Volcanotectonic Evolution and Characteristic Volcanism of the Neovolcanic Zone of Iceland

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Abstract

The thesis focuses on three aspects of the volcanotectonic activity in the Neovolcanic Zone of Iceland: (1) effects of deglaciation, (2) mechanical interaction between central volcanoes, and (3) effects of volcanoes as soft, elastic inclusions on the propagation of volcanic fissures and rift zones. The Neovolcanic Zone contains rocks belonging to the Brunhes normal magnetic epoch, dating back to 0.78 Ma, and represents the on-land expression of the Mid-Atlantic Ridge. This zone, composed of three segments, is the location of most of the volcanotectonic activity in Iceland. Within the Neovolcanic Zone, the Holocene volcanism is primarily confined to the volcanic systems, essentially large groups or (within the rift zone) swarms of volcanic and tectonic features. Most volcanic systems contain a central volcano, many of which have a collapse caldera, and (within the rift zone) a fissure swarm. The volcanic systems are fairly evenly distributed throughout the Neovolcanic Zone. In addition to the polygenic central volcanoes, the Neovolcanic Zone contains numerous monogenic basalt volcanoes. These include table mountains and hyaloclastite ridges, formed in subglacial eruptions, as well as lava shields and volcanic fissures, formed in subaerial eruptions.

The retreat of the ice at the close of the Weichselian, and the associated unloading of the crust and isostatic uplift, has long been held accountable for the increase in volcanic activity in the late glacial and early postglacial periods. Here I present conceptual and numerical models to explain the formation and location of subglacial table mountains and hyaloclastite ridges as, as well as subaerial lava shields and central volcanoes, within the Neovolcanic Zone as a result of the ice retreat. The ice retreat reduced compression in the crust, and created tensile stresses around the deep-seated magma reservoirs, thereby explaining how, where, and when the table mountains and lava shields formed. During the Weichselian, the ice load generated compression that encouraged the development and expansion of shallow crustal magma chamber. Subsequently, the unloading encouraged increased volcanic activity

Numerical models show that there is strong mechanical interaction between central volcanoes within clusters, such as in the central part of Iceland. This interaction encourages shared dykes and seismogenic faulting between nearby volcanoes, as is supported by observations. Central and subglacial volcanoes function as soft, elastic inclusions in the stiffer host-rock (basaltic) lava pile. Stress-modelling results indicate two main points. First, that soft subglacial mountains may hinder the propagation of volcanic fissures, as is supported by models of the 1783 Laki Volcanic Fissure. Second, that large central volcanoes may temporarily arrest the propagation of large parts of or entire rift-zone segment, as is supported by models of the Torfajökull Volcano in South Iceland.

Diese Arbeit beschäftigt sich mit drei Aspekten der vulkanotektonischen Aktivität der Neovulkanischen Zone Islands: (1) den Auswirkungen der Deglaziation, (2) der mechanischen Wechselwirkung zwischen Zentralvulkanen und (3) dem Einfluss von Vulkanen als weiche, elastische Einschlüsse auf die Ausbreitung von vulkanischen Spalten und Riftzonen. Die Neovulkanische Zone Islands enthält Gesteine der Brunhes normalmagnetischen Epoche, jünger als 0,78 Millionen Jahre, und repräsentiert den Mittelatlantischen Rücken an Land. In dieser Zone, die aus drei Segmenten aufgebaut wird, findet der Großteil der vulkanischen Aktivität auf Island statt. Innerhalb der

Neovulkanischen Zone ist der holozäne Vulkanismus vor allem auf die Vulkansysteme beschränkt, die im wesentlichen große Gruppen oder (innerhalb der Riftzone) Schwärme von vulkanischen und tektonischen Strukturen darstellen. Die meisten Vulkansysteme beinhalten einen Zentralvulkan, von denen viele eine Kollapscaldera aufweisen sowie (innerhalb der Riftzone) einen Spaltenschwarm. Die Vulkansysteme sind recht gleichmäßig über die Neovulkanische Zone verteilt. Zusätzlich zu den polygenen Zentralvulkanen enthält die Neovulkanische Zone zahlreiche monogene Basaltvulkane. Diese umfassen Tafelberge und Hyaloklastitrücken, die in subglazialen Eruptionen gebildet wurden, sowie Lavaschilde und vulkanische Spalten, die in subaerischen Eruptionen gebildet wurden.

Der Rückzug des Eises am Ende der Weichseleiszeit, und die damit verbundene Entlastung der Kruste und der isostatische Aufstieg werden seit langem für die Zunahme der vulkanischen Aktivität im spätglazialen und frühen postglazialen Zeitraum verantwortlich gemacht. Hier präsentiere ich konzeptuelle und numerische Modelle, die die Bildung und Verteilung subglazialer Tafelberge und Hyaloklastitrücken sowie subaerischer Lavaschilde und Zentralvulkane innerhalb der Neovulkanischen Zone als Ergebnis des Eisrückzuges erklären. Der Rückzug des Eises reduzierte die Kompression in der Kruste und rief Zugspannungen in der Umgebung der tiefen Magmenreservoire hervor, wodurch erklärt wird, wie, wo und wann sich die Tafelberge und Lavaschilde bildeten. Während des Weichselglazials führte die Belastung durch das Eis zu Kompression, die die Entwicklung und Ausdehnung flacher krustaler Magmenkammern begünstigte. In der Folge wurde durch die Entlastung die Steigerung der vulkanischen Aktivität hervorgerufen.

Numerische Modelle zeigen, dass es eine starke mechanische Wechselwirkung zwischen Zentralvulkanen in Clustern, wie sie im Zentralteil Islands vorkommen, gibt. Diese Wechselwirkung begünstigt Gänge, die in mehreren Vulkanen aufdringen sowie seismogene Störungsaktivität zwischen benachbarten Vulkanen, wie auch Beobachtungen bestätigen. Zentralvulkane und subglaziale Vulkane fungieren dabei als weiche, elastische Einschlüsse innerhalb der steiferen Nebengesteine des (basaltischen) Lavastapels. Ergebnisse von Spannungsmodellierungen weisen auf zwei Hauptpunkte hin. Erstens, dass die weichen subglazialen Berge die Ausbreitung vulkanischer Spalten behindern, wie Modelle der Laki-Vulkanspalte von 1783 zeigen. Zweitens, dass große Zentralvulkane temporär die Ausbreitung ganzer Riftzonensegmente stoppen können wie durch Modelle des Torfajökull-Vulkans in Südisland bestätigt wird.

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1 Introduction

Due to its location on the Mid Atlantic Ridge, and above a mantle plume, Iceland experiences abnormally high levels of volcanic activity. This fact contributes to the countries title of “Land of Fire and Ice”, along with the presence of several ice bodies. The high and extremely varied amount of volcanic activity offers the opportunity to study aspects of volcanology which would possibly be less accessible elsewhere. This is the reason for the location of the work presented here in this study.

This study looks at the types of characteristic volcanic activity in Iceland, and considers their different settings. The study aims to examine the reasons for the locations of the different types of activity in terms of their volcanotectonic setting and geodynamics. The topics studied include the effect of glaciation on the crust and resultant volcanic activity; mechanical interaction between clusters of central volcanoes; the volcanotectonic evolution of the rift zone in the southeast of Iceland; and finally, the presence of large fissure eruptions.

Any study of volcanology takes science a step closer towards understanding and thus protection against the hazards volcanoes pose. Therefore, this study also discusses the types of volcanic hazard in Iceland, and the ways in which this study enhances knowledge of them.

This study is a collection of internationally peer-reviewed journal articles. At the time of writing, two of these were published (Andrew & Gudmundsson, 2007 *Appendix I*; Gudmundsson & Andrew, 2007, *Appendix II*); one in press (Gudmundsson *et al*., 2008, *Appendix IV*); and finally one accepted with minor revisions (Andrew & Gudmundsson, *accepted*, *Appendix III*). In order not to breach copyright laws, the papers have been reformatted to be inserted into this study as appendices. Further work around each of the specific study areas is presented in individual chapters, along with the abstracts of the papers for continuity purposes.

2 The Geodynamic Setting of Iceland

2.1 Mid Atlantic Ridge

The island of Iceland sits astride the Mid-Atlantic Ridge (MAR) in the North Atlantic Ocean, between Greenland and Norway from 63^{o}23'N to 66^{o}30'N. It is the only surface expression of the MAR, whose length is 14,000-15,000 km (Figure 2.1). The MAR is the divergent plate boundary of the Eurasian-African and North and South American plates, and is spreading at an average rate of approximately 2 cm per year.

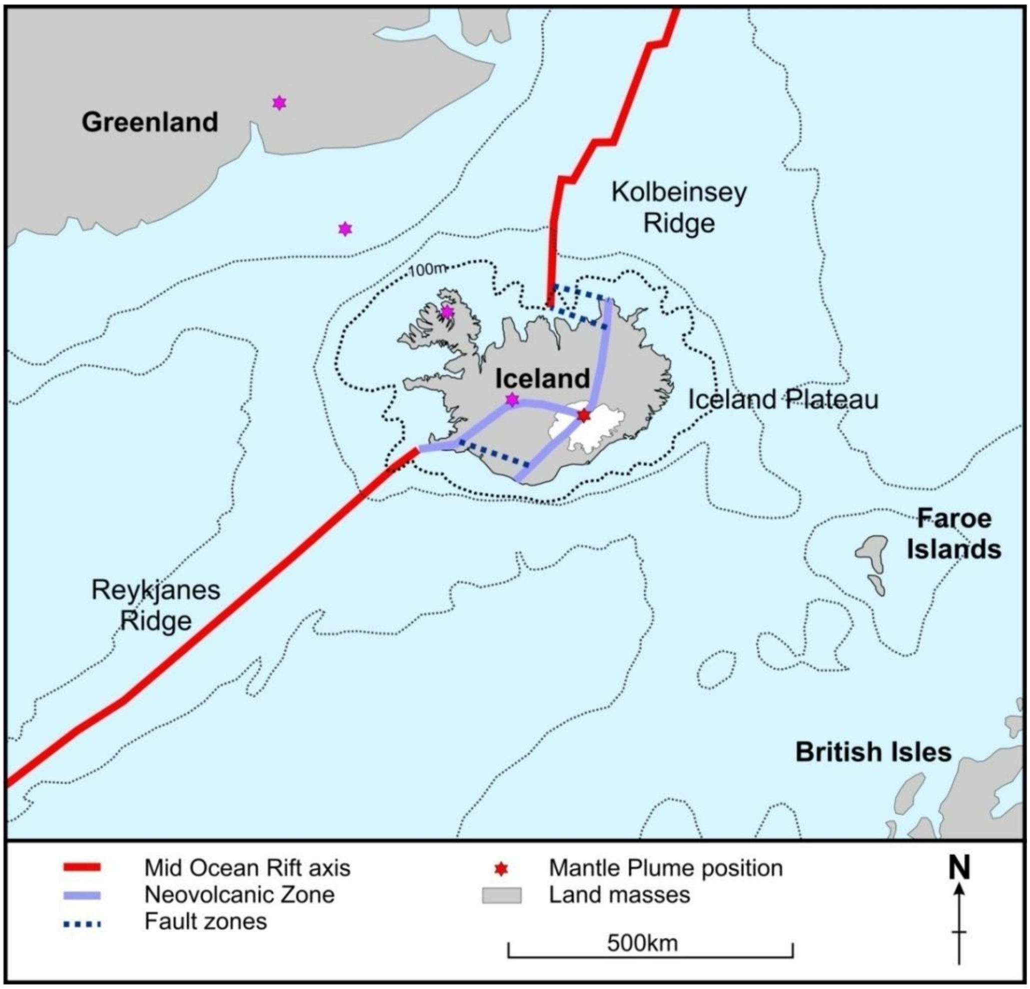

Figure 2.1: Map locating Iceland in the North Atlantic Ocean, at the junction of the Reykjanes and Kolbeinsey Ridges. The Iceland Basalt Plateau is marked by the darker dotted line around Iceland, and the present location of the mantle plume is marked by the red star. The purple stars show the previous locations of the mantle plume, during its creation of the North Atlantic igneous province, their approximate ages coming progressively closer to Iceland from Greenland are: 24 Ma, 15 Ma, 7 Ma and 3 Ma (Saunders, *et al*., 1997). Modified after Tronnes, 2004; Thordarson & Larsen, 2007; Gudmundsson, 2007.

More specifically Iceland is located at the junction of the Reykjanes Ridge to the south and the Kolbeinsey Ridge in the north (Figure 2.1). Iceland is part of the so called Iceland Basalt Plateau (Figure 2.1), an elevated section of oceanic crust forming the floor of the Atlantic Ocean. Iceland comprises 103,000 km^2, which amounts to ~30% of this plateau, whose overall area covers 350,000 km^2 (Thordarson & Hoskuldsson, 2002). The plateau, Iceland and the surround are part of the North Atlantic igneous province, an area developed during the continental breakup in the early Tertiary. This province formed as eastern North America and Greenland drifted away from northwestern Europe, following the initial rise the Iceland mantle plume (Best, 2003).

2.1.1 Mantle Plume

Iceland is located atop one of the hot spots that occur sporadically through the world's tectonic plates. The hot spot is recognisable here, not only by seismic tomography (e.g. Nataf, 2000; Zhao, 2001; Foulger, 2002), but also a magma production rate an order of magnitude higher than in other submarine oceanic rifts. This implies an unusually large volume of underlying hot decompressing mantle, that is, a mantle plume (Best, 2003). The upper 600-700 km of the mantle plume are clearly revealed by the speed decrease of both P- and S-seismic waves (Gudmundsson, 2007) (Figure 2.2).

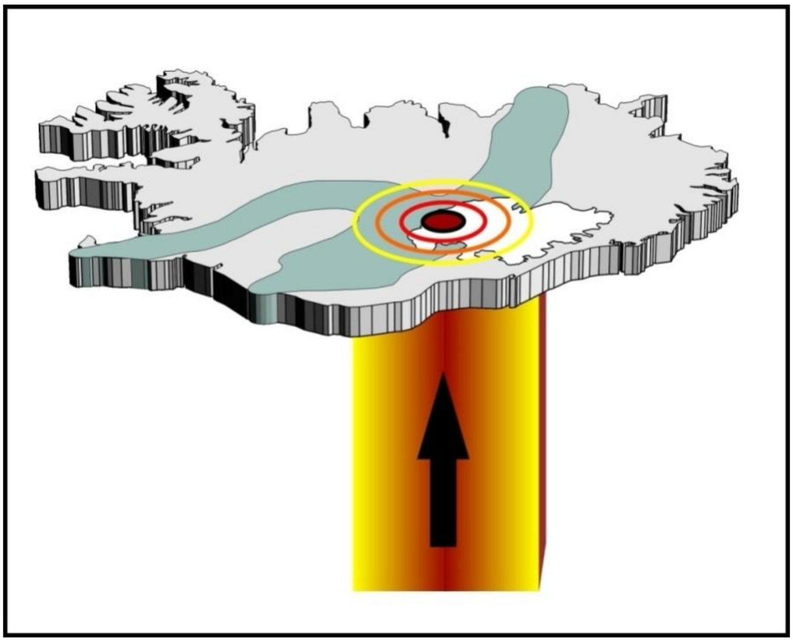

Figure 2.2: Schematic diagram of the mantle plume under Iceland. The approximate location is shown at the surface. Modified after Wolfe *et al*., 1997; Ito, 2001; Shen *et al*., 2002; Gudmundsson, 2007.

