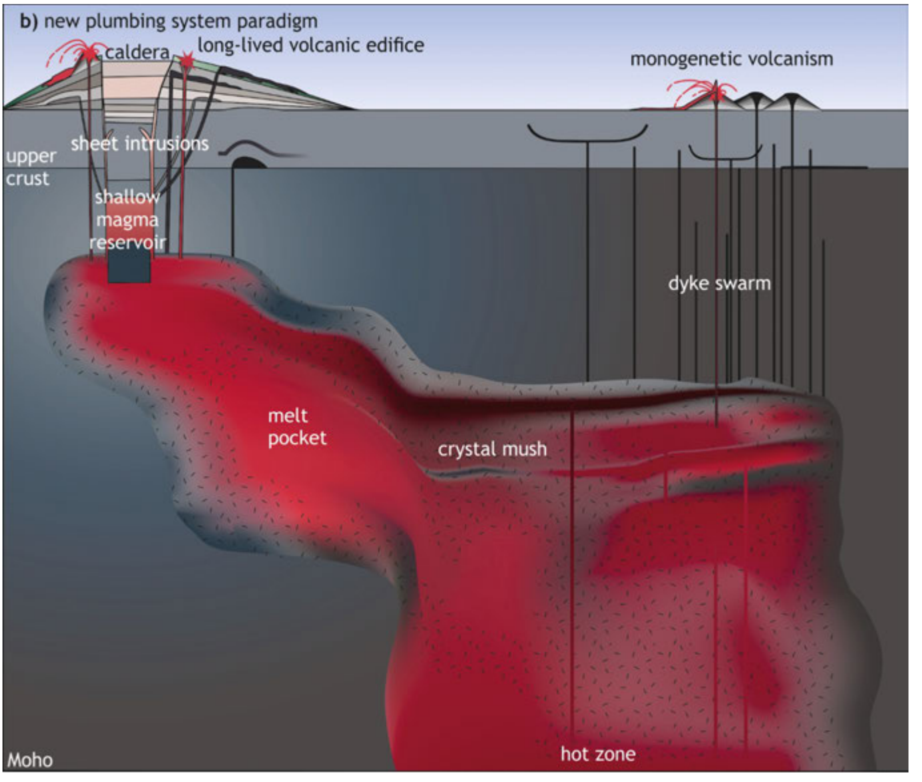

Figure 2: A schematic view of the trans-crustal mush paradigm. Sourced from Burchardt *et al.* (2022).

Whether a stacked sill or a trans-crustal mush model, the creation of space in the crust is required for the formation of magma reservoirs. In the upper crust (i.e. top 5 km), ductile deformation is limited, and two main magma emplacement mechanisms are considered to be viable. The first is the uplift of the host rock above the intruding magma (the magma reservoirs 'roof'), by doming or sliding along faults (Fig 3; Gilbert, 1877; Hawkes and Hawkes, 1933; Morgan *et al.,* 2005). Laccolith emplacement begins with the intrusion of a sill, and growth of the magma reservoir is then facilitated by doming of the overlying host rock. During piston-uplift, the host rock is uplifted along faults, rather than doming. In the doming scenario, the host rock should largely drape the dome-shaped laccolith, producing what is referred to as a 'forced fold' (Gilbert, 1877; Hawkes and Hawkes, 1933). The doming may be accompanied by radial normal faults on the upper part of the dome, and thrusts on the flanks (i.e. Fig 3a; van Wyk de Vries *et al.,* 2014; Magee *et al.,* 2017; Mattsson *et al.,* 2018). In the piston-uplift scenario, the host rock should largely retain its original inclination, and large steeply dipping faults that facilitate uplift should be visible at the edges of the intrusion (Corry, 1988; de Saint-Blanquat *et al.,* 2006; Morgan *et al.,* 2008; Schmiedel *et al.,* 2019). In some scenarios, piston uplift may be a second stage that follows the initial doming (i.e. Mattsson *et al.,* 2018; Burchardt *et al.,* 2019).

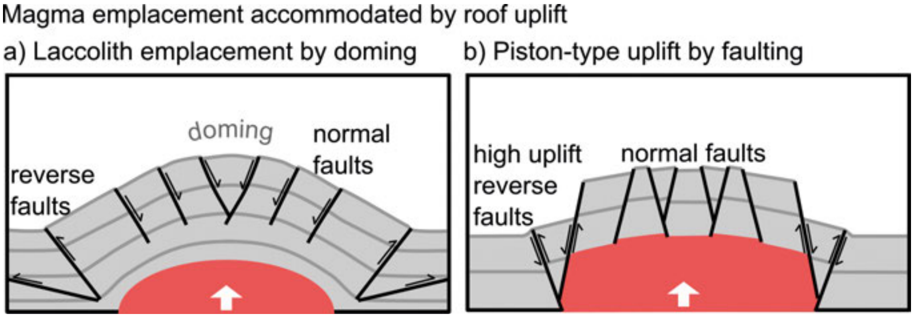

Figure 3: Models for magma emplacement accommodated by roof uplift. Right figure adapted from Schmiedel *et al.* (2019).

The second emplacement mechanism is via subsidence of the rocks below the intruding magma (the magma reservoirs 'floor') into a larger, lower magma reservoir (Fig. 4; Clough *et al.*, 1909; Cloos, 1923; Hamilton and Myers, 1967; Cruden and McCaffrey, 2001; Burchardt *et al.*, 2010). In this mechanism, a mass exchange of magma occurs between the lower and upper reservoirs. The broad term for this emplacement mechanism is floor subsidence, and has been divided into piston, piecemeal, symmetric- and asymmetric-cantilever types, reflecting the geometry of the floor (Tomek *et al.*, 2014; Cruden and Weinberg, 2018). In piston floor subsidence, the floor subsidies as a single horizontal block bound by 'ring' dykes/faults, which facilitate subsidence of the floor block and transfer of magma (Clough *et al.*, 1909; Bonin *et al.*, 2004; Burchardt *et al.*, 2012). Piecemeal floor subsidence is similar to piston subsidence, but occurs via multiple floor blocks bound by faults and dykes (McNulty *et al.*, 2000; Tomek *et al.*, 2014). In contrast, both symmetric and asymmetric cantilever floor subsidence contain a single fault/magma feeding zone, and produce funnel and wedge shaped pluton geometries, respectively (Cruden, 1998; Vigneresse *et al.*, 1999; Cruden and McCaffrey, 2001).

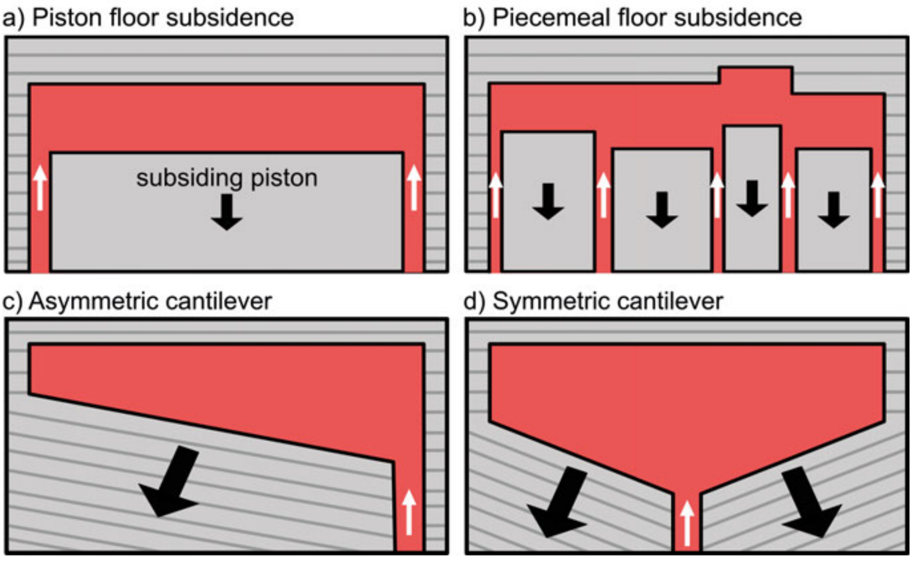

Figure 4: Models for magma reservoir emplacement by floor subsidence.

Tectonic processes associated with local dilation within strike slip faults, reverse faults, normal faults and folds have been also been attributed to contribute to space making for magma bodies (Cruden and Weinberg, 2018). However, it is considered that although the tectonic structures may provide pathways for magma ascent, the faults alone are not sufficient to accommodate the space required for kilometre-sized pluton emplacement (Paterson and Fowler, 1993; Acocella and Rossetti, 2002; Cruden and Weinberg, 2018). Therefore, while tectonics may contribute and influence the overall geometry, emplacement mechanisms such laccolith style uplift or floor subsidence are still required.

Study area

Iceland's highly abundant volcanism has been attributed to the intersection of the Mid Atlantic Ridge (MAR) with the Iceland mantle plume (Thordarson and Larsen, 2007). Volcanism in Iceland is centred along (1) rift zones reflecting the current on-land spreading location of the MAR, which produce magmas of tholeiite compositions, and (2) flank or off-rift zones, which are defined by a lack of rifting and by the production of magmas with alkaline compositions (Fig. 5; Jakobsson, 1972; Jónasson, 2007; Thordarson and Hoskuldsson, 2008). Geophysical and petrological evidence suggests that mafic magma is generated via partial melting of the mantle at the MOHO (Kelley and Barton, 2008; White *et al.*, 2019). From there, the magma may travel to the lower-upper crust boundary (ca. 5 km), and/or a shallow upper crust reservoir (<5 km) before eruption (Tryggvason, 1994; Sturkell *et al.*, 2003; Kelley and Barton, 2008; Flude *et al.*, 2010; Tarasewicz *et al.*, 2012; White *et al.*, 2019). Magmas of silicic composition are generated by both (a) fractional crystallisation, and (b) the partial melting of hydrothermally altered basaltic crust and older silicic volcanic rocks by the intrusion of hot basaltic magma (Carmichael, 1964; Sigurdsson and Sparks, 1981; Macdonald *et al.*, 1987; Jónasson, 2007; Bindeman *et al.*, 2012; Schattel *et al.*, 2014).

The resultant volcanism can be segregated into discrete volcanic systems. A typical volcanic system comprises an elongate fissure swarm, with fissures aligned parallel to the spreading axis, and a central volcano (Sæmundsson, 1979). The central volcano is considered the core hub for volcanic activity, containing a shallow magmatic plumbing system, and importantly, is the main point of eruption for silicic volcanic rocks (Walker, 1966; Sæmundsson, 1979; Thordarson and Larsen, 2007; Thordarson and Hoskuldsson, 2008; Askew *et al.*, 2020).