The combination of the mantle plume and the spreading centre results in an area of abnormally high volcanic activity. This is clearly seen by Iceland's elevation above its surrounding sea floor. The crustal aggregation caused by these two factors has created an unusually thick section of oceanic crust. Typically, oceanic crust in other locations is ~7 km thick, however, in Iceland there is a crustal thickness of ~40 km above the hotspot, thinning to ~16 km at the Reykjanes Ridge (Foulger *et al*., 2003). Whilst there is thick crust here, however, there is very little lithosphere below the crust (Best, 2003).

Profiles across the MAR highlight magnetic reversals, reading in a barcode-like pattern, mirrored on either side of the ridge. These represent different magnetic epochs, the most recent being the Brunhes normal epoch, which extends back to 0.78 Ma. It is the rocks belonging to this age, in the present tectonic form of Iceland, which hold the majority of present volcanic activity. The rocks of the Brunhes normal epoch are in a band, representing the MAR, and cutting through the country, called the Neovolcanic Zone (Figure 2.3).

2.1.2 Rift Jumps

The mantle plume is assumed to be a stationary feature; whilst the spreading axis drifts independently to the plume. The American-Eurasian plate boundary is thought to migrate in a west-northwest direction, at a rate of 0.3 cm/yr, a movement occurring in addition to the spreading motion

(Thordarson & Hoskuldsson, 2002; Garcia *et al*., 2003). Reconstructions of the track of the mantle plume suggest that the activity goes back to ~130 Ma (Saunders, *et al*., 1997; Tronnes, 2004).

The migration causes the active rift to migrate away from the mantle plume and hotspot. In order to rectify the situation, new rifts are formed progressively eastwards from the spreading centre of the MAR, in order to keep its proximity to the hotspot. This process is known as “rift jump”, and causes the extinction of formerly active rifts, and the initiation of new rifts (Garcia *et al*., 2003). Such a process has occurred several times in Iceland, though there are often two parallel rifts active at the same time, due to the length of time involved in the process, as is currently the case in the south of Iceland. The eastward jumps of the rift in the north of Iceland have led to its location 100km east with respect to the Kolbeinsey Ridge (Garcia *et al*., 2003) (Figure 2.1, Figure 2.3). Previous known jumps of the rift have occurred approximately 24 Ma, 15 Ma, 7 Ma and 3 Ma (Tronnes, 2004) (Figure 2.1).

2.2 The Neovolcanic Zone

2.2.1 Rift Zones and Flank Zones

The Neovolcanic Zone is comprised of three segments; these are the North Volcanic Zone (NVZ), the West Volcanic Zone (WVZ) and the East Volcanic Zone (EVZ) (Figure 2.3). Of these segments, the NVZ and the WVZ are active rift zones, whilst the EVZ is sometimes considered as a flank zone (e.g. Hards *et al*., 2000), or a propagating rift zone. The Snaefellsnes Volcanic Zone is a flank zone in the west of the country, as is the Öraefajökull-Snaefell Zone to the east of the EVZ (Figure 2.3). The area between the WVZ and EVZ experiences a comparatively large amount of seismic activity, and is thus known as the South Iceland Seismic Zone (SISZ) (Figure 2.3).

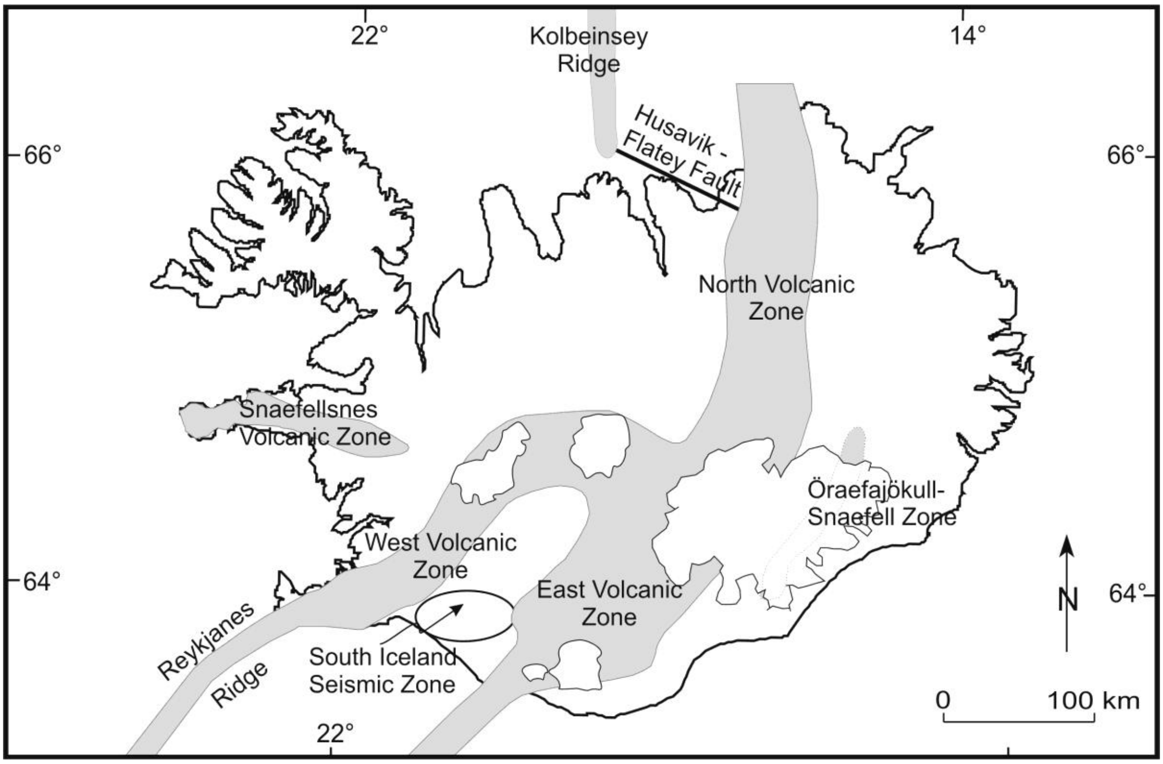

Figure 2.3: Map showing the Neovolcanic Zone at the junction of the Reykjanes and Kolbeinsey Ridges to the south and north respectively. The Snaefellsnes and Öraefajökull Flank Zones are also shown, as well as the South Iceland Seismic Zone. The latitude and longitude are indicated. Modified after Thordarson & Hoskuldsson, 2002; Gudmundsson, 2000.

The NVZ is ~7Ma, and thought to be a reactivation of an older spreading zone in a similar location (Saemundsson *et al.*, 1980; Hardarson & Fitton, 1993). The WVZ is of a similar age to the NVZ, activated ~7-9 Ma when the rift jumped there from the Snaefellsnes Volcanic Zone (LaFemina *et al.*, 2005). The EVZ, by comparison, is thought to have begun its formation ~2-3 Ma within pre-existing crust (Foulger *et al.*, 2003).

According to the theory of rift jump, the EVZ's location parallel to the WVZ shows that it is a rift zone in the making, and will eventually take over from the WVZ, though both are active parallel to each other at present. This mirrors the process that has occurred in other parts of the country in the past. A good example of this is in the north of Iceland, where the main focus of the spreading was previously in the northwest up until ~15 Ma, in what are now the northwest fjords (Figure 2.1). While this rift zone was still active, a new rift zone was evolving as a propagating rift to the east of this, the area now known as Snaefellsnes. The spreading and activity in the Snaefellsnes rift zone increased as the activity in the northwestern rift decreased proportionally. By this process, the Snaefellsnes rift zone became the main focus of the magmatism in the country, though both rift zones remained active parallel to each other for a time (Hardarson *et al.*, 1997).

2.2.2 Spreading Rates

The spreading rate in Iceland varies according to the specific rift zones, as does the spreading vector. The spreading rate of the NVZ is 1.8-2.0 cm/yr, and is shared between the two volcanic zones in the south of the country. The spreading rate on the WVZ increases from 0.2 cm/yr in the northeast of the zone, to 0.7 cm/yr in the southwest. The EVZ is spreading at a rate of 1.9 cm/yr in the northeast, decreasing to 1.1 cm/yr in the southwest. The increase in spreading of the WVZ to the south, and the opposite pattern in the EVZ are thought to be consistent with the propagating rift and rift jump models, where the EVZ is propagating, and the WVZ is deactivating to the southwest (LaFemina *et al.*, 2005).

2.2.3 Petrology

Almost all of the bedrock in Iceland is composed of basalt, which varies depending on its age and the conditions of its formation. The bedrock can be classified into four types according to their age: (1) the Tertiary bedrock, dating 16-3 Ma; (2) the Older Pleistocene formation, dating 3-0.78 Ma; (3) the Younger or Late Pleistocene formation, dating 0.78 Ma-10 ka, and finally (4) the Holocene formation, from the last 10 ka (Gudmundsson, 2007) (Figure 2.4, note that 3 and 4 and marked as one subdivision). These formations lie progressively laterally from the Neovolcanic Zone and present position of the MAR, thus marking the expansion of Iceland with age (Figure 2.4). There are Holocene sandur deposits along the south coast (Figure 2.4), commonly marking the past paths of jökulhlaups.

The Pleistocene formation in particular is strongly characterised by hyaloclastite ridges, glacial tillites, glaciofluvial sediments, and some marine sediments (Gudmundsson, 2007). The presence of such rocks highlights the occurrence of both subglacial activity and glacial erosion during a period of widespread glaciation. The Tertiary formations by comparison hold almost no evidence of glaciation, indicating a much warmer climate than in Quaternary times.

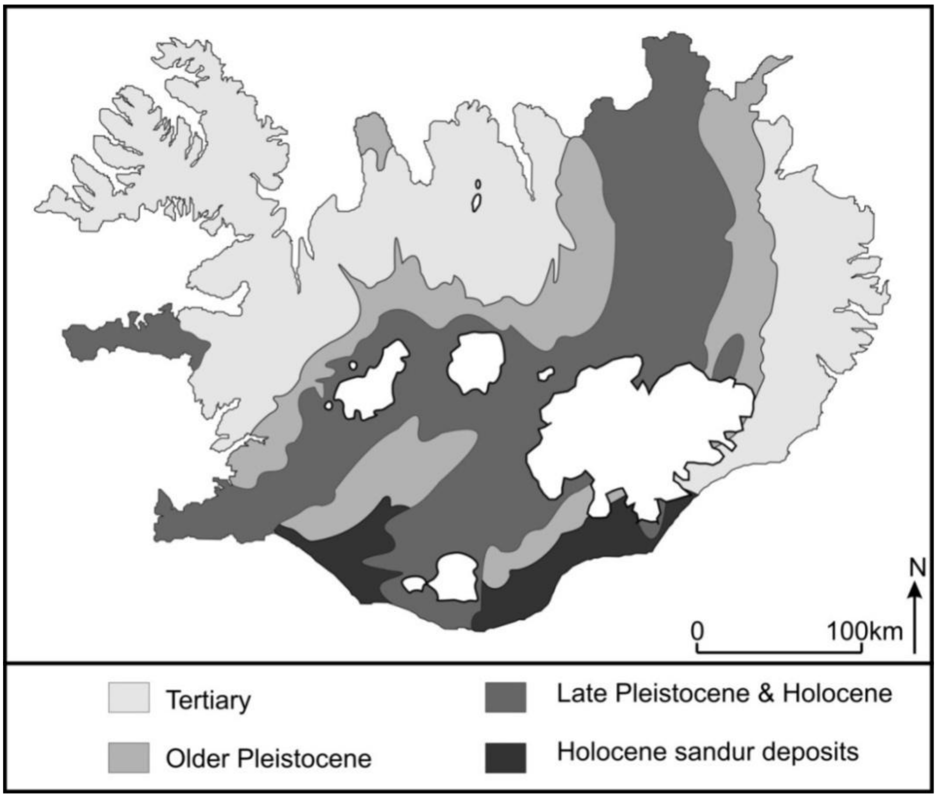

Figure 2.4: Map showing the major geological subdivisions of the bedrock of Iceland. Modified after Thordarson & Hoskuldsson, 2002.

The postglacial igneous rocks (of the Holocene formation) can be further divided into three evolutionary series, each holding a compositional SiO_2 range from basic to acid (Gudmundsson, 2007). The locations of these series can be seen in Figure 2.5:

1) Tholeiite series: These occur in the active rift zones.
2) Alkali series: These are found in the flank zones, where there is no rifting, as well as the southern end of the EVZ.
3) Transitional series: These are found in the transition from propagating to rift zone in the EVZ.

The most widespread rocks in Iceland come from the tholeiitic series (Figure 2.5). Tholeiite volcanism indicates a high percentage of melting, and is commonly associated with Mid Ocean Ridge Basalts (MORB). They are also linked to the presence of the mantle plume, particularly the close proximity to it. Alkalic series by comparison, indicate a lower percentage of melting and are more commonly located slightly further away from the plume and the ridge. In keeping with this, the tholeiitic rift zone and mildly alkalic flank zone volcanism in Iceland is thought to be roughly equivalent to the main shield-building tholeiitic stage, and the pre- and post-shield-building alkaline stages of Hawaiian volcanism (Best, 2003; Tronnes, 2004).

The tholeiitic volcanism is also in part a result of the thin lithosphere beneath Iceland (Figure 2.6A) allowing higher levels of partial melting in the ascending decompressing mantle (Best, 2003). By comparison, other oceanic islands have a much thicker lithosphere (Figure 2.6B). This means that there can only be small degrees of partial melting before the ascending decompressing mantle reaches the more rigid and un-convecting lithosphere (White, 1993; Best, 2003). This latter situation creates predominantly alkaline basalts.

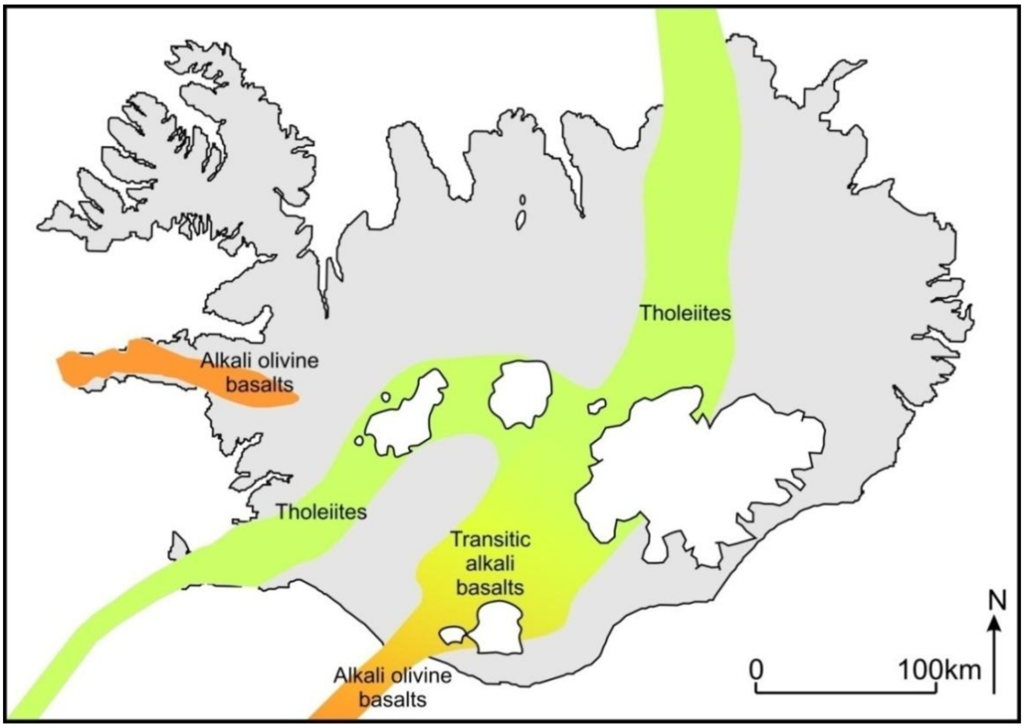

Figure 2.5: Map showing the established regional variation of the chemical composition of eruptive rocks. Modified after Gudmundsson, 2007.