Insights into the underlying magmatic plumbing systems in Iceland can be found in the fjords of the west and east, where up to 2 km erosion has dissected the crust, exposing the roots of the former rift systems (Walker, 1964). The Reyðarártindur pluton is one of four silicic or mafic-silicic plutons exposed in the Lón fjord of Southeast Iceland, termed the Southeast Iceland Intrusive Suite (SIIS) (Fig. 5). These major intrusions we thought to be coeval (Moorbath *et al.*, 1968; Furman *et al.*, 1992), but dating by Padilla (2015) disproved this when zircon crystallisation ages revealed time differences of millennia between the intrusions. At the time of the SIIS emplacement, the tectonic and geologic setting is regarded akin to the modern-day rift or developing-rift zones (Gale *et al.*, 1966; Moorbath *et al.*, 1968; Furman *et al.*, 1992; Padilla *et al.*, 2016).

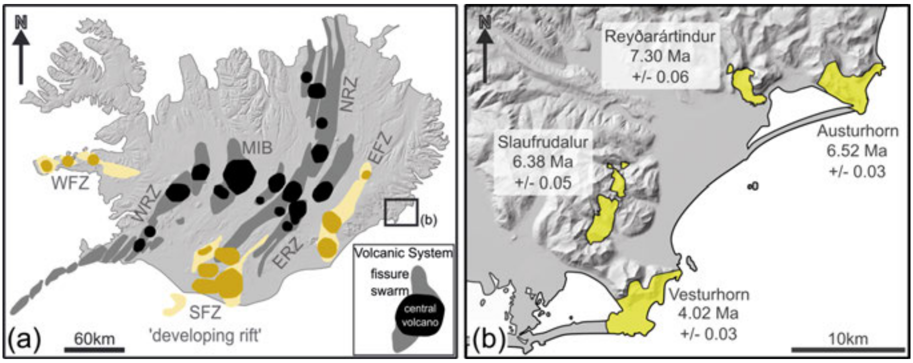

Figure 5: (a) Simplified map of Iceland showing the active volcanic systems (<0.8 Ma) after Jóhannesson & Sæmundsson (2009). A volcanic system is defined as containing a fissure swarm (yellow/grey), a central volcano (black/dark yellow), or both. On this map, the volcanic systems are divided by colour into rift zones (grey/black), and flank zones (yellow/dark yellow). The rift zones are divided into the western (WRZ), eastern (ERZ), northern (NRZ), and mid Icelandic belt (MIB) rift zones. Likewise, the different flank zones are referred to as the western (WFZ), eastern (EFZ) and southern flank zone or 'developing rift' (SFZ) (after Jónasson (2007)). Black box indicates location of (b). **(b)** Map of the Southeast Iceland Intrusive Suite (SIIS). The outline for Slaufrudalur is based on Burchardt *et al.* (2012), for Austurhorn is based on Blake (1966), for Vesturhorn is based on Jóhannesson and Sæmundsson (2014), and Reyðarártindur from the mapping conducted during this study. Ages are from Padilla (2015). Digital elevation models and coastline outline for both maps were downloaded from Landmælingar Íslands (www.lmi.is).

The Reyðarártindur pluton is exposed along the south and eastern sides of Reyðarártindur Peak, its namesake, as well as in the valleys associated with the Reyðará and Karlsá Rivers. The pluton was first mentioned and mapped by Cargill *et al.* (1928), and again by Gale *et al.* (1966) and Walker (1964). Following this mapping, no further investigations were conducted for another 50 years, when Padilla (2015) performed U-Pb dating and measured $\delta^{18}O$ ratios on the zircons, as part of a wider study of the SIIS. The zircon dating produced crystallisation ages of 7.30 ± 0.06 Ma, and $\delta^{18}O$ values that range from +1.5 to +4.8 ‰, reflecting incorporation of hydrothermally altered crust in the parental magma (Padilla, 2015). Additionally, Padilla (2015) reported that pluton was dominantly constructed of granophyre, but contained a zone of mixed and mingled mafic and silicic rocks in the Reyðará valley. Recently, the Reyðarártindur pluton was dated again, this time producing zircon crystallisation ages of 7.40 ± 0.02 to 7.41 ± 0.04 Ma (Twomey *et al.*, 2020), comparable to the Padilla (2015) study. This thesis presents the first in-depth study that is solely focused on the Reyðarártindur pluton that attempts to decipher the emplacement, evolution, and eruption of the magma reservoir.


W3
W5
R12
R44b
E5
R29
R3a
R26b
T5
R16
R7b
T4
R2a
T7
R19c
K10
R21
R28a
R43a
R4b
R4a
K1
T12
R33
R13a
T14b
R19b
R35b
R36
E4b
H5
E6
H12b
R18b
R35a
CZ1

Methods

Many methods were utilised in the three manuscript. Here I outline the main methods that were used

Contact Mapping

The detailed geological mapping of the pluton exposure presented in **Paper I** builds on reconnaissance-level maps by Cargill et al. (1928), Walker (1964), Gale et al. (1966), and Padilla (2015). The pluton was mapped during three field campaigns, the first time in August 2018 and again in June and August 2019. Where possible, contacts were mapped from ground reconnaissance with a hand-held ground positioning system and complemented by mapping on aerial photographs (Bing Maps). Unmanned Aerial Vehicle (UAV "drone") imagery filled in the gaps where access was restricted, and where the aerial photographs are unclear (i.e. due to shadows). Strike and dip measurements of the contact were taken in the field where feasible. All data was imported into the MOVE™ 2019 software (https://www.petex.com/products/move-suite/), and the contact was drawn in 3D view, overlain on a 2 m spatial resolution digital elevation model (Arctic DEM).

Sample collection and preparation

10 samples of the host rock and 175 samples of the pluton were collected for the purpose of thin section preparation and geochemical analysis (**Papers I** and **III**). The sampling aim for the pluton rock units (**Paper I**) was to produce an even and representative sampling distribution, with a higher sampling density in the Reyðará River area where there was a wide range of lithologies. For the host rock units (**Paper III**), the aim was to collect representative samples and a profile extending upwards from the pluton-host rock contact. The minimum sample size collected was 10 cm^3, excluding the weathered surface. Thin sections slabs were cut at Uppsala University, and then sent for thin section preparation at the Slovak Academy of Sciences. Samples for geochemical analysis were cut at Uppsala University and cleaned via scrubbing in water with a brush and soaking in an ultrasonic bath. At least 100 g of each sample

was then crushed in a steel jaw crusher and milled to a fine-grained powder in a tungsten carbide mill at the Natural History Museum (Naturhistoriska Riksmuseet), Stockholm. Approximately 20 g of each sample was sent to ALS geochemistry in Vancouver, Canada where major and trace element geochemistry was performed.

3D reconstruction of the pluton shape

A 3D reconstruction of the pluton shape was created to estimate the volume of the magma reservoir (**Paper I**), and analyse the roof shape with respect to the magma emplacement (**Paper II**). The MOVE^{TM} 2019 software (https://www.petex.com/products/move-suite/) software was used for 3D reconstruction of the Reyðarártindur pluton. 46 cross sections of the roof oriented N–S and 22 oriented NW–SE (300–120°) were drawn and the roof surface was created from these using Delaunay Triangulation (Fig. 6). Further information on the construction of 3D reconstruction and a detailed list of inferences is provided in **Paper I**.

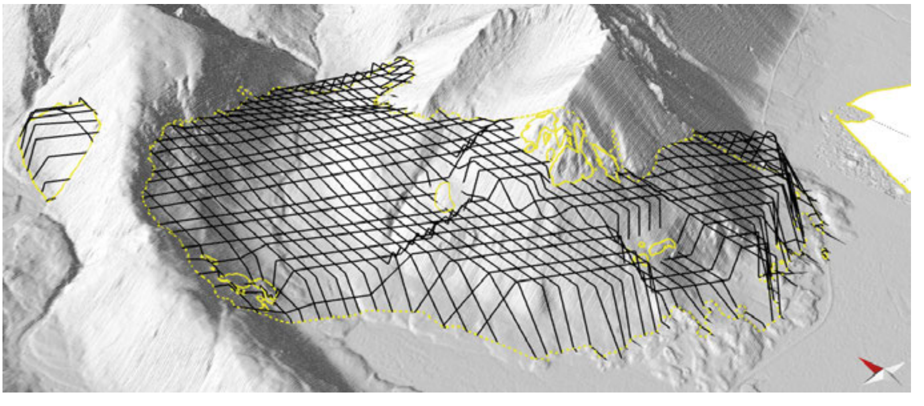

Figure 6: Map of the cross section lines used in the 3D reconstruction of the Reyðarártindur pluton.

Mineral identification and chemistry

Mineral identification was required for the pluton rock units in **Paper I**, and the partial melt and host rock units in **Paper III**. Initially, optical microscopy was employed to identify the minerals and textures in the rocks. Additionally, on selected thin sections, spot Wavelength Dispersive X-ray Spectroscopy (WDS) analysis and the production of elemental maps via WDS were conducted (Fig. 7). In **Paper I**, the purpose of spot analysis and elemental maps was to obtain feldspar zoning profiles, whereas in **Paper III**, the focus was to assist in the identification of many of the minerals. I conducted all of the spot analyses myself on the field emission source JEOL JXA-8530F Hyperprobe Scanning Electron Microscope (SEM) at the Department of Earth Sciences, Uppsala University. For the elemental maps presented in **Paper I**, I collected the data on the same SEM. However, for **Paper III**, the elemental maps were produced using Energy Dispersive Spectroscopy (EDS) by Johan Lissenberg at the School of Earth and Environmental Sciences, Cardiff University, Wales. For run times and analytical conditions, see the respective papers.