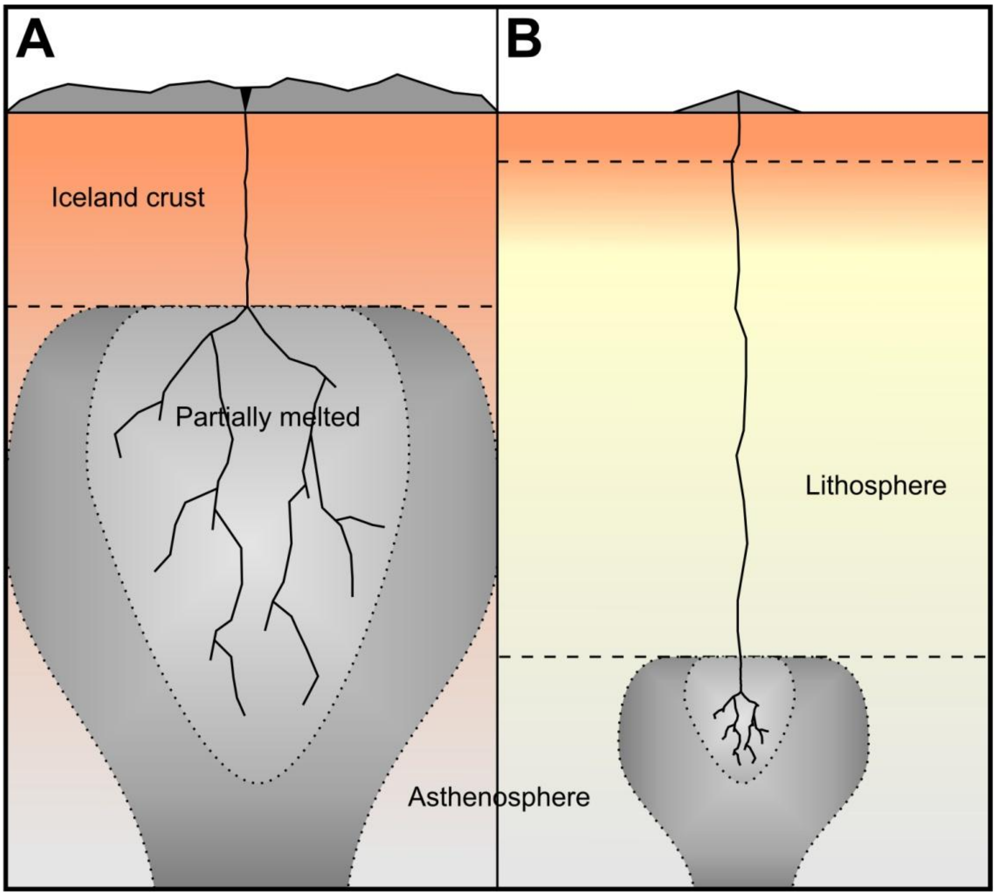

Figure 2.6: Schematic diagram showing partial melt production in the ascending decompressing mantle below a very thin lithosphere in Iceland (A), and a thick lithosphere beneath other oceanic islands (B). Modified after Best, 2003.

2.2.4 Geological Features

2.2.4.1 Volcanic Systems

Volcanism within the Neovolcanic Zone is mainly concentrated in volcanic systems and these are the main geological features in Iceland. There are 30 volcanic systems, distributed evenly throughout the Neovolcanic Zone (Figure 2.7). The volcanic systems can be defined using petrological characteristics (Jakobsson *et al*., 1978) or tectonic characteristics (Saemundsson, 1978), or both. They have a typical lifetime of 0.5-1.0 Ma (Thordarson & Larsen, 2007). The systems are the present-day form of the regional swarms of dykes and faults, identified by Walker (1960), that are exposed in the Tertiary and Pleistocene lava pile.

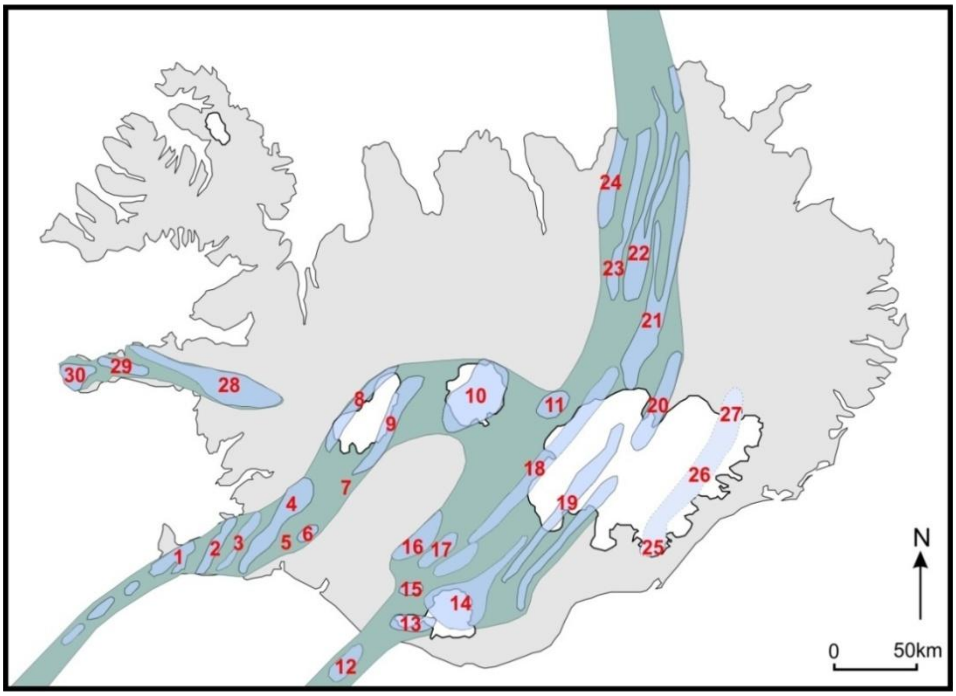

Figure 2.7: Distribution of active volcanic systems: (1) Reykjanes-Svartsengi, (2) Krysuvik, (3) Brennisteinsfjöll, (4) Hengill, (5) Hromundartindur, (6) Grimsnes, (7) Geysir, (8) Prestahnjukur, (9) Langjökull, (10) Hofsjökull, (11) Tungnafellsjökull, (12) Vestmannaeyjar, (13) Eyjafjallajökull, (14) Katla, (15) Tindfjöll, (16) Hekla-Vatnafjöll, (17) Torfajökull, (18) Bardarbunga-Veidivötn, (19) Grimsvötn, (20) Kverkfjöll, (21) Askja, (22) Fremrinamur, (23) Krafla, (24) Thestareykir, (25) Öraefajökull, (26) Esjufjöll, (27) Snaefell, (28) Ljosufjöll, (29) Helgrindur, (30) Snaefellsjökull. Modified after Johannesson & Saemundsson, 1998; Gudmundsson, 2000; Thordarson & Larsen, 2007.

The volcanic systems mainly feature a fissure (dyke) swarm or a central volcano, or both (Saemundsson, 1978; 1979; Jakobsson *et al*., 1978; Jakobsson, 1979). The fissure swarms are elongated structures, and are normally aligned sub-parallel to the axis of the volcanic zone hosting the system (Thordarson & Larsen, 2007). Most volcanic systems in the flank zones lack well developed fissure swarms. The central volcanoes, where present, act as a focal point for volcanic activity and are normally the largest edifices within the system (Thordarson & Larsen, 2007).

Volcanic activity on the volcanic systems is linked to plate movements. The spreading and subsequent rifting of the crust occurs in distinct rifting episodes, most commonly confined to a single system at one time. However, near-concurrent activity on two or more systems is known to have happened (Thordarson & Larsen, 2007 and references within). Episodes of rifting are characterised by recurring earthquake swarms, and volcanic eruptions on the fissure swarms and within the central volcanoes.

2.2.4.2 Central volcanoes

There are 23 central volcanoes located within only 19 of the volcanic systems, indicating that some of the volcanic systems hold more than one central volcano (Thordarson & Larsen, 2007). The characteristics of a central volcano are: that it erupts frequently; it can extrude basaltic, intermediate and acid lavas; it is associated with and fed by a shallow crustal magma chamber, and is often associated with a collapse caldera (Gudmundsson, 1995). Central volcanoes are also normally associated with well defined swarms of tension fractures, normal faults and volcanic fissures (Gudmundsson, 2000).

Central volcanoes are built by repeated eruptions from a central vent. The vent is maintained by a long-lived plumbing system including the shallow magma chamber. The presence under several of the active central volcanoes of shallow magma chambers has been confirmed by both geodetic (Sturkell *et al*., 2006) and petrologic (Sigmarsson & Steinthorsson, 2007) data.

The remaining volcanic systems without central volcanoes hold high-temperature geothermal fields, which may represent central volcanoes in their earliest stage of growth (Johannesson & Saemundsson, 1998; Thordarson & Larsen, 2007). The Vestmannaeyjar volcanic system (Figure 2.7) is also thought to be developing a central volcano, because the volcanic activity in the system is concentrating around the island of Heimaey (Johannesson & Saemundsson, 1998; Mattsson & Hoskuldsson, 2003).

2.2.4.3 Fissure Swarms

Of the 30 volcanic systems, 20 feature a fissure swarm (Johannesson & Saemundsson, 1998). These exist in varying states of maturity, ranging from embryonic through to mature (Thordarson & Larsen, 2007). The lengths of the intact volcanic fissures are equal to the surface lengths of their feeder dykes (Gudmundsson, 2000).

There are 4 embryonic fissure swarms, holding one or a few discrete volcanic fissures. 12 of the fissure swarms are considered well developed and mature. These are distinct narrow and elongated features (5-20 km wide and 50-200 km long), holding a high density of normal faults, fissures and tension cracks (Thordarson & Larsen, 2007).

2.3 Volcanic Activity

Whilst there is an even distribution of the central volcanoes throughout the Neovolcanic Zone (Figure 2.8), there are other types of volcanic activity that are unevenly distributed. A notable pattern in the characteristic volcanism is the distribution of the shield volcanoes in the NVZ and WVZ. The EVZ, by comparison holds no shields, but holds all of the “large fissures” (Figure 2.8).

2.3.1 Shield Volcanoes

When discussing the shield volcanoes in Iceland, it is important to make the distinction between Icelandic shields and those on other basaltic islands such as Hawaii and the Galapagos Islands. Outside Iceland, shield volcanoes are polygenetic structures representing the accumulation of a large amount of fluid lava, giving a low profile edifice. Such shield volcanoes are mainly observed in intra-oceanic locations. In Icelandic terms these shield volcanoes would be classified as central volcanoes. In Iceland, shields are monogenetic structures, and the common terminology used to define them is lava shields (Walker, 1971; Jakobsson *et al*., 1978; Andrew & Gudmundsson, 2007, *Appendix I*).

Shield volcanoes can be seen throughout the Tertiary and Pleistocene lava piles, showing a common presence throughout the geological history of Iceland (Gudmundsson, 2000; Andrew &

Gudmundsson, 2007, *Appendix I*). The Holocene shields are almost uniquely confined to the early post-glacial times (Rossi, 1996; Rossi & Gudmundsson, 1997), and are all thought to be older than 3500 years (Thorarinsson *et al*., 1959).

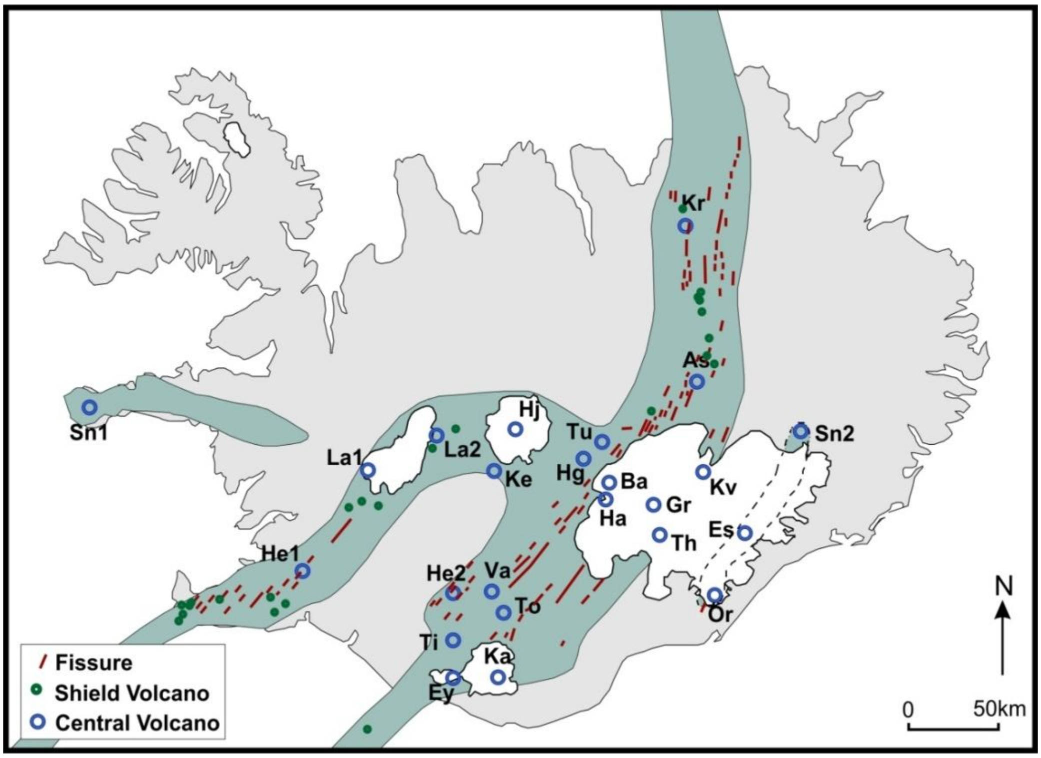

Figure 2.8: Distribution of central volcanoes, shield volcanoes and some fissures. The central volcanoes labelled and are as follows: Sn1 = Snaefellsjökull, He1 = Hengill, La1 & 2 = Langjökull, Hj = Hofsjökull, Ke = Kerlingarfjöll, Ey = Eyjafjallajökull, Ti = Tindfjallajökull, Ka = Katla, He2 = Hekla, To = Torfajökull, Va = Vatnafjöll, Ör = Öraefajökull, Es = Esjufjöll, Sn2 = Snaefell, Th = Thordarhyrna, Gr = Grimsvötn, Kv = Kverkfjöll, Ha = Hamarinn, Ba = Bardarbunga, Hg = Hagongur, Tu = Tungnafellsjökull, As = Askja, Kr = Krafla. Modified after Johannesson & Saemundsson, 1998; Gudmundsson, 2000 (and references within).