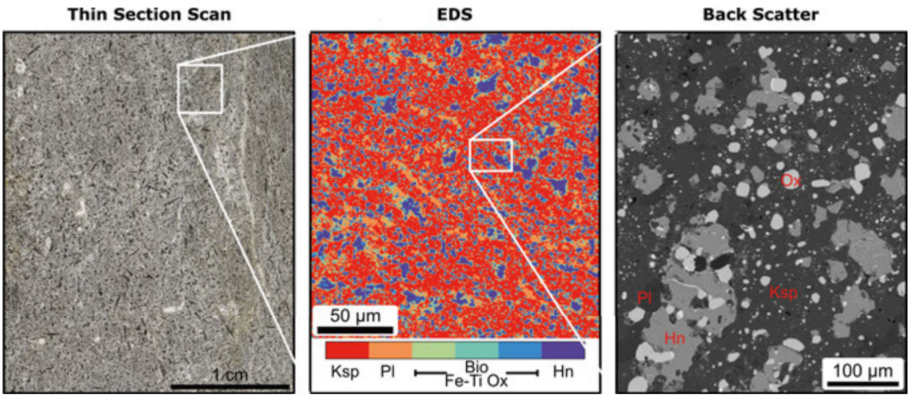

Figure 7: Example of a thin section, elemental scan, and backscatter image

Photogrammetry and virtual outcrop analysis

Structural orientation data of faults, dykes, lava beds and fractures were collected for **Paper II** for the purpose of structural analysis. The structural measurements were collected using two different methods. The first was directly in-field using (a) the FieldMove Clino Pro application (www.mve.com/digital-mapping/), and (b) by analogue compasses accompanied by a handheld GPS. The second was by digital mapping of structural orientation data from virtual outcrops (Fig. 8). Digital mapping has the advantage of being able to take numerous measurements from zones that would otherwise be inaccessible. To create virtual outcrops, overlapping photos of the outcrops were acquired using an UAV, and then imported into the Agisoft Photoscan™ software for processing. The resulting .obj file was then imported into the LIME v2.0 software (https://virtualoutcrop.com/lime/; Buckley *et al.* 2019), where the 'structural data from 3 points' tool was used to acquire the orientation of measurable faults, dykes and lava bedding. After data acquisition, all the measurements were collated in the Petroleum Experts MOVE v2019.1 software, where stereonets and rose plots were constructed.

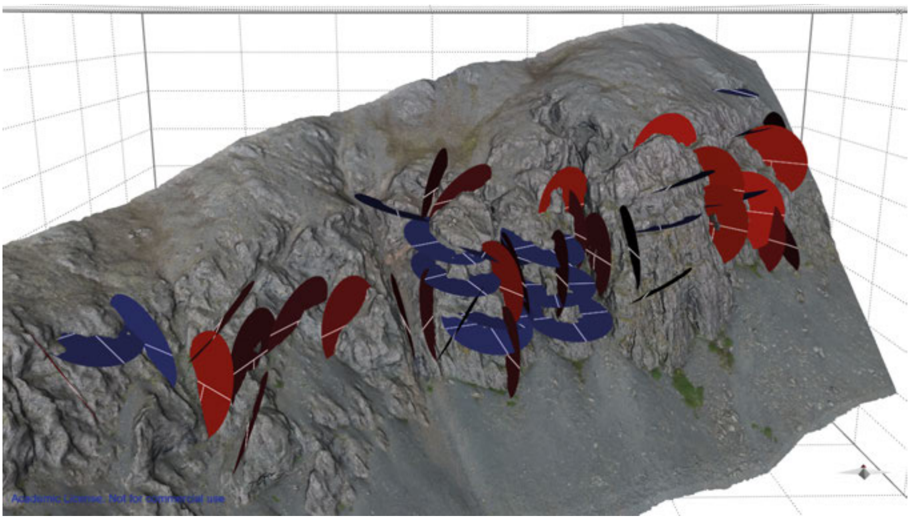

Figure 8: Example of a virtual outcrop. Blue disks represent measured orientations of the lava flows, and red disks the measured orientation of faults, fractures and dykes.

Additional methods used in the manuscripts

The papers in this thesis also employ data collection methods that were undertaken by commercial labs and co-authors. Specifically, finite element modelling in COMSOL was conducted by co-authors Christoph Hieronymous and Sonja Greiner for **Papers I** and **II**; whole rock geochemical analysis was conducted for **Papers I** and **III** via the commercial services of ALS Geochemistry, Vancouver; EDS scans were obtained for **Paper III** by Johan Lissenberg at Cardiff University, Wales; and porosity and tensile strength measurements were obtained for **Paper III** by Michael Heap at the University of Strasbourg, France. These methods are described in full in the respective manuscripts.

Summary of Papers

Paper I: Rapid assembly and eruption of a shallow silicic magma reservoir, Reyðarártindur pluton, Southeast Iceland.

Here, we use field mapping, 3D pluton reconstruction, geochemistry and a thermal model to (a) present the first modern detailed map of the Reyðarártindur pluton, and (b) investigate the assembly of magma and eruptive history of the Reyðarártindur pluton.

Mapping results showed that the host rock to the Reyðarártindur pluton primarily consists of Neogene basalt lava flows, and in a small zone in the northwest, some intercalated rhyolite lavas from the Lón Volcanic Complex (Fig. 9). The roof basalts to the Reyðarártindur pluton are mostly sub-horizontal (±10°), and the roof contacts are sharp and usually concordant with the roof basalts. Adjacent roof exposures occur with vertical offsets of up to 200 m, creating both structural highs and lows. The floor of the pluton is not exposed, and the lowest exposure of the pluton roof is near sea level. The long axis (NW–SW) of the pluton is 4.72 km long and the short axis (SW–NE) is 1.65 km wide, giving a surface area of ca. 8 km^2 in map view. In 3D, the pluton has a shape characterized by flat roof segments that are vertically offset, giving the pluton a 'castle-like' shape. A minimum volume of 2.5 km^3 was calculated from the 3D reconstruction.

Primarily, the exposed pluton is constructed of a single rock unit, the Main Granite (69.9–77.7 wt.% SiO_2). The Main Granite is macroscopically homogenous feldspar-phyric granophyric microgranite. We interpret that the Main Granite was intruded first in the intrusion sequence. Two further units are locally exposed as enclaves mingling with the Main Granite at the base of the exposure, the Granite Enclaves (67.4–70.2 wt.% SiO_2), and the Quartz Monzonite Enclaves (61.8–67.3 wt.% SiO_2) (Fig. 9). These units are distinguished from the Main Granite both by texture and whole-rock geochemistry. We interpret that these two units were intruded after the Main Granite. Locally, mixing between the Enclaves and the Main Granite occurred, which produced the geochemically distinct Enclave Host Granite (66.7–71.3 wt.% SiO_2).

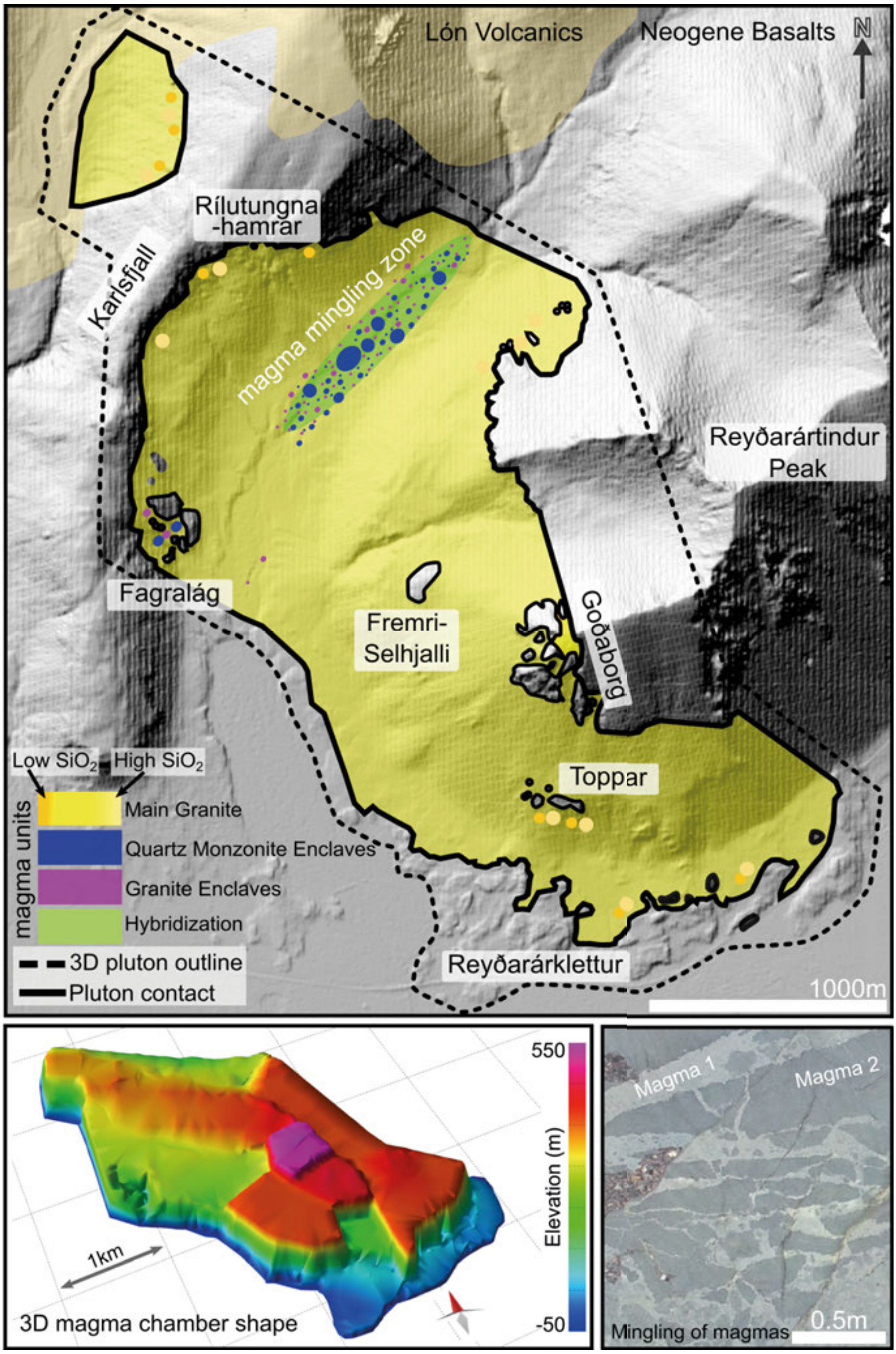

Figure 9: Top: The new map of the Reyðarártindur pluton. **Bottom left:** 3D reconstruction of the pluton. **Bottom right**: Photo of enclaves in the magma mingling zone exposed in the Reyðará River.

Geochemically, the suite of rocks from the pluton (including the Main Granite, Enclave Host Granite and Enclaves) plot together on Harker Diagrams. The coherence of the trends indicates the intrusive rock units are geochemically related, likely via differentiation by fractional crystallisation of amphibole, feldspar(s), and oxides. The implication of this result is that the units are all likely sourced from the same magma reservoir. Feldspar phenocrysts in the units display disequilibrium textures at their rims, and the cores document growth in a more primitive melt than their rims. Therefore, we consider that the growth of the feldspar phenocrysts was not co-genetic with the host melt, and that they were removed from a mush and entrained within a host melt of slightly different composition during transport to the Reyðarártindur reservoir.

The pluton roof is intruded by dykes from the pluton, and in two locations displays depressions associated with large dykes: Rílutungnahamrar and Fagralág (Fig. 10). Within these particular dykes, the rock is partially to wholly tuffisitic, and rock compositions range from quartz monzonite to granite. We interpret these zones as eruption-feeding conduits from the pluton. The compositional range of the conduit samples, corresponding to the different magma types observed in the pluton, suggest that the silicic magma injection events responsible for the magma mingling preserved in the pluton could have been an eruption trigger.

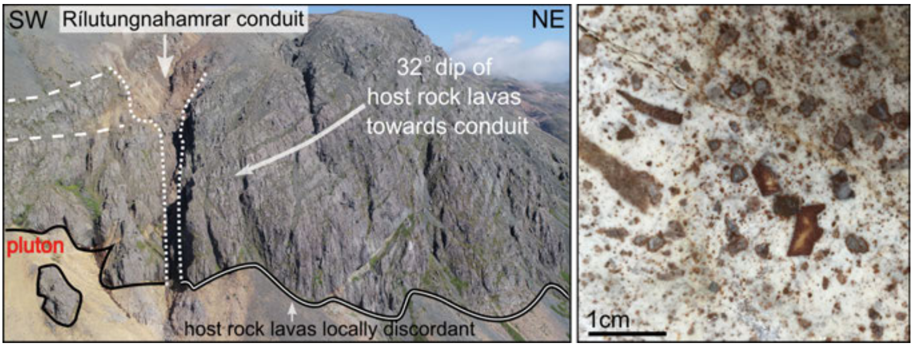

Figure 10: Left: Photo of the Rílutungnahamrar conduit, which shows the host rock lavas dipping towards the dyke axis. **Right:** Photo of the pyroclastic breccia exposed within the conduit.

To arrive at a rough quantitative estimate of possible pluton longevity and eruption timeframes, we ran two thermal cooling models, one with the exposed thickness of the pluton (600 m) and one with double that thickness because the full thickness of the pluton is not exposed. In the first scenario, the total melt volume of 2.5 km^3 is halved by 1000 years, and decreases to zero by 3000 years. In the second scenario, the cooling timescale increases about three-fold. Notably, in both models, the top 75 m of the pluton would have reached rheological lockup ca. 1000 years after formation. Therefore, we can reason that the eruption timeframe of the pluton was limited to 1000 years, as there is no preserved evidence of later magmas cross-cutting older, cooler magmas leading to the conduits.

In conclusion, this study of the Reyðarártindur pluton suggests that silicic magma batches may migrate from their source mush reservoir and assemble in the shallow crust prior to eruption. The timescales from establishment to eruption may be geologically short (i.e. less than 1000 years) and the longevity of the reservoir limited to approx. 5000 years. Intrusion of new silicic magma into the reservoir may be sufficient to trigger an eruption, as there was no trace of basaltic magma in the conduits. Overall, the Reyðarártindur pluton is likely a smaller, ephemeral part of a wider plumbing system that feeds a central volcano.

Paper II: Volcanic unrest as seen from the magmatic source — Reyðarártindur pluton, Iceland

Here, we use the Reyðarártindur pluton as a case study to explore how space was created for the formation of a magma reservoir of silicic magma. Secondly, we investigate the deformation of the host rock associated with emplacement and eruption, and the conditions that led to eruption. To do this, we use field mapping and photogrammetry to analyse the orientation of the lava layers, and fracture, fault and dyke structures in the basaltic host rock to the pluton. We then combine this information with the 3D pluton shape reconstruction from **Paper I** to present an emplacement model.

Measurements of lava attitudes distributed over the entire pluton roof reveals that the host lavas above the pluton are generally sub-horizontal (0–12˚), and concordant with the pluton roof (Fig. 11). The lava attitudes are comparable with the regional lava attitudes measured outside of the pluton walls. Sparse outcrops and the contact trace suggest that the pluton wall contacts are sub-vertical and discordant to the host lavas. The orientation of the lava beds at the wall contacts are also sub-horizontal. Two dip anomalies of the lava layers in the pluton roof were identified. The first dip anomaly occurs at Reyðarárklettur where the lavas locally dip ca. 22˚ to the south. The second dip anomaly is associated with a long dyke/conduit that extends the width of the pluton, the Rílutungnahamrar-Falkáhnaus dyke. On the eastern side of the dyke, the lavas locally dip 20–30˚ towards the dyke axis. On the western side of the dyke, no dip anomaly occurs. As the dyke was established as a conduit in **Paper I**, we consider that the local subsidence was associated with eruption.

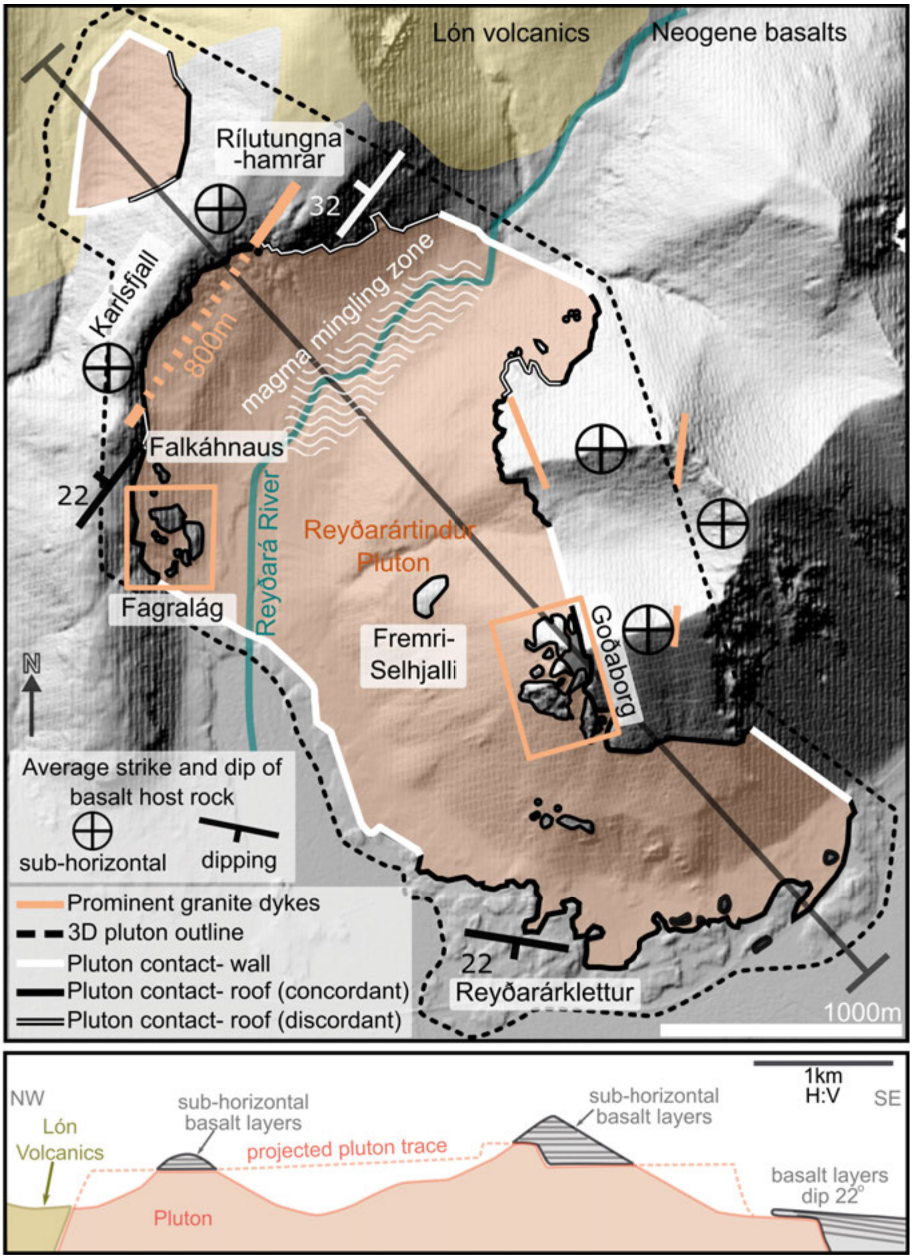

Figure 11: Upper: Map of the Reyðarártindur pluton, which contains average strike and dip values of the basaltic host rock. **Lower:** Cross-section NW–SW through the pluton which shows the 'stepped' roof structure.

In contrast to basalts away from the pluton, the basaltic lavas overlying the pluton roof are fractured, faulted and intruded by granitic dykes that originate from the pluton. The fractures are steeply dipping to sub-vertical (70–90°), and many are arranged in conjugate sets (Fig. 12). Some of the fractures exhibit displacement of the lava layers of up to 5 m (i.e. a fault), and slickenlines observed in the field show sub-vertical displacement along the fault planes. Dykes intrude some of the fractures.

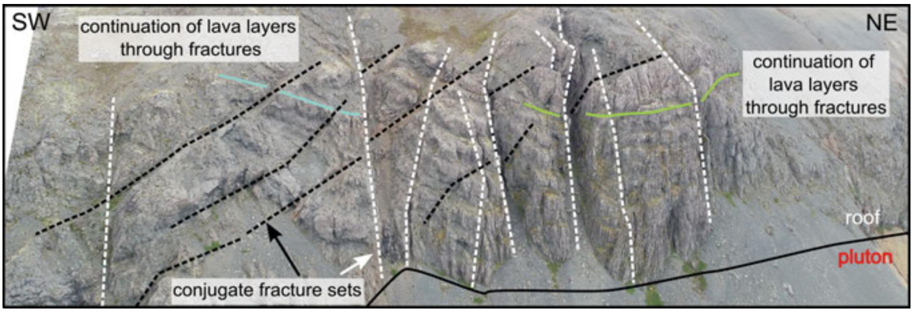

Figure 12: The conjugate fracture sets penetrating the host rock above the pluton.