An interesting feature of the shield volcanoes is that they are commonly located on the margins of, in between, or outside the volcanic systems. They are composed of primitive lavas, that is picrite and olivine tholeiite (Rossi 1996; Rossi & Gudmundsson, 1997), which in turn affects their size, the former being commonly much smaller. The primitive lavas indicate that they directly tap the deep seated magma reservoir, at the crust-mantle boundary, rather than a shallow crustal magma chamber.

2.3.2 Fissures

As noted earlier (Section 2.2.4.3), there are fissure swarms throughout the Neovolcanic Zone; however, the EVZ holds anomalously large fissures that are not present in the NVZ and WVZ (Figure 7.1). These large fissures experience much larger volumes of lava than the other fissures, and their eruptions are much more prolonged, lasting for months to years, and feature numerous eruptive episodes.

These eruptions, also known as flood-lava events, are thought to be fed, as with the shields, from the magma reservoirs due to the large volumes of erupted material (Sigmarsson *et al*., 1991; Bindeman *et al*., 2006). This theory has been heavily disputed in the past. The large fissures are commonly composed of mixed cone rows with a combination of spatter and scoria cones, and can feature volcanogenic chasms (Thordarson & Larsen, 2007).

There are two notable large fissures in the EVZ, which define the term; they are the Laki (Skaftar Fires) eruption of 1783-84, and the Eldgja eruption of 934 AD. The Laki eruption opened a 27 km long fissure and erupted ~15 km^3 of lava (Thordarson & Self, 1993; 2003). The Eldgja eruption formed a 57 km long fissure and erupted ~20 km^3 of lava (Thordarson *et al*., 2001). The lavas of the two eruptions were evolved tholeiite from Laki and transitional alkali from Eldgja (Thordarson *et al*., 2001). Both eruptions are considered to be major volcanotectonic rifting episodes.

2.3.3 Subglacial Volcanism

Given the glacial history of Iceland, and the presence of 4 major ice bodies on the island (e.g. Figure 2.3), subglacial volcanism is a strong feature. When magma reaches the surface under ice bodies, the heat transfer to the ice is so efficient that it normally enters into a subaqueous environment. This is in the form of a water filled cavity or an ice-dammed lake (Allen, 1980; Tronnes, 2004).

Subglacial volcanic mountains are commonly very topographically high structures. They are comprised of pillow lavas, overlain by pillow breccias and hyaloclastite tuff. These formations reflect the decreasing hydrostatic pressure as the mountain grows higher, and thus shallower in the water or ice, during the eruption (Allen, 1980).

There are two major types of subglacial volcanoes, which are directly comparable to their subaerial counterparts. Table mountains are shield volcanoes formed subglacially (with similar primitive magmas) and similarly found on the margins of the volcanic systems (Schiellerup, 1995; Andrew & Gudmundsson, 2007, *Appendix I*). Table mountains are thought to be a good indicator of the ice sheet elevation and the thickness at the time of their eruption (Licciardi *et al*., 2007).

The second type of subglacial volcanism is hyaloclastite ridges. These are again comparable to their subaerial form of fissures. They are formed by subglacial fissure eruptions, and are elongated. These can be comprised of hyaloclastite cones, which are formed where the subglacial eruptions have localised to individual vents (Andrew & Gudmundsson, 2007, *Appendix I*). There are many hyaloclastite ridges and cones located to the east of the NVZ (Bourgeois *et al*., 1998) (Figure 5.2), though they occur throughout the Neovolcanic Zone. Lithological evidence suggests that a significant part of the young hyaloclastites were deposited towards the end of each glacial period, as the ice was thinning (Sigvaldason *et al*., 1992).

2.4 Volcanic History

2.4.1 Pleistocene

In order to consider the volcanic history of the Neovolcanic Zone in the Holocene, it is necessary to first look back into the Pleistocene. Whereas for the Holocene, the volcanic history record is quite clear, it becomes less precise looking further back (Mortensen *et al*., 2005). The record has been comprised through tephrochronology, ice core records and, more recently, historical records.

Iceland has been subject to several glaciations over the past 4 Ma, where an ice cap has covered the whole island (Bourgeois *et al*., 1998; Einarsson *et al*., 1988). The most recent glacial period was the Weichselian, which started ~70 ka and ended ~12 ka (Maclennan *et al*., 2002). Volcanic activity is thought to have begun to increase with the retreat of the ice sheet. In particular, there has been an increase in hyaloclastite activity towards the end of the Pleistocene (Bourgeois *et al*., 1998; Andrew & Gudmundsson, 2007, *Appendix I*).

2.4.2 Early Holocene Activity

The increase in volcanic activity continued into the early Holocene. In some parts of Iceland, it was as high as 30-50 times higher immediately after deglaciation compared to more recent times (Maclennan *et al*., 2002). The high eruption rates are thought to have persisted for <1.5 ka after the deglaciation, during this period of time more than 75% of the postglacial volume of magma was erupted (Maclennan *et al*., 2002). This increase in activity has been attributed to the increased magma productivity, related to the isostatic uplift, following the glacial retreat. The exact mechanism behind this increased productivity is a matter of much debate (Federova *et al*., 2005; Jull & McKenzie, 1996; Maclennan *et al*., 2002; Sigvaldason *et al*., 1992; Sigvaldason, 2002; Slater *et al*., 1998).

The increase in activity has been studied in many different volcanic systems, such as Snaefellsnes (Hardarson & Fitton, 1991), Dyngjufjöll, and the Reykjanes Peninsula (Sigvaldason *et al*., 1992; Gee *et al*., 1998). The production of shield volcanoes has been directly linked to the glacial retreat (Rossi, 1996; Andrew & Gudmundsson, 2007, *Appendix I*). The number of central volcanoes in Iceland since the last glacial period is also thought to be higher than before.

2.4.3 Historical Activity

The Holocene volcanic activity has been well documented, with a good range of tephrochronology data, ice core data and historical accounts (Thordarson & Larsen, 2007 and references within). Thordarson and Larsen (2007) performed an extensive study into the volcanic history, its distribution throughout the Neovolcanic Zone and patterns within, during historical times (the past 1100 years). Of the 30 volcanic systems, 16 have been active in historical times, with 205 recorded events; of which 192 represent single eruptions.

The study by Thordarson and Larsen (2007) shows that the distribution of activity within the Neovolcanic Zone has been remarkably uneven. The volcanic systems of the EVZ produced approximately 80% of the volcanic events, and the bulk of this activity came from 4 of the systems, Grimsvötn, Hekla, Katla and Bardarbunga-Veidivötn (Figure 2.7). There is a similar pattern with the volume of erupted magma in the Neovolcanic Zone. The EVZ accounts for 82% of the total volume, and 97% of that from the EVZ (79% of the total output) comes from just 4 of the systems, Grimsvötn, Hekla, Katla and Bardarbunga-Veidivötn – which also contain the large fissures.

The eruption frequency in the EVZ is also higher in the 4 most active volcanoes, with a recurrence interval of years to decades. For the whole of Iceland the repose periods between eruptions vary from <1 year to >20 years. This leads to the conclusion that the EVZ is the dominant contributor to volcanic production in historical time. It can be considered as the main corridor for magma to the surface in historical times (Thordarson & Larsen, 2007).

3 Modelling

3.1 Types of Models

Before introducing numerical modelling, as used in this work, it is important to introduce the concept of modelling as a tool in structural geology, and in this case, volcanotectonics and rock mechanics. A model is essentially a representation of a structure or scenario. In structural geology, models are used to understand the formation and interaction between geological structures.

Models can be created in a very realistic way. However, they can never be a perfect representation of reality. As a result of this there can be an infinite number of models created before a solution is reached. It is common that models simplify reality, and that the solution chosen as representative is closest to the reality (Logan, 2002).

Model solutions should normally be based on data collected either from the field or from literature reviews. The solution should be compared to observations in the field, or used as predictions for future scenarios. There are many types of models using varying approaches, a few of which are outlined here.

Theoretical Models

A theoretical or conceptual model is a description of the components or variables being studied in a particular situation, together with a hypothesis for the relationship between the variables. Such models are normally accompanied by a pictorial representation of the variables and their relationships. Theoretical models can help to illustrate theories to other people, to clarify points of further analysis, and to predict possible relationships between the variables.

Analogue Models

Analogue models use scaled versions of a certain phenomenon to create a more understandable or analytical form. Essentially they are an accurate representation of a structure, for example a caldera or fault, created using materials that share the same relative scaled properties as those in natural form. By reproducing a structure, and applying forces, it is possible to speed up the process on a smaller scale and assess how it occurred. However, it should always be taken into account that it is impossible to scale all parameters correctly, and thus accuracy is compromised by this.

Analytical Models

Analytical models use equations to examine changes within a system. In structural geology, or more specifically, rock mechanics, they calculate stress, strain and displacement caused by applied loads. The calculations are made according to the laws of physics and rock mechanics. These are commonly equations used to examine processes such as dyke injection in a volcanotectonic context.

Analytical models are solved “by hand”, however as the models become more and more complex (and the equations comprising them) a computer program can be used, whereby they are then termed numerical models.

3.2 Numerical Models

Numerical models are used to simulate physical problems when analytical solutions become too complex (Logan, 2002). Numerical models create an environment where a model can be broken down into elements and nodes, through a process known as discretisation. Each element represents its own equation solved by the program. The numerical modelling programs allow complex structures to be designed or represented, and have loads applied, then solved.

There are many types of numerical models one of which, the Finite Element Method (FEM), allows the creation of 3D elements, solving the problem in terms of volume (Logan, 2002). The method uses an approximation of differential equations meaning that the results are also an approximation, and thus less accurate in general. However, this method allows easier creation of more complex models, and is thus used in this study. All models here were performed using the Finite Element program Ansys (www.Ansys.com; Logan, 2002).

Numerical modelling involves three stages:

1. Pre-processing: At this stage the user defines the model and the environmental factors to be applied to it. This means designing the geometric structure of the model, which can be done either within the program, or in a computer drawing package and then imported. This geometric representation is then divided into elements and nodes, forming a mesh. The nodes on this mesh are the points where features such as displacement are calculated.
2. Analysis: Properties such as Young's modulus and Poisson's ratio are applied to the materials within the model. Also loads such as force, pressure and gravity are applied. The solution of the models shows the effects of such forces on the structure modelled.
3. Post-processing: The programs allow the visualisation of the results in picture, graph and numerical forms. These results allow the user to identify the implications of the analysis.

3.3 Models of Iceland

In this study all numerical models made were based on the geodynamics of Iceland. As a result of this all models had a load of -5 MPa tension applied perpendicular to the spreading axis, in order to replicate the rifting of the divergent plate boundary. Series of models with varying levels of complexity were run for each of the problems studied. The results given for each problem were chosen as the best match for the scenario considered.

Mechanical properties were assigned to the different materials used in the models, and these differ according to the material. The properties applied were Young's modulus and the Poisson's ratio. Young's modulus, or stiffness, of a material is one of the main measures of its elastic properties. It is the ratio between stress and strain in the one-dimensional Hookes law, and is indicated by the slope of the stress-strain curve (Hudson & Harrison, 1997; Jaeger *et al*., 2007). The Young's moduli assigned to the materials used in this study were taken as educated approximations and generally accepted for basalts and lava flows. Where different values are used, they are specified in the papers. Poisson's ratio is the ratio between lateral contraction and elongation of a material, with values commonly between 0.1 and 0.3 in solid rocks (Pollard & Fletcher, 2005). In all materials in all models used in this study, it is assigned a value of 0.25.

When making the numerical models, it is important to consider interference from the imposed boundary conditions. In order to counteract any interference, whilst models were all made according to scale, they are generally made considering a larger area surrounding the area in question. Results shown of the problems are always given with a small cartoon depicting the area modelled and the area given in the result.

4 Aims

This study aims to examine the volcanotectonic evolution of Iceland. This includes the development and behaviour of some of the characteristic volcanism in the Neovolcanic Zone. The aims are achieved through the means of numerical modelling. The specific aims are as follows:

- To examine the relationship between deglaciation and the increase in volcanic activity from a mechanical perspective

- To propose a model for the creation of shield volcanoes being restricted to the early postglacial

- To further these ideas of shield volcanoes to include other types of volcanic activity, specifically central volcanoes

- To examine the possibility of mechanical interaction between central volcanoes

- To examine the mechanical interaction of volcanoes with other geological structures of varying scales

- To look at the activity of the EVZ and the characteristic volcanism, specifically the phenomena of large fissures

- To put all these findings into the context of hazards and risk on both a proximal and distal scale

5 Deglaciation and Volcanic Activity

5.1 Introduction

The increase in activity coinciding with the retreat of the Weichselian ice sheet has been mentioned above (Sections 2.4.1, 2.4.2). High levels of volcanic activity at this time are not disputed, and have been well documented. There is good evidence that the volume production of most of the volcanic systems peaked at this time, and has decreased markedly since (Annertz *et al*., 1985; Gudmundsson, 1986; Vilmundardottir & Larsen, 1986; Jull & McKenzie, 1996; Bourgeois *et al*., 1998).

The so called "pulse" in activity has been attributed by some authors (Gudmundsson, 1986; Sigvaldason *et al*., 1992) to the tapping of pooled crustal magma chambers caused by the changes in tectonic stress and magmatic pressures during the unloading of the ice. Other authors (Jull & McKenzie, 1996; Hardarson & Fitton, 1997; Slater *et al*., 1998; Maclennan *et al*., 2002) attribute the pulse to higher magma supply derived from increased melt generation in the upper mantle stimulated by the glacial rebound.

Theoretical models by Jull & McKenzie (1996) support this second theory, that rapid glacial unloading can create short lived pulses of decompression-induced melt production and eruptive activity. Theoretical and analytical models, supported by numerical models were made by Andrew & Gudmundsson (2007, *Appendix I*), on the first theory. These show the tectonic influences of the ice load on the volcanic systems, primarily looking at the shield volcanoes. Here this theory is extended to include hyaloclastite ridges and cones, table mountains and central volcanoes.

5.2 Glacial Periods

Due to the divergent plate boundary in Iceland, the normal state of stress in the crust is an extensional regime with absolute tension at the surface. There is normally melting at the crust-mantle boundary and below, creating a magma reservoir at this level, as is the case at most divergent and intraplate volcanic areas (Figure 5.1A).

The addition of ice creates a state of compression at the surface. At the last glacial maximum there was approximately 2 km of ice, which would create compression as deep as the magma reservoir (Figure 15, Andrew & Gudmundsson, 2007, *Appendix I*). The compression of the ice causes a down-bending of the crust (Figure 5.1B), where there is compression above the neutral surface, and relative tension below. As this is the case around the magma reservoir, a situation is created where the area surrounding the reservoir is altered. Increased melting is likely to occur as the ice begins to recede, and relative stresses begin to be released.

5.3 Subglacial Volcanism

As the ice recedes, the compression it creates is decreased from the margins working inwards. This corresponds logically to the ice receding from the coast of Iceland inland. As the ice retreats tensile stresses around the magma reservoir lead to a situation of increased porosity. This allows an increased level of melting. With the release of compression, dykes are able to inject to the surface underneath the ice, forming subglacial volcanoes.