We consider that the pluton was emplaced by piecemeal floor subsidence which began with multiple horizontal intrusions of the Main Granite at different localities at different depths, hence the stepped pluton roof (**Paper I**) (Fig. 13). Our main evidence for this is the sub-horizontal layering of the lavas above the pluton, which is traceable across faults and beyond the pluton walls. Continued magma supply would have promoted the thickening and subsequent merging of individual intrusions by the subsidence of multiple blocks of the intrusion floor.

The conjugate fracturing/faulting in the roof suggest upwards pressure from the magma reservoir and testify to overpressure. We consider that the fracture sets were not created during initial pluton emplacement, but rather during later magma recharge. Overpressure eventually culminated in eruption, evidenced by the exposed conduits. The mingling of magmatic units within the conduits indicates that injection of new magma into the reservoir triggered the overpressure/eruption. We envisage that cooling and sealing of the piecemeal subsidence network preceded eruption, causing overpressure with magma recharge. Hence, the sequence of events deciphered from Reyðarártindur shows that eruption would have been preceded by recharge into the pluton and small-scale seismicity in the pluton roof, which would likely be detectable via seismic monitoring networks (Fig. 13).

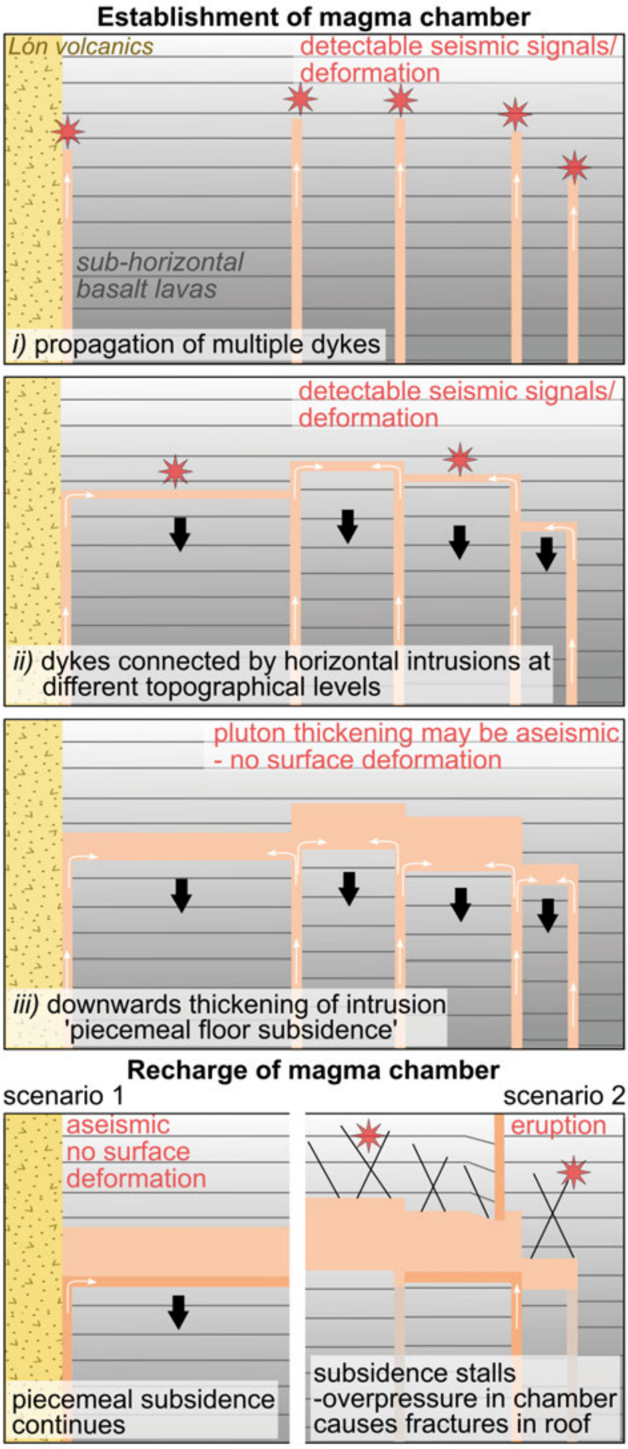

Figure 13: Model for the emplacement to eruption of the Reyðarártindur pluton that explains the observed 3D shape, wall and roof rock contacts, and fracture sets.

In order to quantify what, if any, deformation would be expected at the Earth's surface during eruption, we constructed a simple numerical model that replicates the field observations of subsidence towards the Rílutungnahamrar-Falkáhnaus conduit. The results of the model show that the Rílutungnahamrar-Falkáhnaus eruption would have been observed along a rift-parallel fissure located in a half-graben, linked to the asymmetric subsidence of the reservoir roof. According to our numerical model, the surface deformation pattern is highly asymmetric and concentrates above the collapsing part of the roof, not the dyke. The 100 m subsidence at the magma reservoir roof only partially translates to the surface, where the maximum surface displacement is ca. 15% of the maximum subsidence at 2 km depth. Subsidence of the block may have released significant seismic energy, especially if the subsidence occurred en-masse.

In conclusion, we show that piecemeal floor subsidence was the main mechanism that facilitated the emplacement of the magma reservoir. Recharge led to overpressure in the magma reservoir, which caused small-scale faulting and fracturing of the pluton roof, and eventually lead to an eruption. We envisage that cooling and sealing of the piecemeal subsidence network preceded eruption, causing overpressure with magma recharge. The eruption at the Rílutungnahamrar-Falkáhnaus conduit was associated with trapdoor-style roof subsidence, which would have been observable at the Earth's surface and may have released significant seismicity.

Paper III: Cap-rock formation above a magma reservoir, Reyðarártindur pluton, Iceland.

Magma bodies can physically and chemically react with their host rocks to form what is referred to as a 'cap-rock' above a magma body (i.e. Hayba and Ingebritsen 1997). The properties of the cap-rock influence many processes, including (a) the cooling rate of the magma body, and therefore the longevity of the magma reservoir, (b) the transport of heat, fluids and gasses away from the magma body, and consequently the development of magmatic-hydrothermal systems, and (c) the overpressure conditions required for the injection of dykes, and thus, eruption. In this study, we investigate the formation of a cap-rock in the in the host rock directly above the Reyðarártindur pluton. To do this, we use field mapping, sampling, compositional analyses, geothermobarometry, rock strength testing and partial-melting modelling. With this information, we create a model for dynamic cap-rock formation. We then consider the effects of the cap-rock properties and structures on the failure of the cap-rock and on heat transfer between the magma and the host rock.

The roof contact typically has a sharp interface between the granitic magma (**Paper I**) and the former basalt-lava host rock (now hornfels). The host rock was shattered and altered for ca. 10 m above the pluton contact (Fig. 14). Evidence of shattering is preserved in the irregular veins of granitic material preserved between hornfels clasts. We consider that the shattering occurred during horizontal magma intrusion at the beginnings of magma reservoir development. Following shattering, sintering and recrystallisation (contact metamorphism) changed the basalt to a sanidinite hornfels with a characteristic black and shiny appearance. Some partial melting of the host rock occurred, and melt pooled locally at the magma-host rock contact.

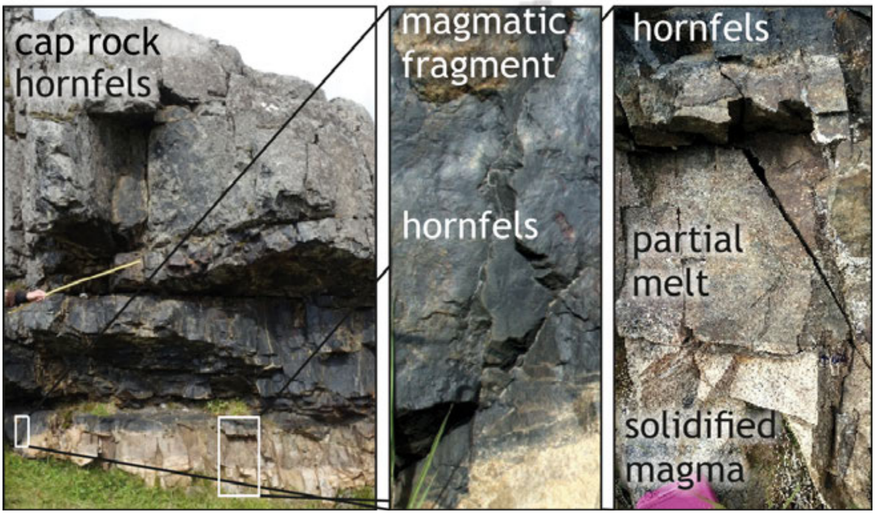

Figure 14: Field photos of the cap-rock. This zone shows the hornfels and the accumulation of partial melt at the pluton-host rock interface.

The mineral assemblage and composition of the hornfels indicate temperatures exceeding 800˚C (Fig. 15). The hornfels is extremely hard and impermeable (density 3157.2 kg×m^{-3}; tensile strength 39.5 MPa; connected porosity 0.00). We also observe later stage fractures and faults that cut through the cap-rock, the arrangement and orientation of which demonstrates that they were formed by magmatic overpressure in the reservoir (**Paper II**). The fractures also give us a means to "date" the formation of the cap-rock, showing that it must have formed early, following the establishment of the magma body. Due to the unusually strong cap-rock, dykes from the magma reservoir could only intrude along the fractures, instead of creating their own pathways (**Paper II**). Still, a few of the dykes fed surface eruptions (**Paper I**). A model for cap-rock formation is then presented (Fig. 16).

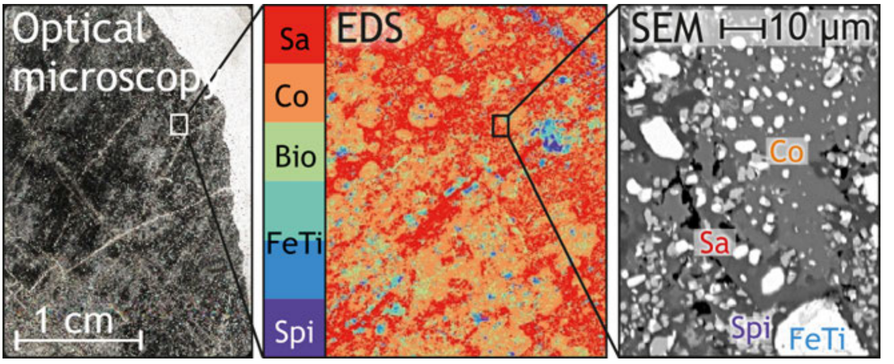

Figure 15: Results from optical microscopy, EDS and SEM compositional analysis. The former basalt now has a completely metamorphic mineral assemblage. Sa= sanidine, Co= cordierite, Bio= biotite, FeTi= Fe Ti oxides, Spi= hercynite spinel.

Our findings show that the contact metamorphism occurring at the magma-host rock interface can cause changes in the physical, chemical and mechanical properties of the host rock. The cap-rock then acts as a 'lid' to the magma body, insulating the magma, and increasing the longevity of eruptible magma within the magma body. Additionally, due to a higher strength, larger over-pressure may be required for dyke formation and eruption. Heat loss from the magma body would have initially been restricted to conductive heat transfer. Convective transfer of fluids, gasses and heat would have developed later, once faults, fractures and dykes penetrated the cap-rock. This case study highlights the importance of processes occurring at the magma-host rock interface, and the subsequent implications for the thermal longevity of the magma body and the development of the associated magmatic-hydrothermal system.

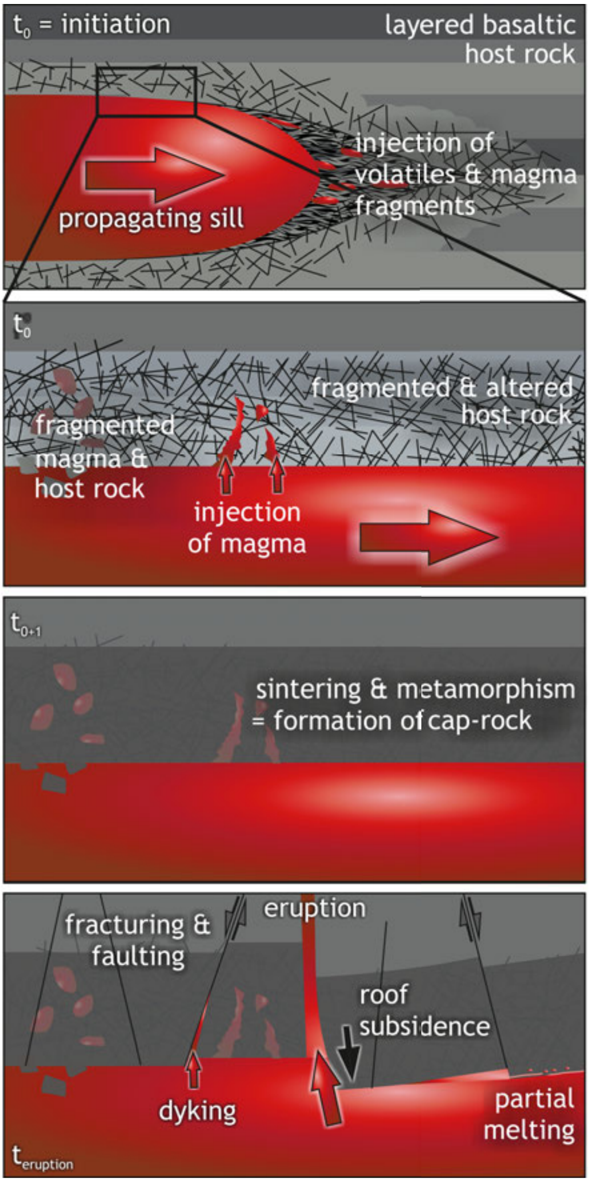

Figure 16: Model of the cap-rock development above the Reyðarártindur pluton. At t_0, horizontal intrusion of magma shatters a damage zone approximately 10 m thick in the surrounding basaltic host rock. Pyroclastic granite (tuffisite) from the sill intrudes the fractures. Simultaneously, local metasomatism of the rocks closest to the contact occurs, promoted by available fluids and gasses from the sill. Following sill emplacement t_{0-1}, contact metamorphism recrystallizes the host rock, some host rock is partially melted, and sintering of the clasts in the shatter zone occurs to create a coherent, pore and fracture free cap-rock. During a later stage associated with magma reservoir recharge ($t_{eruption}$) the host rock is fractured, faulted and eruption occurs.

Conclusions and Outlook

Here I unite the models from **Papers I–III** to present a full model for the evolution of the Reyðarártindur magma reservoir. In the last section, I discuss what the findings mean in a wider scientific context.

Model for the evolution of the Reyðarártindur pluton

A melt-dominated magma body of silicic composition with a minimum volume of 2.5 km^3 was emplaced in the upper crust during the Neogene at approximately 1–2 km depth. The tectonic setting was likely a rift zone or an establishing rift zone, which is consistent with findings from Gale *et al.* (1966), Moorbath *et al.* (1968), Furman *et al.* (1992) and Padilla *et al.* (2016). The model for evolution of the magma reservoir is as follows:

I. Batches of the Main Granite were extracted from an underlying mush reservoir and propagated upwards via a network of dykes. The magma extracted was predominantly melt, containing only ca. 5 vol.% crystals, which were entrained during melt migration (**Paper I**; Fig. 17 i).

II. The dykes arrested at different depths where the magma then propagates laterally, parallel to the host rock lava layers, connecting the dyke network. The connection of the dyke networks allowed multiple blocks of the reservoir floor to subside downwards into the underlying magma reservoir, so-called piecemeal floor subsidence. Consequently, the resulting pluton has a 'stepped' roof. During piecemeal subsidence, the Reyðarártindur reservoir grew by mass exchange of magma between the lower source reservoir and the upper Reyðarártindur reservoir (**Paper II**; Fig. 17 ii–iii).

III. Simultaneous with horizontal magma intrusion, the host rock at the top of the magma reservoir was locally shattered, intruded by the sill, sintered, and underwent contact metamorphism to sanidinite facies. The resulting roof rock was stronger than the surrounding basalts, pore and fracture free, and impermeable. It acted as a cap-rock to the magma reservoir, restricting the loss of magmatic fluids and gasses, and limiting heat loss to conductive heat loss only (**Paper III**; Fig. 17 ii–iii).

IV. At a later stage, recharge of the magma reservoir occurred via batches of a further granite and a quartz monzonite magma. Although geochemically distinct, the units were all extracted from the same source reservoir as the Main Granite (**Paper I**). Cooling and sealing of the faults and magma pathways between the magma source and the Reyðarártindur magma reservoir inhibited continued subsidence of the reservoir floor. Therefore, magma recharge caused overpressure in the magma reservoir, fracturing and faulting the reservoir roof (**Paper II**; Fig. 17 iv).

V. The fractures and faults were used as preferential pathways by dykes, and allowed for the establishment of convective heat, fluid and gas transport from the magma reservoir. Convective heat transport and loss of volatile phases promoted faster cooling and solidification of the magma reservoir (**Paper III**; Fig. 17 iv).

VI. Once the overpressure associated with recharge was sufficient to allow dyke propagation all the way to the Earth's surface, eruption occurred. Evacuation of magma through conduits caused localised subsidence of the overlying host rock. This subsidence would have been visible at the Earth's surface (**Paper II**; Fig. 17 iv).

The possible eruptive timeframe was likely limited to 1000 years after initial magma reservoir establishment, although local pools of melt could have existed in the core or base of the reservoir for up to 4000 years (**Paper I**).

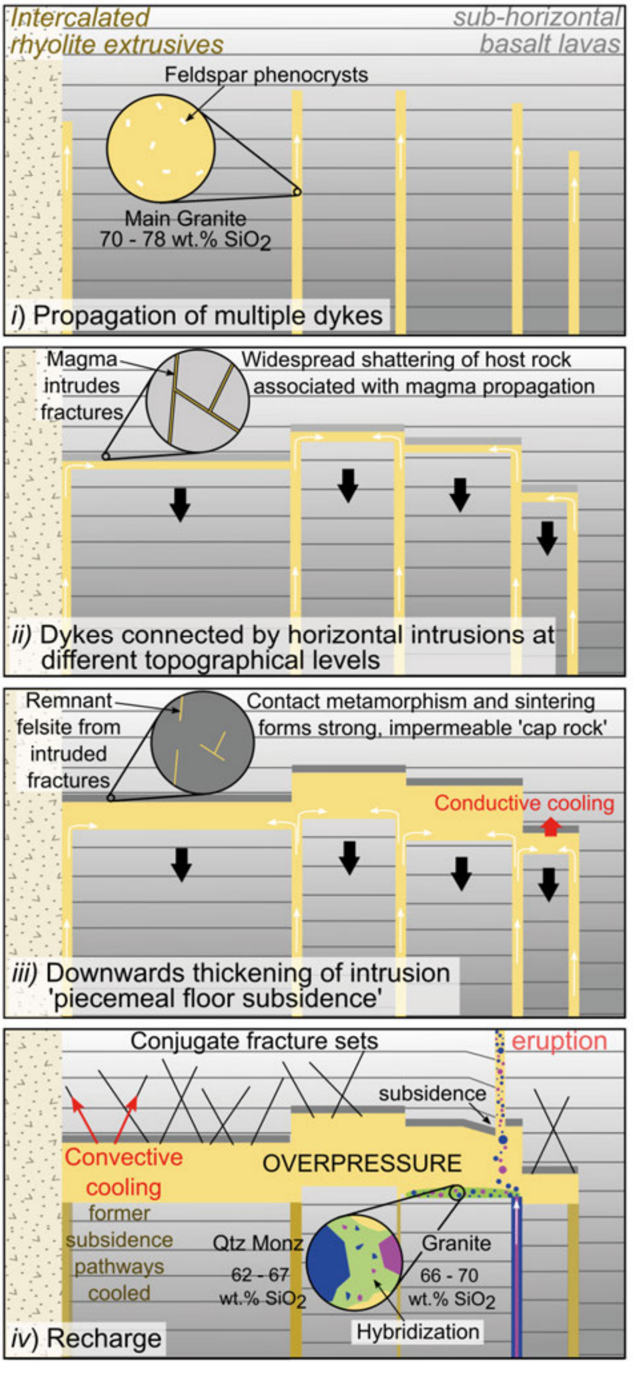

Figure 17: Conceptual model for the evolution of the Reyðarártindur magma reservoir.

Reyðarártindur in a wider scientific context

What insights do the studies give on the wider plumbing system?

What does the source magma system to Reyðarártindur look like?