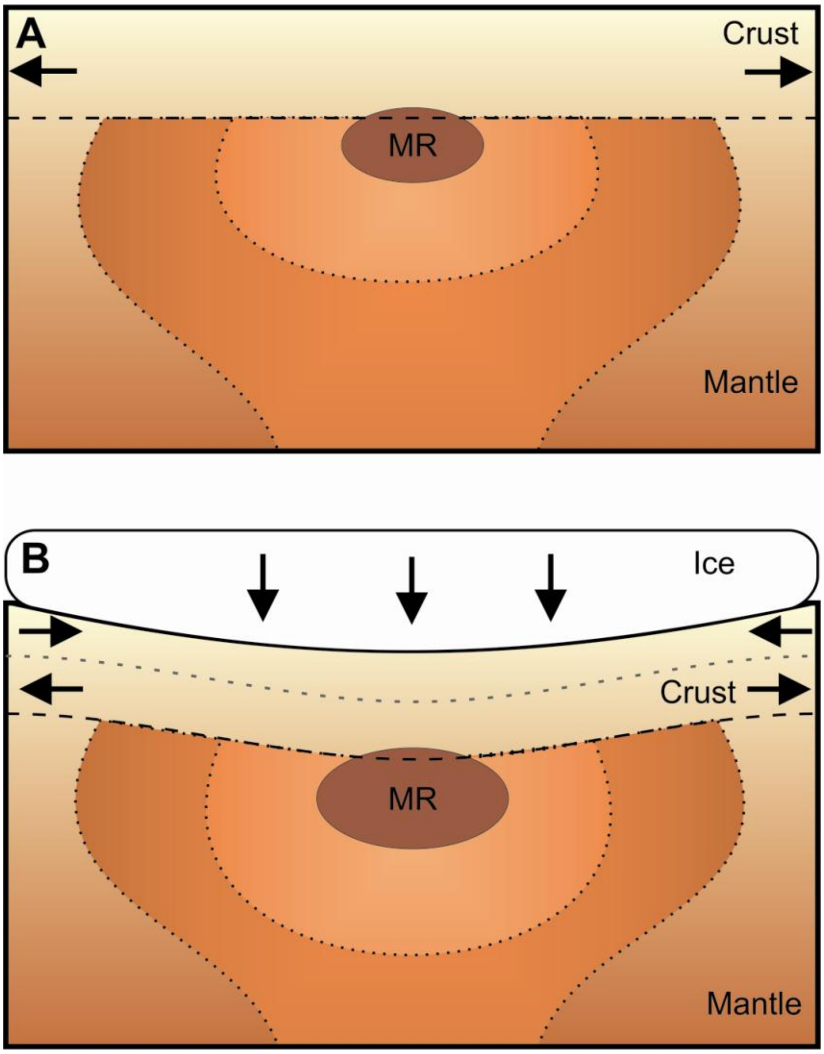

Figure 5.1: Schematic diagram showing the areas of melt at and under the crust/mantle boundary (shown as dashed black line): A) the crust in Iceland under "normal" conditions, with relative tension indicated by arrows. B) the crust in Iceland under compression from ice sheet, creating compression and a difference in stresses above and below the free surface (depicted a dotted grey line). The magma reservoirs are marked as "MR", and the area of melting changes under each regime.

The eruptions of the table mountains have been correlated with the thinning of the ice as the ice sheet retreats (Licciardi *et al*., 2007), and this correlation has shown that the ice persisted longer in the WVZ than in the rest of Iceland. In the north of Iceland, the hyaloclastite ridges are known to be on the coastal side of the NVZ in the Fjallgardar ridge (Figure 5.2), adding support to this idea (Bourgeois *et al*., 1998). This ridge also shows the subglacial volcanic activity on the margins or outside the volcanic systems (Figure 5.2).

The relaxation of the compression from the ice increases the likelihood of dyke injections being able to propagate to the surface. The eruptions are thought to be fed from the magma reservoirs at the margin of the volcanic systems, rather than from the shallow crustal magma chambers, as they are still under the compression from the ice (Figure 5.3).

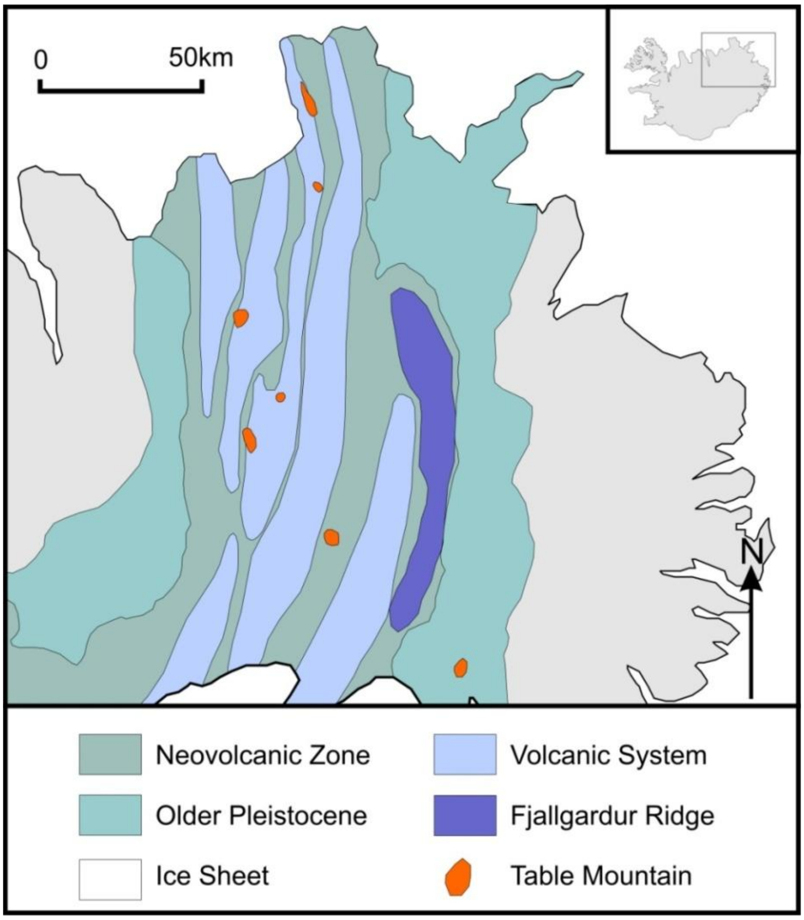

Figure 5.2: Map of northeast Iceland (located in figure) showing the location of some (not all) table mountains and hyaloclastite ridges. Note the location of the subglacial features at the margins or outside the volcanic systems. Modified after Bourgeois *et al*., 1998; Licciardi *et al*., 2007.

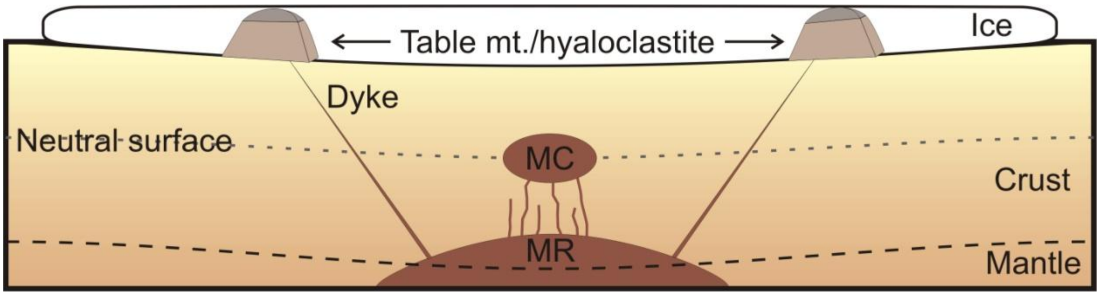

Figure 5.3: Schematic diagram showing a receding ice sheet and decompressing crust. Dykes coming from the magma reservoir (MR) are mostly stopped at the neutral surface forming a shallow crustal magma chamber (MC). Due to the decompression, some dykes are able to reach the surface, feeding directly from the magma reservoir, these form table mountains and hyaloclastite ridges.

5.4 Shield Volcanoes

When the ice has retreated, the first type of volcanic activity to appear is the shield volcanoes, also on the margins or outside of the volcanic systems (Andrew & Gudmundsson, 2007, *Appendix I*). These tap the magma reservoirs as with the subglacial volcanism, and are composed of primitive lavas. Further details on the shield volcanoes are given by Andrew & Gudmundsson (2007, *Appendix I*) as part of this work.

5.5 Central Volcanoes

There are thought to be more central volcanoes in the Holocene than previously in Iceland. In addition to this, as mentioned before, there is a well-documented increase in the activity of the central volcanoes, immediately after the ice retreat. The formation and increased activity of the central volcanoes can be explained by theoretical mechanical activity following the ice retreat.

As mentioned in section 5.2, when the ice compresses the crust, and then begins to recede, there is a difference in the stresses above and below the neutral surface (Figure 5.1). The relative tension in the lower part of the crust not only allows increased melting around the magma reservoir, but can also allow injection of dykes. These dykes are able to propagate as far as the neutral surface where they become arrested by the contrast in mechanical properties, in this case, the change in stress regime (Figure 5.4A).

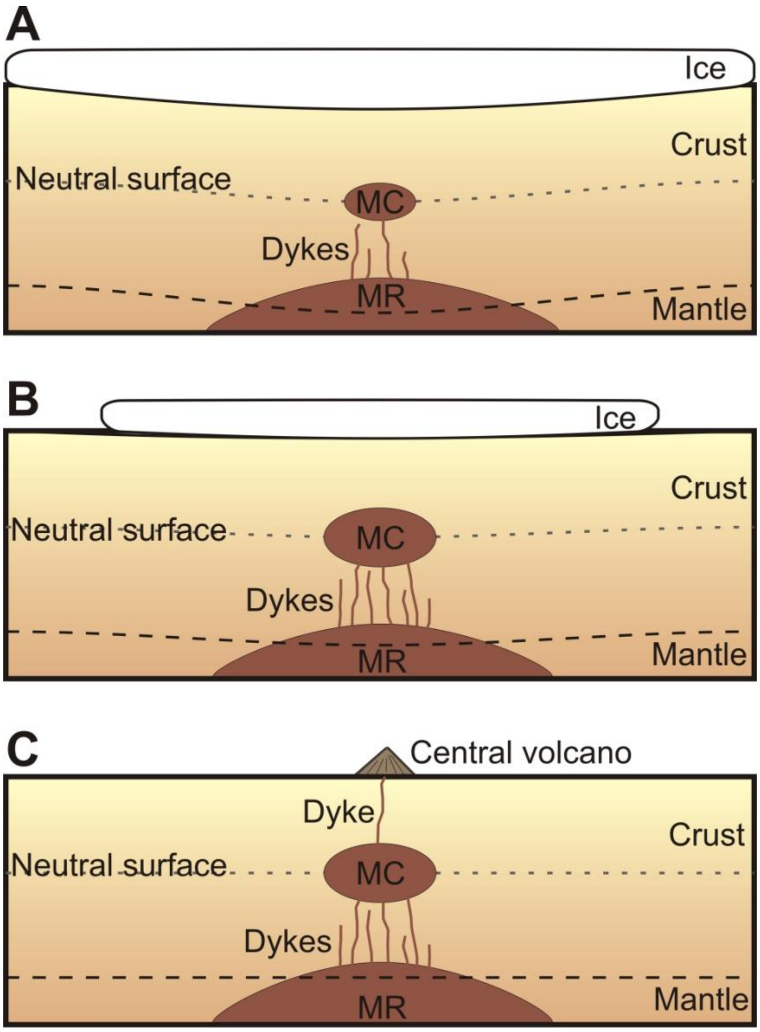

Figure 5.4: Schematic diagram showing stages of development of central volcanoes due to receding ice. A) the compression of the ice allows dykes to feed from the magma reservoir (MR), but only reach the neutral surface, where they are arrested due to the change in stress regime, and begin to form a shallow crustal magma chamber (MC). B) as the ice recedes an increased number of dykes reach the magma chamber and so it grows. C) when the ice has finally gone, dykes are able to feed from the magma chamber to the surface creating central volcanoes.

At a certain threshold, continued injections of magma to this neutral surface lead to the creation of a shallow crustal magma chamber (Figure 5.4A). Due to the compression of the ice, the shallow magma chamber cannot inject dykes to the surface, but continues to receive and fill with magma from below (Figure 5.4B).

When the ice finally retreats, the relaxation of the compression of the upper crust, and thus isostatic uplift, create a situation where the newly developed magma chamber can inject dykes to the surface (Figure 5.4C). Due to the changed stress regime, that is a favourable stress field throughout the crust (relative tension), and favourable overpressure within the magma chamber, these dykes are better able to reach the surface. Therefore there is a high level of volcanic activity at this time.

5.6 Distribution, structure, and formation of Holocene lava shields in Iceland

Here we present the results of new field observations of Holocene shields and provide numerical models to explain their location, time of formation, primitive composition, and large volumes. We made models with varying ice-sheet size and thickness (glacial load), the ice resting on a mechanically layered crust, and studied the stress effects that the load would have on a double magma chamber, that is, a small, shallow crustal chamber at 3 km depth and a deep-seated reservoir at 20 km depth (the base of the crust). Such a pair of chambers is typical for volcanic systems and associated central volcanoes (composite volcanoes) in Iceland. For an ice sheet covering an entire volcanic zone or more, that is, 100 km or wider, the ice-induced compressive stress extends to the deep-seated reservoir and into the upper mantle. Consequently, such a loading suppresses magma accumulation in the reservoir and associated volcanism. During the late-glacial period, when the ice sheet is only 20 km wide, the glacial load generates tensile stresses around the deep-seated reservoir, increases its fracture porosity and magma content, and extends the reservoir laterally and vertically into the upper mantle. Consequently, when the lava shields (and, somewhat earlier, the table mountains) were erupted, much more melt or magma was available to feed a single eruption than during the later part of the Holocene. And because of the greater vertical extent of the reservoir, this magma tended to be hotter and more primitive than that issued in later-formed fissure eruptions. Also, the stress field generated at the end of the glacial and in the early Holocene favoured dyke injections at the marginal parts of, or in between, the volcanic systems, thereby explaining the location of the lava shields.

The full text of this paper is presented as *Appendix I* at the end of this study.

6 Mechanical Interaction

6.1 Introduction

6.1.1 Inclusions

For the purposes of analysing volcanoes in a tectonic perspective, it is possible to consider them as inclusions in the bedrock or crust matrix. An inclusion may be defined as a body with material properties that contrast with those of the surrounding material, commonly referred to as the “matrix”. In the context of rock mechanics, it is a material body hosted by a larger body with different elastic properties (Gudmundsson, 2006).

Inclusions in an elastic body can change the stress fields immediately surrounding them, most commonly concentrating the magnitudes of stress. When inclusions come into close proximity of each other, they can interact mechanically, as is studied here in the context of volcanoes (Gudmundsson & Andrew, 2007, *Appendix II*; Andrew & Gudmundsson, *accepted, Appendix III*; Gudmundsson *et al*., 2008, *Appendix IV*). They can also affect other geological structures, such as fractures or fissures. These can be considered essentially as cracks, and their behaviour alters, and is controlled, according to the stress field created by the inclusion (Li & Chudnovsky, 1993).

Volcanoes as Inclusions

An inclusion can be either stiffer or softer than its matrix, meaning it has either a higher or lower Young’s modulus respectively. Where an inclusion contains a fluid only, it cannot be considered with elastic properties and is thus assigned a Young’s modulus of zero, and is regarded as a hole in two dimensions, or a cavity in three dimensions. In the case of volcanology, magma chambers (and thus central volcanoes) can be considered in simplified terms, and modelled, as holes. When the magma chamber is solidified as a pluton, it is considered as a stiff inclusion.