For floor subsidence to occur, the source magma reservoir must have been directly below the Reyðarártindur magma reservoir. Within the source reservoir, we know that differentiation by fractional crystallisation was the process responsible for the geochemical separation of the magmas producing the Main Granite, Granite Enclave and Quartz Monzonite rock units. In **Paper I**, we consider that the separate evolution could have occurred within three different scenarios: (1) within a stratified source reservoir, (2) discrete melt lenses within a mush zone, or (3) by extraction of melt at different times from a melt-mush reservoir (cf. Miller and Miller, 2002; Bachmann and Bergantz, 2004; Pistone *et al.*, 2017; Edmonds *et al.*, 2019). Mush formation has been proposed under volcanoes of all compositions and settings in Iceland (Gunnarsson *et al.*, 1998; Hansen and Grönvold, 2000; Alfaro *et al.*, 2007; Jónasson, 2007; Flude *et al.*, 2010; Chekol *et al.*, 2011; Passmore *et al.*, 2012). Likewise, there have also been arguments for melt-dominated reservoirs at all depths in Iceland (Greenfield and White, 2015; Edmonds *et al.*, 2019; White *et al.*, 2019; Saubin *et al.*, 2021). A consequence of differentiation by fractional crystallisation is that there must be a remaining mush or crystal graveyard that the melt has been either extracted from, or the crystals have settled to.

To get a full picture of the system, we must consider how silicic melts are generated in Iceland. Generation of silicic magmas in Iceland has been attributed to three different processes: (a) fractional crystallisation, (b) the partial melting of hydrothermally altered basaltic crust and older silicic volcanic rocks by the intrusion of hot mafic magma, and (c) a combination of both, so called assimilation and fractional crystallisation (Carmichael, 1964; Sigurdsson and Sparks, 1981; Macdonald *et al.*, 1987; Jónasson, 2007; Bindeman *et al.*, 2012; Schattel *et al.*, 2014). The fractional crystallisation model is often preferred for flank zone systems, and the partial melting model for rift zone systems (e.g. Schattel *et al.*, 2014). In the partial melting models, the silicic magma generated is often separate from the mafic magma body that provided the heat (i.e. Walker, 1966; Gunnarsson *et al.*, 1998; Árnason, 2020; Saubin *et al.*, 2021). Padilla (2015) found that the oxygen isotope ranges of Reyðarártindur zircons suggest considerable incorporation of hydrothermally altered crust into their parental magmas prior to crystallisation. Based on our own mapping we can add that the altered silicic volcanic rocks of the Lón volcanic complex provided ample and close-by material to melt. Thus, a system where partial melting of the crust as well as differentiation by fractional crystallisation was occurring to generate the melt series.

Another piece of the puzzle are the feldspar phenocrysts, which were grown in a melt more primitive than, and out of equilibrium with, the Main Granite, Quartz Monzonite and Granite Enclave melts that gave rise to the rocks they now reside in. In **Paper I**, we hypothesised that during melt migration the plagioclase phenocrysts were removed from a mush, and entrained within the melt, explaining their disequilibrium. Additionally, the phenocrysts do not display oscillatory zoned cores, which in other studies of mush systems have been attributed to repeated recharge and melt flux through the mush system via a more mafic magma (i.e. Ginibre *et al.*, 2002; Neave *et al.*, 2013).

With all the above information, I speculate that the source reservoir for Reyðarártindur was most likely a mush/melt-mush system that was formed above/associated with a larger mafic reservoir. The original intermediate composition melt was generated by partial melting of the surrounding hydrothermally altered host rock (cf. Padilla, 2015), which was assimilated into the mafic magma. Fractional crystallisation occurred, generating a more mush-like reservoir and melts of quartz monzonite to granite composition. The reservoir may have been separate from the main mafic reservoir, and was therefore not affected by small recharge events and flux of new melt through the system, which would have caused oscillatory zoning in the phenocrysts. Extraction of melt at different times/locations within the reservoir generated the series of magmatic rock units exposed at Reyðarártindur. Plagioclase phenocrysts were entrained from other parts of the source reservoir during melt migration.

What role did Reyðarártindur play within the plumbing system?

In **Paper I**, we conducted finite element modelling and determined that the eruptible lifetime of the Reyðarártindur magma reservoir was significantly less than that of an Icelandic central volcano. Therefore, we concluded that the reservoir was likely an ephemeral part of the wider plumbing system that fed the associated central volcano. An implication is that one might anticipate the presence of several different short-lived shallow felsic magma reservoirs associated with longer-lived central volcanoes. This implication couples well with geochemical studies of erupted deposits from central volcanoes and the results from the IDDP-1 drilling program at Krafla volcano.

For example, at Krafla, the solidified host rock the IDDP-1 drilling program encountered has been termed a “felsite”, and has been interpreted as solidified remnants of previous intrusion, which the new melt has intruded (Elders *et al.*, 2011; Zierenberg *et al.*, 2013; Saubin *et al.*, 2021). Additionally, Rooyakkers *et al.* (2021) established different geochemical signatures for different rhyolite eruptions there, and implied that these different geochemical signatures represented different granitic magma bodies. Similarly, Árnason (2020) proposed a model where there are multiple cycles of ascending silicic melts that stagnate at shallow levels, and may be “ignited” by basalt to erupt.

Likewise, at Kerlingarfjöll, Flude *et al.* (2010) established three different erupted rhyolite lineages, which were attributed to different episodic rhyolite sources. At Thingmuli, Charreteur *et al.* (2013) proposed a model including multiple separated silicic intrusions in the upper crust. Similarly, at the eroded central volcanoes of Reyðarfjordor and Breiðdalur, different silicic sequences/linages have been identified (Gibson, 1963; Askew *et al.*, 2020).

When does a magma reservoir grow vs when does it erupt?

A dyke will propagate from a magma reservoir when the overpressure within the reservoir exceeds the critical value for failure, and cracks propagate into the wall rock, initiating magma transport (McLeod and Tait, 1999; Rivalta *et al.*, 2015). Eruption will occur if the magma within the dyke contains sufficient energy to reach the Earth's surface (McLeod and Tait, 1999; Cañón-Tapia, 2014; Caricchi *et al.*, 2021). Hence, dyke propagation and eruption hinge on the build-up of overpressure within the magma reservoir. Overpressure can be created by a variation in volume, either by the injection of new magma, or volatile exsolution (i.e. Cañón-Tapia, 2014; Caricchi *et al.*, 2021). In **Papers I** and **II**, we argued that the injection of new magma was the most likely scenario at Reyðarártindur. Prior to eruption, 2.5 km^3 of space for the reservoir was established via piecemeal floor subsidence and mass transfer with the underlying source reservoir. However, this process must not have been operating at the time of recharge and overpressure build up. The reasons could be that (1) the floor subsidence faults/magma pathways had cooled and sealed, and could no longer accommodate subsidence of the blocks, or alternatively (2) corresponding recharge had also occurred in the source reservoir, and hence there was no space or under-pressure for roof subsidence.

In **Paper II**, we consider that cooling and sealing of the magma pathways/faults between the floor subsidence blocks occurred in the time between magma reservoir emplacement and recharge. This observation is a bit contradictory to our observations and modelling that the Reyðarártindur reservoir must have erupted soon (within cooling timeframes) after emplacement (**Paper I**). On the one hand, the Reyðarártindur magma reservoir was a larger body of hot magma, and therefore would have had a longer cooling timeframe compared to the feeder network, but on the other hand, the magma pathways were deeper in the crust, and therefore had a smaller thermal contrast between the magma and the host rock. Perhaps recharge and overpressure was also a factor in the source reservoir. Alternatively, the answer could lie in the formation of the cap-rock at the top of the magma reservoir, insulating the top of the magma body for longer than expected timeframes. A secondary implication of the magma body erupting sooner rather than later after emplacement is that is supports the theory that younger magma reservoirs are more likely to erupt (Caricchi et al., 2021). In conclusion, eruption will occur if the space making mechanism by which the magma body was emplaced by in the first place, in this case, piecemeal floor subsidence, fails to accommodate the incoming magma.

What does Reyðarártindur tell us about the cooling of magma reservoirs?

The development of the high-strength, impermeable cap-rock indicates that shallow magma reservoirs may not cool as fast as we have traditionally thought (**Paper III**). Furthermore, in the quest to reach supercritical fluids, the IDDP-1 program at Krafla volcano, Iceland, accidentally penetrated a magma-host rock interface and intersected a melt-dominated magma body of granitic composition at 2 km depth (Mortensen *et al.*, 2010; Elders *et al.*, 2011). It has been proposed that this magma reservoir could have been emplaced in the crust as long as 300 years ago (Rooyakkers *et al.*, 2021). The finding of melt-dominated magma at such a shallow crustal level, which could be as old as 300 years, is in complete contradiction to our models of cooling and crystallisation at the roof interface, and storage of magma in a dense crystalline 'mush' (Cashman *et al.*, 2017; Eichelberger, 2020). Perhaps the development of a cap-rock such as documented above the Reyðarártindur pluton in **Paper III** could help explain the melt-dominated state of the magma body at Krafla. Additionally, the floor subsidence emplacement mechanism means that there was a larger (likely mafic dominated) source reservoir below the Reyðarártindur magma reservoir, which could have provided heat to the upper reservoir and/or a higher geothermal gradient. The implication that shallow magma reservoirs cool slower than traditionally thought is positive news for the longevity of magmatic-geothermal systems, and hence, exploration for geothermal energy.

How do the results improve the interpretation of volcanic unrest and hazard mitigation?

Volcanoes are monitored primarily by seismicity, ground deformation, and gas measurements (Sparks, 2003). Volcano scientists in charge of monitoring volcanoes apply models based on the knowledge of magmatic plumbing systems to interpret the signals and provide scenarios for what could occur within the magmatic system (Sparks, 2003; National Academies of Sciences Engineering and Medicine, 2017). This study provides a scenario for the creation, recharge, and eruption of a magma reservoir, and **Paper II** demonstrates what seismic and deformation signals would have been associated with the different phases. The results have particular relevance to central volcanoes in the active rift zones of Iceland. Specifically, kilometre-scale melt-dominated magma bodies of silicic composition may accumulate in the upper crust prior to eruption (**Paper I**). The eruption may occur within geologically short timescales after emplacement (<1000 years). Therefore, the shallow emplacement of magma should be viewed as a warning that an eruption could initiate.

The Reyðarártindur magma reservoir was emplaced by floor subsidence, and although there was some deformation of the lava flows at the roof, the measurable uplift signals would have been minor to negligible (and localised to a zone in the south; **Paper II**). Calculations of intruding magma volume from the uplift signals would have dramatically underestimated the volume of the intruding magma. Hence, volcano monitoring personnel should consider any volumes estimated as absolute minimums. Furthermore, once magma reservoirs were established, growth of the magma reservoir may be aseismic and cause no surface deformation. As magma recharge is considered a primary eruption trigger, the recharge of a magma reservoir without detectable signals is concerning. However, prior to eruption at the Reyðarártindur, there would have been detectable seismicity in the reservoir roof, as overpressure in the chamber led to roof fracturing and faulting across the entire reservoir roof.

Magma chambers are often modelled as a Mogi source (Masterlark, 2007; Sigmundsson *et al.*, 2018; Taylor *et al.*, 2021). However, this study shows that in reality, magma reservoirs have complex 3D shapes (**Paper I**). The elongated shape is not unique to Reyðarártindur, as the much larger neighbouring Slaufrudalur pluton also has a long axis significantly wider than the short axis, although Slaufrudalur has a relatively flat roof (Burchardt *et al.*, 2010). Hence, modellers should consider different geometrical shapes of magma reservoirs during deformation modelling. Furthermore, the formation of a strong, impermeable cap-rock above the reservoir (**Paper III**) suggests that future studies should use high values of tensile strength for the host rock in their calculations and models.

An additional implication of the cap-rock formation is that gas and fluids would have been trapped in the magma reservoir until it was fractured during the overpressure phase (**Paper III**). This means that there may not have been any significant volcanic gas emissions at the Earth's surface after cap-rock formation. However, after fracturing there may have been a release of gas and strong volcanic gas emissions, although focussed to areas of high fracture density, such as above the reservoir "corners".

Implications for the geological evolution of Southeast Iceland

This thesis shows for the first time the age relationship between the (undated) Lón volcanic complex and Reyðarártindur. The Lón volcanic complex must be older than the Reyðarártindur magma reservoir, as it forms the host rock to the pluton in the northwest (**Paper I**).

It should also be noted that there are uncertainties in terms of intrusion depth. The burial depths in Eastern Iceland were established by Walker (1960) and are based on zones of regional zeolite-facies metamorphism, dyke density and the increase in dip of the lava flows as a function of depth. The Lón fjord of Southeast Iceland is part of the deepest exposed level with a calculated burial depth of 2 km (Walker, 1960; Blake, 1966). However, we do not actually know when the maximum burial occurred in relation to the intrusion of Reyðarártindur. Based on the granophyric textures of the pluton rock units, the emplacement depth must be deeper than the Sandfell Laccolith, which was emplaced at ca. 500 m and shows very fine grained, almost volcanic textures (Hawkes and Hawkes, 1933; Mattsson *et al.*, 2018). Volcanic textures in the rocks of the conduits in the roof of Reyðarártindur are likely related to decompression during eruption and cannot be used to estimate the depth of the pluton roof (Hugh Tuffen, pers. comm.). Hence, we deem it likely that Reyðarártindur emplaced somewhere between 1–2 km depth.

Other studies focusing on the volume, isotopes and chemistry of Icelandic lavas over time have revealed pulses of magmatic activity with a high flux coincidental with the zircon crystallisation age of Reyðarártindur, around 8–7 Ma (Watkins and Walker, 1977; McDougall *et al.*, 1984; Hanan and Schilling, 1997a; Kitagawa *et al.*, 2008). The pulsating magmatic flux over time in Iceland has been attributed to changes in source mantle endmember components (Zindler *et al.*, 1979; Hanan and Schilling, 1997b; Thirlwall *et al.*, 2004; Kokfelt, 2006; Kitagawa *et al.*, 2008). In particular, the heightened 8 to 7 Ma flux has been attributed to an enrichment of recycled crustal materials in the mantle source (Kitagawa *et al.*, 2008). This 8 to 7 Ma time period is also coincidental with the initiation of the modern Northern Rift Zone from the Snæfellsnes Rift Zone by an eastward rift jump (6 to 7 Ma Saemundsson *et al.*, 1980; or 8.5 Ma Garcia *et al.*, 2003). Rift jumps occur to relocate the plate boundary over the mantle plume centre (Sæmundsson, 1979; Jóhannesson, 1980; Hardarson *et al.*, 1997). When a rift jump occurs, the centre of magmatism, and heat input, reestablished in a different, older section of crust. This has been hypothesized to cause an abundance of silicic volcanism from the remelting of old volcanic centres and silicic segregations in the crust (Gunnarsson *et al.*, 1998). Notably, the location of Reyðarártindur partially overlaps with the outermost flank of the older Lón central volcano (Fig. 9). Lón contains an abundance of both felsic and mafic rocks that may have provided hydrothermally altered material that could have been recycled into the deeper source reservoir of Reyðarártindur. Therefore, the combination of a high magmatic input and abundant available material for recycling in the crust likely contributed to the development of the Reyðarártindur pluton and associated central volcano.

Concluding statement

The silicic intrusions of Southeast Iceland are still a poorly studied area, despite them being a window into the volcanic and magmatic processes that occur in the roots of the active volcanoes of Iceland. The work on the Reyðarártindur pluton conducted in this thesis helps fill a significant gap in understanding the processes behind the accumulation and eruption of silicic magma in the Icelandic upper crust. In particular, the work contributes to models of magma emplacement, interpretation of volcanic unrest, eruption triggers and the development of geothermal systems. Furthermore, this study is particularly complementary to the ongoing work and models of emplacement for the shallow melt-dominated magma body discovered during geothermal drilling at Krafla. Future work should focus on establishing the relationship between the Lón volcanic complex and the Reyðarártindur pluton, the source reservoir to the Reyðarártindur pluton, and the permeability network developed in the cap-rock.

Summary in Swedish

Forskare som övervakar aktiva vulkaner är intresserade av hur magma transporteras genom jordskorpan till ett vulkanutbrott. På Island brukar magma förflytta sig från djupt belägna magmakammare där magman produceras till de övre 5 km i jordskorpan där magma lagras innan ett möjligt vulkanutbrott. Forskare kan inte direkt observera aktiva magmakammare. Därför undersöker vi gamla, stelnade ”fossila” magmakammare som numera ligger vid jordytan för att förstå de processer som pågick under deras bildning och vad som kan ha orsakat vulkanutbrott från dessa magmakammare. Våra forskningsresultat kan förbättra utbrottsprognoser som baseras på tolkningar av vulkaniska jordskalv och förändringar i vulkanens yta som en följd av processerna på djupet. Denna avhandling undersöker Reyðarártindur, en fossil magmakammare på Island som var aktiv för 7 miljoner år sedan, och rekonstruerar dess historia från bildning till utbrott. Detta är den första vetenskapliga studien av Reyðarártindur vilket krävde en grundläggande kartering av magmakammaren och omkringliggande berg.

Studierna i denna avhandling etablerade att magmakammaren har haft en borgliknande form med en volym på minst 2,5 km^3 och att den fylldes snabbt. De olika magmatyper som fanns i kammaren har hög kiselhalt. Magma med hög kiselhalt kan orsaka explosiva vulkanutbrott. Till exempel var det ökända Eyjafjallajökullutbrott 2010 som påverkade flygtrafiken över hela norra halvklotet ett utbrott av kiselrik magma.

Reyðarártindur bildades snabbt genom en mekanism som kallas för golvsättning. Det innebär att magman trängde in i ett utrymme som skapades genom att berget i botten av den nya magmakammaren sjönk ner i den djupare magmakammare varifrån magman nu trängde upp. Det var i den djupare magmakammaren som magman hade bildats och utvecklats innan Reyðarártindur kom till. När den nya magmakammaren sedan bildades krossades berget i kammarens tak. De höga temperaturerna och gaser från magman gjorde att det krossade berget omvandlades och svetsades ihop till något som kan liknas ett lock. Detta lock var starkare än det ursprungliga berget och isolerade magman. Dessutom förhindrade locket att gaser från magman kunde frisättas, och till en början vulkanutbrott. Istället byggdes ett högt tryck upp i magmakammaren.

Till slut var trycket så högt att magma från Reyðarártindur lämnade kammaren i ett eller flera vulkanutbrott från vilket visar sig i resterna av gamla utbrottskanaler där magmakammarens tak (locket) har brustit. Resterna av magman i dessa utbrottskanaler påvisar typiska tecken på ett explosivt utbrott likande Eyjafjallajökull 2010. Anledningen till utbrottet var att ny magma trängde in i magmakammaren. Vanligtvis visar utbrottsprodukter på Island, och speciellt från Eyjafjallajökull, att den nya magman som triggade utbrottet var basaltisk, dvs. med låg kiselhalt, hög temperatur och högt gasinnehåll. Sådan magma förmår mobilisera den trögflytande kiselrika och explosiva magma som annars gärna sitter kvar i kammaren och stelnar. Resultaten av studierna i denna avhandling visar istället att den nya magman som trängde in i Reyðarártindur hade hög kiselhalt. Anledningen till utbrottet kan också härledas till att golvsättningen fungerade otillräckligt bra för att kompensera för den tillkommande magman och att kammarens tak (locket) motverkade en tryckutjämning tills att det till slut brast.

Trycket i kammaren skapade även ett flertal mindre sprickor i taket innan utbrottet, och dessa skulle kunna ha uppmätts med hjälp av seismometrar om vulkanen övervakats för 7 miljoner år sedan. Under utbrottet bildades det en sänka vid vulkanens yta som troligen fylldes med aska och pyroklastiskt material från utbrottet.

Simuleringar av hur magmakammaren svalnade och stelnade visar att kammaren bildades och hade utbrott under en tidsperiod på mindre än 1000 år, vilket är väldigt kort. Vulkaner på Island kan vara aktiva i ungefär 1 miljon år, vilket innebär att Reyðarártindur var troligtvis enbart en kortvarig del av ett större system av magmakammare i den vulkanens liv.

Denna studie av en magmakammares anatomi, utveckling och utbrott kommer att hjälpa de som övervakar aktiva vulkaner att tolka signal som uppstår i vulkanens rötter.

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Like a volcano at the Earth's surface, I consider this work to be the tip of a very large magma system that is hidden from sight. Many people need to be acknowledged for its success. Furthermore, this thesis was conducted during an unexpected pandemic and therefore everyone's support is doubly acknowledged.

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