Where the inclusion is of a different material to the matrix and is subject to loading, it will develop a local stress differing to that of the surrounding material. The stress within the inclusion will be higher or lower than that in the surrounding material if the inclusion is stiffer or softer respectively (Savin, 1961; Nemat-Nasser & Hari, 1999; Jaeger *et al*., 2007; Andrew & Gudmundsson, *accepted, Appendix III*).

6.2 Mechanical Interaction

The mechanical interaction between the central volcanoes in Iceland has been a main theme of this study. Numerical modelling has highlighted the presence of clusters of volcanoes, mainly (but not exclusively) in the southwestern and northeastern ends of the EVZ. Clusters mean that the volcanoes are in sufficiently close proximity to each other to be able to interact.

The interaction between volcanoes can be through varying means, such as one volcano experiencing interference from a proximal volcano undergoing an eruption. It can also be through the sharing of dykes. The latter can occur when there are zones of tensile stress concentration, which encourage the emplacement of dykes. There must also be the correct orientation of the principal stresses running between the volcanoes to allow the dyke to be emplaced. This interaction and emplacement of dykes is limited by distance between the volcanoes, and also the favourable stress directions created by the spreading axis (Gudmundsson & Andrew, 2007, *Appendix II*; Andrew & Gudmundsson, *accepted, Appendix IIII;* Gudmundsson, *et al*., 2008, *Appendix IV*). Such interaction can be validated by eruptions such as the 1996 Gjalp eruption between the central volcanoes Grimsvötn and Bardarbunga (e.g. Gudmundsson *et al*., 1997)

It is not only the central volcanoes that interact. Other types of volcanoes as inclusions have also been studied here (Andrew & Gudmundsson, *accepted*, *Appendix III*). Models of hyaloclastite mountains, such as the Laki mountain (Figure 6.1), have shown that it affects the propagation of fractures and fissures. Similarly on a larger scale, the Torfajökull central volcano has effects on the propagation of the rift zone in the southwestern part of the EVZ (Andrew & Gudmundsson, *accepted, Appendix III*).

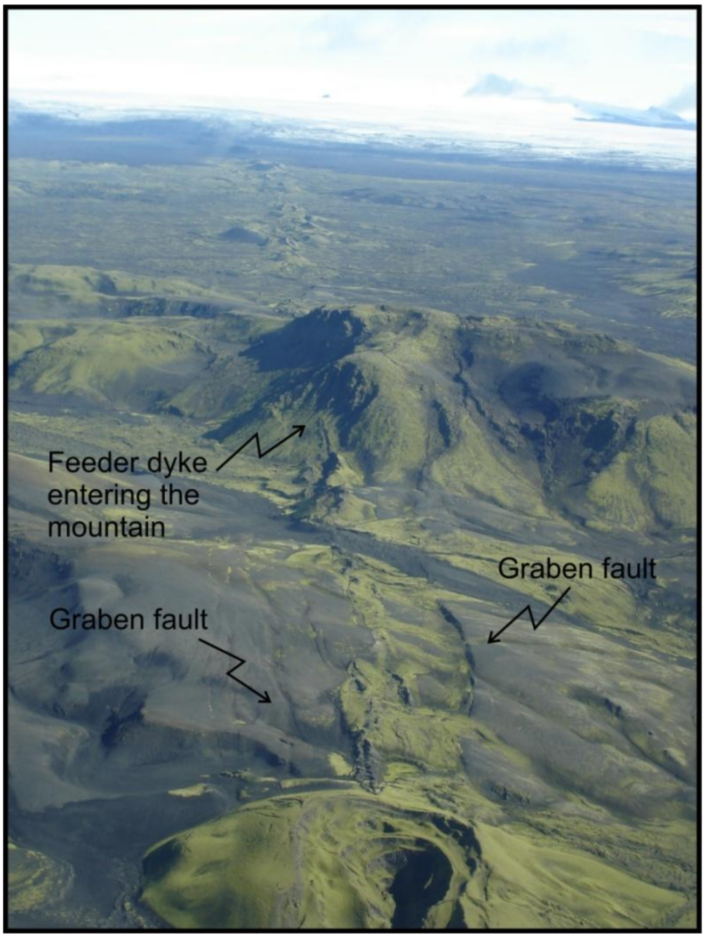

Figure 6.1: Photograph taken from air over Laki hyaloclastite mountain looking northeast along the fissure row towards Vatnajökull in distance. The faults of the graben are marked, as is the feeder dyke entering Laki mountain.

6.3 Mechanical interaction between active volcanoes in Iceland

We test the possibility of mechanical interaction between eight central volcanoes in the central part of the active Iceland rift zone. The average distance between the volcanoes is 30 km; all are thought to have shallow magma chambers, and many contain collapse calderas. We analysed many finite-element models with the volcanoes subject to a tensile stress of 5 MPa (equal to the maximum in situ tensile strength of the crust) in a direction parallel to the spreading vector, N105E. The results show zones between many nearby volcanoes where the tensile stresses exceed the in situ tensile strength of the crust. The results indicate that mechanical interaction between volcanoes in a pair, such as simultaneous dyke emplacement, seismogenic faulting, and deformation, may be common in this part of Iceland, in agreement with observations.

The full text of this paper is presented as *Appendix II* at the end of this study.

6.4 Volcanoes as elastic inclusions: their effects on the propagation of dykes, volcanic fissures, and volcanic zones in Iceland

We present three main numerical results. The first, using the hole model, shows the mechanical interaction between all the active central volcanoes in Iceland and, in particular, those forming the two main clusters at the north and south end of the East Volcanic Zone (EVZ). The strong indication of mechanical interaction through shared dykes and faults in the northern cluster of the EVZ is supported by observations. The second model, using a soft inclusion, shows that the Torfajökull central volcano, which contains the largest active caldera in Iceland, suppresses the spreading-generated tensile stress in its surroundings. We propose that this partly explains why the proper rift zone northeast of Torfajökull has not managed to propagate through the volcano. Apparently, Torfajökull tends to slow down the rate of southwest propagation of the rift-zone part of the EVZ. The third model, again using a soft inclusion, indicates how the lateral propagation of two segments of the 1783 Laki fissure became arrested in the slopes of the hyaloclastite mountain Laki.

The full text of this paper is presented as *Appendix III* at the end of this study.

6.5 Effects of dyke emplacement and plate pull on mechanical interaction between volcanic systems and central volcanoes in Iceland

The two principal ways by which rift-zone volcanic systems become loaded are (1) the magmatic overpressure induced by dykes, and (2) the plate "pull" associated with extension in the direction of the spreading vector (105°). This paper shows that both loading conditions give rise to mechanical interaction between volcanic systems in general, and their central volcanoes in particular.

Here we show that the magmatic overpressure of a regional dyke may reach tens of mega-pascals. We model the effects of simultaneous dyke injections, each dyke with an overpressure of 10 MPa, in the echelon systems on the Reykjanes Peninsula. The results indicate N-trending zones of high shear stress between the nearby ends of the volcanic systems, favouring strike-slip faulting. Geometrically similar shear-stress zones develop between the volcanic systems on the peninsula when acted on by a plate pull of 5 MPa in a direction parallel with the spreading vector. The results agree with the observation that there are many N-trending strike-slip faults on the Reykjanes Peninsula. When the same plate pull is applied to a cluster of 8 central volcanoes in Central Iceland, zones of high tensile stress develop between many of the volcanoes. These highly stressed zones encourage mechanical interaction between the volcanoes, such as simultaneous dyke emplacement and seismogenic faulting, as is supported by observational data.

The full text of this paper is presented as *Appendix IV* at the end of this study.

7 Volcanotectonics of the East Volcanic Zone

7.1 Introduction

As is discussed in section 2.2.1 the EVZ is the youngest section of the Neovolcanic Zone. The EVZ can be loosely divided into two sections, the northeastern half, above Torfajökull (Figure 7.1), where it shows classic rifting activity, similar to that of the NVZ and WVZ. Southwest of Torfajökull, the EVZ is considered a flank zone, where it is thought to be developing into a classical rift zone. The division of propagating and classic rift is defined by the differences in petrology at either end of the EVZ, the tholeiitic series in the northeastern end and alkali in the southwestern end (Figure 2.5). The slowing of this development into a rift zone has been found in this study (Andrew & Gudmundsson, *accepted*, *Appendix III*) to be at least partially attributed to the "buffer" effect of Torfajökull.

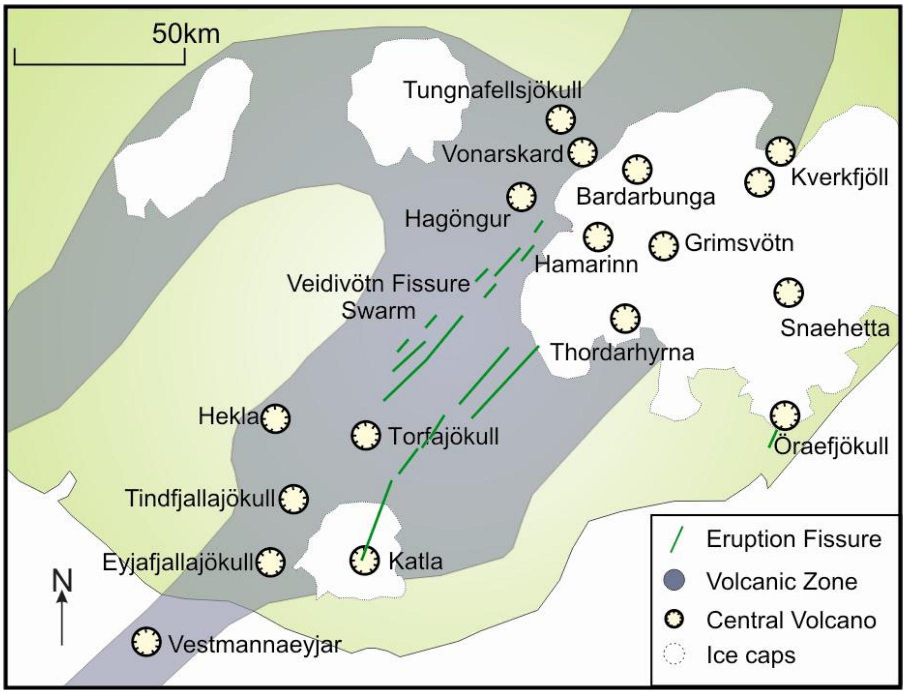

Figure 7.1: Map of EVZ showing the central volcanoes and the large fissures. The Veidivötn fissure swarm is also shown, with its proximity to Torfajökull central volcano. Modified after Johannesson & Saemundsson, 1998.

7.2 The Vestmannaeyjar Volcanic System

At the southwestern tip of this flank zone, lies the Vestmannaeyjar volcanic system (Figure 7.1). Within the system, the volcanic activity has been concentrating around the island of Heimaey, as can be seen by the five eruptions creating its form (Figure 7.2). Within the context of the system, this would appear to be a marked localisation of activity (Figure 7.3).

This concentration of activity, and the occurrence of more evolved lavas on Heimaey compared to the rest of the system, has led to the belief that the Vestmannaeyjar volcanic system is developing a central volcano (Mattsson & Hoskuldsson, 2003). This would mean a shallow magma chamber located under an edifice representing the focal point of activity in the system (Thordarson & Larsen, 2007). It has been suggested that the magmas of the 1973 Eldfell eruption on Heimaey and the 1963 eruption of Surtsey at the southwestern end of the system, were derived from the same magma batch. The Eldfell magma is thought to have resided in a shallow crustal magma chamber and fractionated for an additional 10 years from the Surtsey magma (Sigmarsson, 1996).

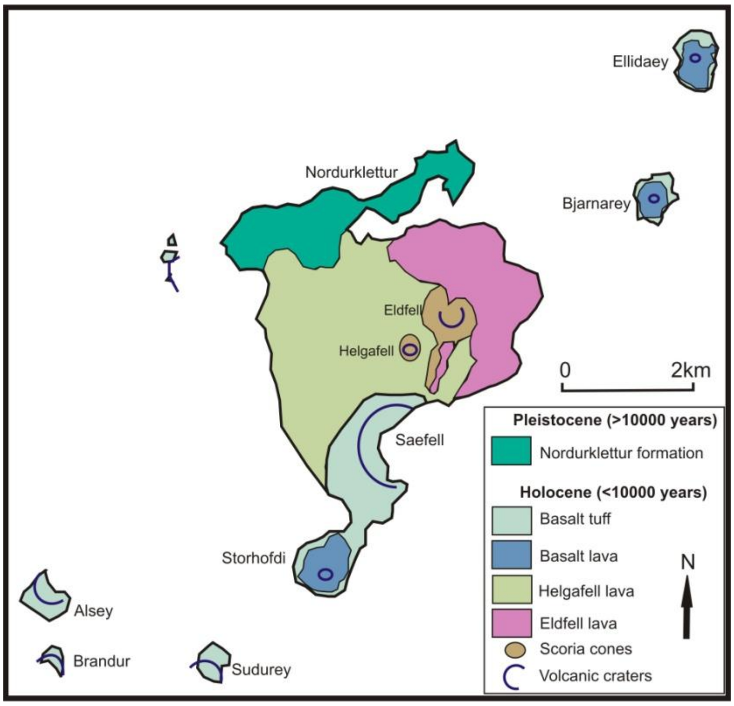

Figure 7.2: Geological map of Heimaey and surrounding subaerial eruption sites in the Vestmannaeyjar volcanic system. Modified after Jakobsson *et al*., 1973; Thordarson & Hoskuldsson, 2002.

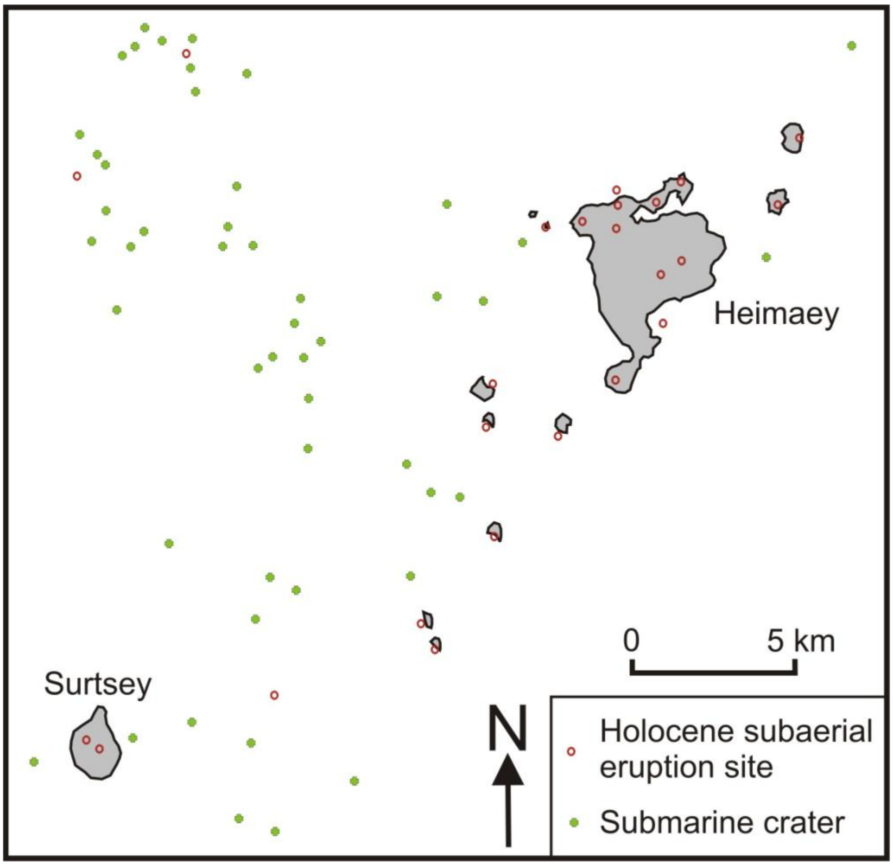

Figure 7.3: Map depicting the submarine and subaerial sites within the Vestmannaeyjar volcanic system. Modified after Mattsson & Hoskuldsson, 2003.

7.3 Flank Zone Models

Volcanic flank zones are thought to be rift zones in an early stage of formation. This is not only observed in Iceland, but also in other rift zone locations, such as the Galapagos spreading centre (Christie & Sinton, 1980). The most obvious example of this in Iceland is the EVZ in its gradual replacement of the WVZ, but it is also seen in the Öraefajökull-Snaefell Zone. The case of the Snaefellsnes Volcanic Zone is slightly different and not considered here.

As a flank zone is a developing structure, and therefore it should be possible to see a progression of stages of development along its length. In addition to this, there should be a corresponding diversity of magma composition and levels of productivity (Hards *et al*., 2000).

Hards *et al*. (2000) make a comparison between the EVZ and the propagating rift zones (flank zones) of the Galapagos spreading centre as studied by Christie and Sinton (1980). The model of Christie & Sinton (1980) shows that as the evolution of the rift zone proceeds, the magma supply rate gradually increases and the cooling rate of the magma chamber decreases correspondingly. The study by Hards *et al*. (2000), applying this to the EVZ is briefly presented here (Table 7.1), with examples according to the volcanoes and situation of the EVZ.

Whilst this model does not fit exactly to the EVZ, the general pattern of alkali basalt, transitional alkali basalt and tholeiite can be applied. Hards *et al*. (2000) note that the tectonic setting, that is the maturity of the volcanic zone, appears to be a controlling factor on petrological variations. They enhance this model further by observing a pattern explicable in terms of maturity of the rift zone and proximity to the mantle plume. This is evident in the composition of the basalts, the percentage of evolved magma erupted, the degree of fractionation and the stability of the magma chambers along the EVZ (Figure 7.4).

Composition & Diversity of erupted magmas	Examples from the EVZ
- lavas at rift tip unfractionated and alkali in composition - low degrees of melting and rapid cooling	Vestmannaeyjar = alkali
- area dominated by crystal ingestion and contamination processes - co-existence of primitive and highly fractionated (silicic) magmas	Katla = transitional alkali
- longer residence times - magma replenishment relatively infrequent – allows production of rhyolites	Torfajökull = rhyolite
- further increase in magma supply rates - system of stable shallow magma chambers - tholeiite basalt magma composition	Grimsvötn = tholeiite basalt

Table 7.1: Table showing the composition and diversity of the magmas of the evolving flank zone, with examples taken from the EVZ. Modified after Hards *et al*., 2000.

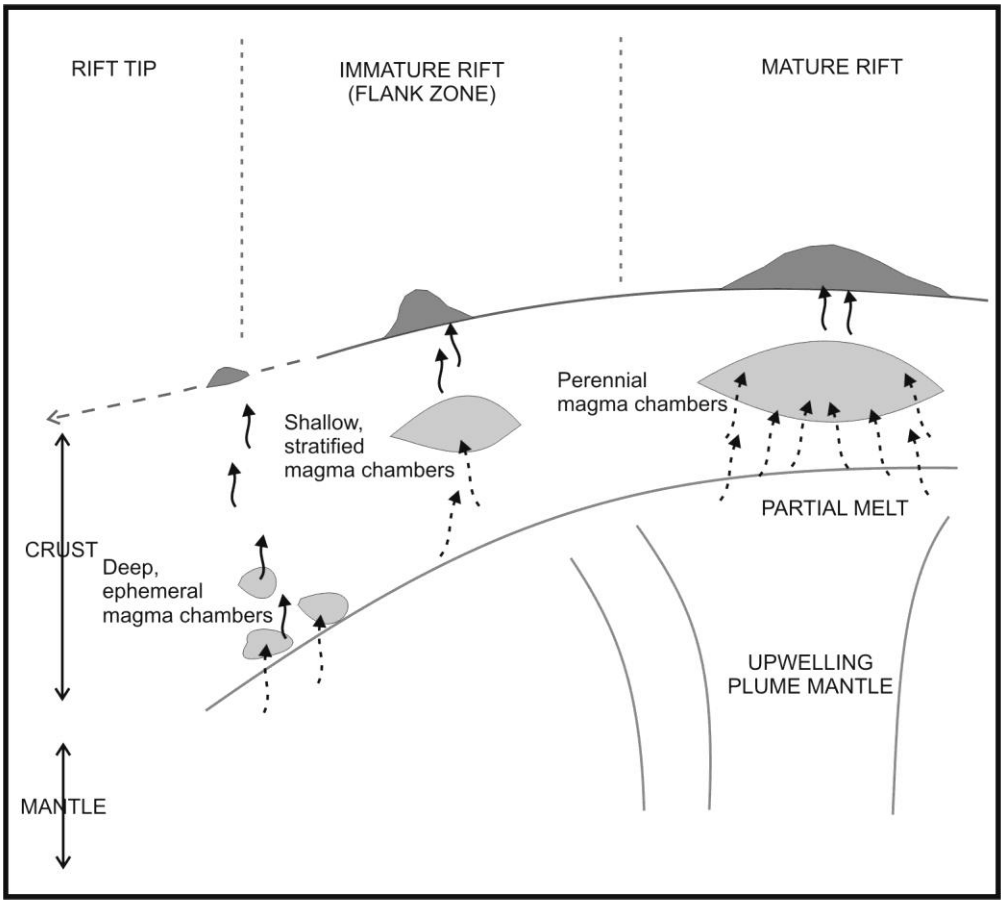

Figure 7.4: Schematic diagram showing relationship between the mantle plume, evolution of the rift zone, and the depth and type of magma chambers. Modified after Hards *et al*., 2000.

It should be noted that some authors argue that the difference in composition of the EVZ and associated differing tectonic regimes do not hold. Thirlwall ((2008) *pers. Comm.*) attributes the difference in composition to the increased depth of melting that is thicker crust, above the mantle plume. Deeper melting implies tholeiitic compositions. By comparison, in the Vestmannaeyjar the crust is thinner, and therefore shallower melting occurs, implying an alkaline composition.

7.4 Large Fissures

The large fissure eruptions have been discussed above (Section 2.3.2), noting their location almost solely in the EVZ (Figure 7.1), and in between the clusters of volcanoes in the northeastern and southwestern ends of this zone (Andrew & Gudmundsson, *accepted*, *Appendix III*). The presence of the large fissures is one of the many interesting characteristics of the EVZ, making it differ from the other parts of the Neovolcanic Zone. Some examples of these large fissures have been discussed in other parts of this study.

In order to understand the anomalous types of fissures it is important to first look at other fissures throughout Iceland and other volcanic areas. The majority of fissures in Iceland are evenly distributed throughout the Neovolcanic Zone (Figure 2.8). The orientation of these fissures is usually determined by the stresses created according to the spreading vector.

7.4.1 Creation of Fissures

A fissure is the surface expression of a dyke, that is, an injection from the magma chamber or reservoir. In its simplest form a fissure is simply a extension fracture (mode I), meaning a crack developed at right angles to the direction of maximum principal tension. The opening of the fracture/dyke is encouraged by the overpressure of the fluid (magma) inside, and the opening occurs during a rifting event within a volcanic system. It is the combination of the extensional tectonics and the overpressure which initiate the fissure. The overpressure of the magma is a function of buoyancy, overpressure within the magma chamber and the stress difference in the principal stresses. The condition for the rupture of the magma chamber or reservoir and the initiation of the dyke (or inclined sheet) is:

$$P_l + P_e = \sigma_3 + T_o$$

where P_l is the lithostatic stress or overburden pressure at the rupture site, P_e is the excess pressure at rupture, σ_3 is the minimum compressive or maximum tensile principal stress, and T_o is the local in situ tensile strength at the rupture site (Gudmundsson, 2006).

7.4.2 Flow in fissures

Upon opening, the fissure experiences possibly the highest mass flux, that is the effusion rate of the magma, and the fissure elongates in its early stages (Giberti & Wilson, 1990; Ida, 1992). The elongation of the fissure means that it propagates laterally, with continuous flow along its length. Further into the eruption, however, the flow becomes localised into individual fingers and then vents (Figure 7.5), which allow the prolonged activity of the eruption (Ida, 1992; Wylie *et al.*, 1999).

It is worth noting that lava shield eruptions start in essentially the same way, with a small fissure eruption. However, at the stage of localisation of flow, a single vent dominates and grows to form the lava shield (Rossi, 1996). The key variable in the creation of a fissure or a shield eruption (aside from the source of the magma in some cases) is thus the average effusion rate (Sinton *et al.*, 2005).

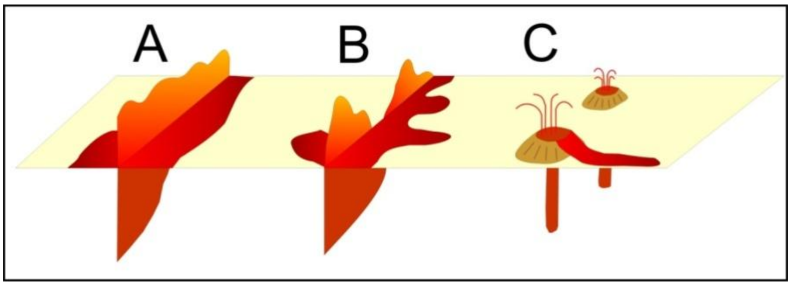

Figure 7.5: Schematic diagram of flow in a fissure: A) the fissure opens and widens with a flux of flow at the start along the length of the fissure, or fissure segment; B) the flow in the fissure becomes localised into fingers; C) the localised fingers eventually form vents and cones. Modified after Ida, 1992.

7.4.3 Fissures in the EVZ

Volcanic fissures in the EVZ are much larger than in other areas, for example the Laki and Eldgja eruptions were 27 km and 57 km long respectively. The locations, orientations and the initiation of the fissures can be explained by the reasons outlined above. However, the reason for the size of the fissures has so far not been explained.

As discussed above, the EVZ differs from the other parts of the Neovolcanic Zone in several ways, one of which being that it is thought to be younger. The characteristics of the crust will alter slightly with age, thus as a younger segment of the Neovolcanic Zone, the EVZ's characteristics will differ. As a rift zone becomes more mature, it experiences more fracturing due to the rifting and volcanic activity. It is possible that the EVZ is stiffer than the other segments of the Neovolcanic Zone as, due to its age, it has not experienced as much activity. This could then in turn mean that when volcanic fissure eruptions begin they have the opportunity to behave in a different way.

If a fissure opens in a stiffer homogeneous crust, it would be more likely to stay open for a longer period of time than in a softer crust. This follows because in a stiffer crust, the crack is able to stay open along its entire length, despite to the variations in the overpressure of the magma. In a soft material, the overpressure means that the magma forces itself outwards in places, localises and creates vents, whilst in a stiffer material this process is harder, due to the resistance of the crust. Therefore the pressure and thus flow, is continuous along the length of the crack for a longer period, and the eruption is sustained for longer and at greater widths (Gudmundsson *et al.*, 2007).

8 Hazards

8.1 Introduction

When looking at any aspect of volcanoes in any location around the world, it is important to look at the matter of hazards and risk. This is especially important when looking at volcanoes in populated areas. Hazards can be defined as extreme geological events capable of causing disaster. Such events and their effects are fundamentally determined by their location, magnitude, timing and frequency (Alexander, 2000). In the context of a volcano, the hazard is the probability of a potentially dangerous volcanic process occurring. A volcanic eruption can present a wide range of hazards, some direct, such as lava flows destroying property or being in the path of a pyroclastic flow, others are indirect such as damage to crops from climate change (Parfitt & Wilson, 2008).

The concept of volcanic risk takes in a much broader area, involving the actual cost of the volcanic event. The cost is not limited to buildings and infrastructure affected, but also includes human lives. Thus risk can be defined as the probability of loss of human lives, property and productive capacity in a region subject to a volcanic hazard.

Volcanic hazards and risk are normally calculated and managed by direct monitoring of the volcano. This study is not involved in direct monitoring of volcanoes, or the construction of hazard and risk maps. However, the fundamental idea of hazards and risk analysis is understanding the processes by which the hazard, i.e. the volcanoes, work and erupt.

Volcanology as a subject still holds a great deal of unknowns. Indeed any modelling of volcanoes by any means, including the numerical models used in this study, is extremely hard to validate. This study attempts to model and validate with data from the field, and from patterns in previous eruptions. As a result, the study gains a better understanding of some of the volcano-tectonic processes operating in Iceland, and contributes to reducing hazard and risk.

8.2 Volcanoes and Ice

Geographically speaking, Iceland presents some features which create different types of volcanic hazard. One of these features is the presence of ice bodies that are spread throughout the country in various sizes. There is ice covering many of the central volcanoes in Iceland (Figure 2.8, Figure 7.1), notably including Katla and Grimsvötn in the EVZ. These two volcanoes are under the two largest ice bodies, and are also amongst the most historically active volcanoes in the country (Thordarson & Larsen, 2007).

When a volcano erupts underneath a body of ice, the ice is melted creating a flux of meltwater, which flows out from the edge of the ice cap. This is known as a jökulhlaup, and Iceland is perhaps the best known location for the phenomena (Francis, 2001).

Vatnajökull: Gjalp

The 1996 subglacial eruption of Gjalp under the Vatnajökull ice cap melted 3 km^3 of ice (Gudmundsson *et al.*, 1997). This volume of melted ice increased to 4 km^3 in the subsequent six weeks of the eruption, and drained southwards towards Grimsvötn and out to the edge of the ice body at roughly the same rate that it was formed. It then burst from the southern edge of the ice body with a flow rate equivalent to that of the Amazon River, ~200000 m^3s^{-1} (Francis, 2001). The result of this flood was the destruction of infrastructure along the edge of this section of the ice cap (Figure 8.1), all the way down to the coast. Deposition led to a marked increase in the size of the sandur plain, and the road was washed out completely (Figure 8.1).

Figure 8.1: Photograph of remains of bridge damaged by the jökulhlaup associated with the 1996 Gjalp eruption under Vatnajökull, see picnic bench for scale.

As a result of the repeated jökulhlaups in this area throughout historical time, and the sandur plain (Figure 2.4), there are not many settlements in the area immediately surrounding this part of the ice cap. However, for those that are present, the infrastructure, and the tourist centre and associated activities at Vatnajökull, mean that the hazard and risk posed here are relevant.

Myrdalsjökull: Katla

Another well documented location of jökulhlaups, also in the EVZ, is due to eruptions of Katla underneath Myrdalsjökull. These are the main hazards associated with the Katla volcano, leading to studies into their magnitude and recurrence times in order to obtain a hazard assessment (Eilasson *et al*., 2006). Jökulhlaups from Katla directly affect local communities, such as the town of Vik on the south coast. This has led to a network of monitoring of the volcano and a subsequent hazard and emergency response plan for the local population.

8.3 Shield Volcanoes

The study here (Andrew & Gudmundsson, 2007, *Appendix I*) links the formation of Holocene lava shields in Iceland to the retreat of the Weichselian ice sheet. The study does not explain the existence of other lava shields evident in the lava pile, dating back to the Tertiary and Pleistocene. This indicates that the existence of lava shield and their formation is still not entirely explained or understood. However, as discussed in this study (Andrew & Gudmundsson, 2007, *Appendix I*) it would appear that this is the best explanation for their formation, and thus they can be cautiously regarded as not being a significant hazard today, though not ruled out entirely.

8.4 Central Volcanoes

The work on mechanical interaction between central volcanoes in Iceland (Gudmundsson & Andrew, 2007, *Appendix II*; Gudmundsson *et al*., 2008, *Appendix IV*; Andrew & Gudmundsson, *accepted, Appendix III*) has allowed an insight into potential links between them. The position of some of these central volcanoes under ice bodies is discussed above and presents perhaps the highest risk. However, links between the volcanoes, and the volcanic systems may allow further insights into behavioural patterns. It may also allow better preparation, by forewarned knowledge, of simultaneous activity of volcanoes in the future.

8.5 Large Fissures

The question of large fissures has been discussed in detail in this study. As their creation and the processes surrounding them are still not understood, they remain unpredictable. Indeed, even though their locations seem to be confined to the EVZ at present, the magnitude of the events poses a risk on both a local, national and international scale.

Laki and Eldgja

The eruptions of Laki and Eldgja are well documented as having very far reaches in their effects. On a proximal, that is a local or even national, scale the Laki eruption in 1783 is thought to have killed roughly 1/3 of the population of Iceland (Thorarinsson, 1979; Ogilvie, 1986). This not only included human fatalities due to contaminated air, but also due to widespread famine as the livestock of the country were heavily affected by contaminated and sometimes inedible vegetation.

The location of Iceland in high latitude means the height of the tropopause in the atmosphere above the country is relatively very low. A fissure eruption such as Laki or Eldgja, with long lived degassing of sulphur, means that gases enter into the atmospheric air circulation systems and are thus able to circumnavigate the globe at this latitude.

Studies of weather in Europe in the 1780's by Kington (1980) have allowed authors to study the direction of a cloud of gas from the Laki fissure eruption as it travelled across northern Europe (e.g. Grattan & Brayshay, 1995; Grattan & Sadler, 1999). This enables a correlation with contemporary records, and shows the international or distal effects of such an eruption (Grattan & Charman, 1994; Grattan & Brayshay, 1995). The Laki eruption has been held accountable for peaks in death rates in Britain, contributing to the French Revolution, and even failed rice crops in the Far East (Brayshay & Grattan, 1999; Scarth, 1999).

So called mass-loading of the atmosphere by these large fissure eruptions also causes cooling of the atmosphere. Studies of the Eldgja eruption show that it was possibly the largest volcanic pollutant of recent history. However, its climatic effects are not thought to have surpassed those of Laki, as Eldgja was a much more prolonged eruption, drawing the sulphur emissions out over several years (Thordarson *et al*., 2001). The eruption columns in the Eldgja eruption allowed the injection of gases into the upper troposphere and lower stratosphere, where they could be dispersed by the westerly jet stream (Lamb, 1970). The mass-loading is thought to have taken ~4-6 weeks to circumnavigate the northern hemisphere and reach Greenland, where the gases are traceable in ice cores (Thordarson *et al*., 2001). The Eldgja eruption is thought to have potentially caused a cooling of approximately 1.2^{o}c (Thordarson *et al*., 2001).

A large fissure eruption today?

As there have been no such large fissure eruptions in modern times, the hazards and risk posed by such an eruption are not clear. Any ideas on the effects of an eruption are therefore mostly speculation. There are a few things that are clear however, first being that any volcanic eruption in Iceland effects air traffic within its vicinity. This means that flights from North America to Europe and vice versa will be disrupted. Though this is not so much a hazard or risk, it is a disruption and an important secondary effect of an eruption.

Mortality in Europe due to reduced air quality is less of a problem in modern society, as for example medicine has come a long way since 1783. Respiratory problems may be exacerbated, but not necessarily proving fatal. The effects of reduced air quality on agriculture may be more of a problem, and the knock-on effect on the economy by increasing food costs and indirectly affecting trade between countries.

On a local scale, tourism in Iceland is a fast-growing economy. The hazard posed to the population and tourists directly by proximity to an eruption is something that should be considered. In turn, the effect on the economy of Iceland due to loss of earning etc. from an eruption of this kind could cause large problems for Iceland.

9 Conclusions

- There is a clear increase in volcanic activity in the late Pleistocene and early Holocene. This has been linked to the retreat of the ice if the last glaciation. Numerical models have examined the mechanical aspect of this deglaciation. They have shown that under glacial compression the compressive stresses extend to the magma reservoir and upper mantle, suppressing the magma accumulation there.
  - As the ice retreats, tensile stresses around the magma reservoir lead to increased porosity and magma content. Dykes are able to inject to reach the surface under the ice, forming subglacial volcanoes; then progressively the margins of the ice forming shield volcanoes. At this time, directly above the magma reservoir, ice compression allows magma to inject to a neutral surface. At this point the dykes are arrested by contrasting mechanical properties and form a shallow crustal magma chamber.
  - Continued ice retreat gradually reduces the compression and finally encourages active volcanism on central volcanoes.
  - These processes have created a situation with subglacial volcanoes and shield volcanoes on the margins of, or outside the volcanic systems. Central volcanoes and fissure swarms are located within the volcanic systems and define them.
- Volcanoes can be considered mechanically as elastic inclusions either softer (with a lower Young’s modulus) or stiffer (with a higher Young’s modulus) than the surround host-rock matrix. As inclusions they can be modelled to evaluate their mechanical interaction.
  - Central volcanoes have been found to mechanically interact with each other, where they are in sufficiently close proximity. This is particularly strong in “clusters” such in the Vatnajökull region in the northeast of the EVZ. They are able to share dykes where tensile stresses exceed the in situ tensile strength of the crust. The volcanoes are able interact through dyke emplacement, seismogenic faulting, and deformation. This has been validated by observations of eruptions.
  - Volcanoes as inclusions can also interact with other geological structures such as faults and fissures. This is the case of the Laki hyaloclastite mountain “buffering” the propagation of the Laki fissure eruption. This has been validated by field observations. On a larger scale, it is proposed that the Torfajökull central volcano uses this “buffering” process to slow the propagation of the EVZ from rifting to propagating.
- The location of large fissure eruptions such as Laki or Eldgja has been briefly examined. The differences of the EVZ as a propagating rift or flank zone were studied, and it is proposed here that the differing properties of the EVZ due to its relative youth could explain the magnitude of these eruptions, in mechanical terms.
- All these findings were put into the context of hazards and risk, concluding that the risk of subglacial eruptions and the resultant jökulhlaups poses the most pressing risk. However the hazard posed by the large fissure eruptions is of a much larger scale and more studies are needed to gain a better understanding of these processes.

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11 Appendix I

Distribution, structure, and formation of Holocene lava shields in Iceland

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Abstract

Lava shields in Iceland are monogenic shield volcanoes mostly composed of primitive basalts, picrite and olivine tholeiite. The Holocene shields are restricted in space and time: in space they are essentially confined to the North and West Volcanic Zones; in time they are essentially confined to the early part of the Holocene. The volcanic zones are characterised by volcanic systems, that is, 5-20 km wide and 40-150 km long swarms of tectonic fractures and volcanic fissures where most of the Holocene volcanotectonic activity takes place. However, most lava shields are not located inside the volcanic systems but rather at their margins or in between systems. In addition to the lavas of the shields being more primitive those issued from nearby volcanic fissures, the average volume of a shield is an order of a magnitude larger than that of the lava from a fissure. Here we present the results of new field observations of Holocene shields and provide numerical models to explain their location, time of formation, primitive composition, and large volumes. We made models with varying ice-sheet size and thickness (glacial load), the ice resting on a mechanically layered crust, and studied the stress effects that the load would have on a double magma chamber, that is, a small, shallow crustal chamber at 3 km depth and a deep-seated reservoir at 20 km depth (the base of the crust). Such a pair of chambers is typical for volcanic systems and associated central volcanoes (composite volcanoes) in Iceland. For an ice sheet covering an entire volcanic zone or more, that is, 100 km or wider, the ice-induced compressive stress extends to the deep-seated reservoir and into the upper mantle. Consequently, such a loading suppresses magma accumulation in the reservoir and associated volcanism. During the late-glacial period, when the ice sheet is only 20 km wide, the glacial load generates tensile stresses around the deep-seated reservoir, increases its fracture porosity and magma content, and extends the reservoir laterally and vertically into the upper mantle. Consequently, when the lava shields (and, somewhat earlier, the tablemountains) were erupted, much more melt or magma was available to feed a single eruption than during the later part of the Holocene. And because of the greater vertical extent of the reservoir, this magma tended to be hotter and more primitive than that issued in later-formed fissure eruptions. Also, the stress field generated at the end of the glacial and in the early Holocene favoured dyke injections at the marginal parts of, or in between, the volcanic systems, thereby explaining the location of the lava shields.

Keywords: lava shields, fissure eruptions, magma chambers, crustal stresses, volcanic hazard

1. Introduction

Holocene lava shields (monogenic shield volcanoes) are common in Iceland (Fig. 1). They range in volume from a fraction of a cubic kilometre to more than twenty cubic kilometres, and in composition from very primitive picrites to olivine tholeiites. Their morphology and eruptive mechanisms have been discussed by Rossi (1996, 1997), who also distinguished between the typical volumes of the picrite shields (normally very small) and the olivine tholeiite shields (some of which are large).

The Holocene lava shields are restricted in time and space. As regards time, most Holocene lava shields formed in early postglacial time, or some 5000-10000 years ago (Rossi, 1996). More specifically, apart from the shield on top of the island of Surtsey, the youngest of the Vestmannaeyjar Islands (Fig. 1), there are hardly any Holocene shields in Iceland younger than about 3500 years (Thorarinsson et al., 1959). As regards space, most of the lava shields occur within two bands in the West and North Volcanic Zones (Fig.1). There are no shields in the East Volcanic Zone apart from the island of Surtsey.
As regards distribution, a remarkable feature is that the Holocene lava shields are commonly outside or at the margins of the nearby volcanic systems (Fig. 1). A volcanic system can be defined using petrological characteristics (Jakobsson et al., 1978) or tectonic characteristics (Saemundsson, 1978), or both. The Holocene systems are analogous to the regional swarms of dykes and faults, identified by Walker (1960), that are exposed in the Tertiary and Pleistocene lava pile. Volcanic systems are essentially swarms of tectonic fractures and basalt volcanoes formed as a result of plate-pull (Gudmundsson, 2000). There are some 30 Holocene volcanic systems within the active volcanic zones (Fig. 1). A volcanic system has a typical lifetime of 0.5-1.0 million years (Thordarson and Larsen, 2007).

Since the volcanic systems are the focus of Holocene volcanic and tectonic activity, the fact that the Holocene shields commonly occur outside them indicates that the shields are related to different tectonic regimes. It has been suggested that many Holocene lava shields in Iceland erupted under special post-glacial tectonic conditions (Gudmundsson, 1986; Sigvaldason et al., 1992; Schiellurup, 1995; Bourgeois et al., 1998; Maclennan et al., 2002). Whilst these post-glacial tectonic conditions may not apply at present, the formation of a lava shield must be taken into account when looking at volcanic hazards in Iceland.

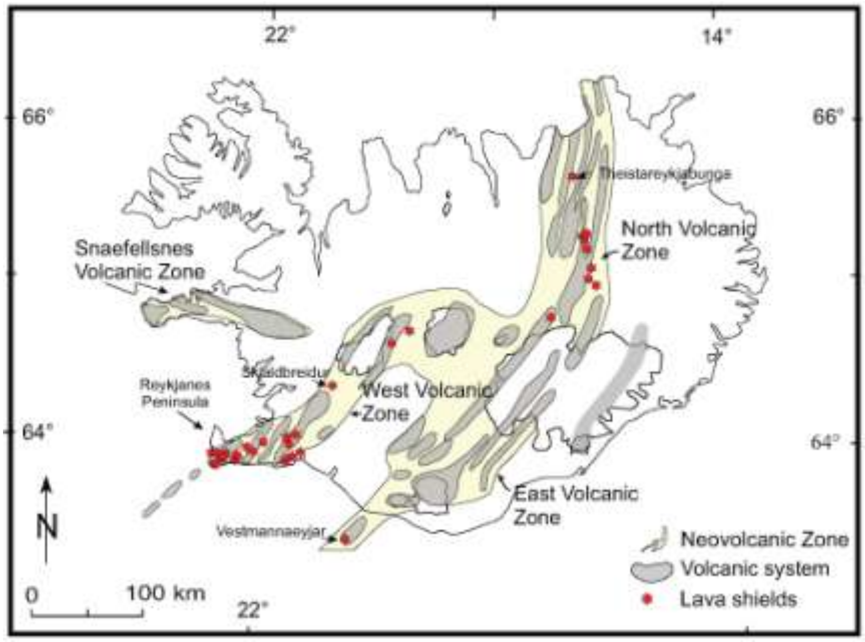

Fig. 1. Map of Iceland with showing the locations of the main Holocene lava shields (red stars) within the active volcanic zones (yellow) as well as the main volcanic systems (grey). Except for the lava shield on top of the island of Surtsey in the Vestmannaeyjar Volcanic Systems (cf. Fig. 2), all the Holocene shields are located in the North and West Volcanic Zones, their highest density being on the Reykjanes Peninsula. Also shown are the locations of some of the largest lava shields in Iceland: Theistareykjabunga in the North Volcanic Zone and Skjaldbreidur in the West Volcanic Zone, as well as the Vestmannaeyjar Volcanic System (modified from Gudmundsson, 2000).

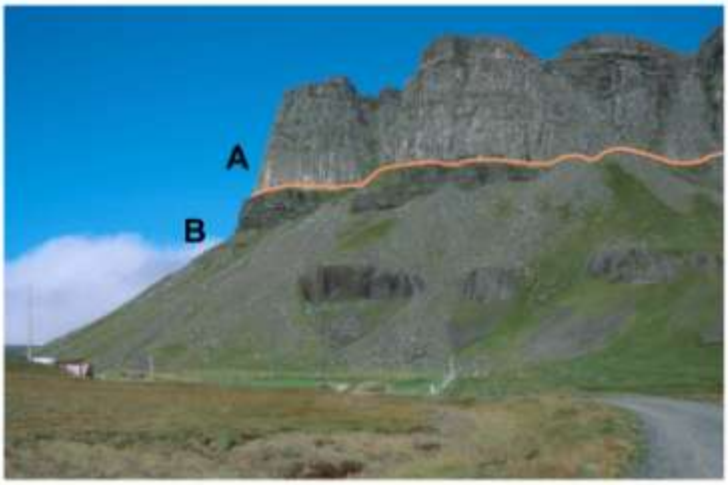

Fig. 2. Parts of two Pleistocene lava shields in the mountain Thyrill in the fjord Hvalfjördur in West Iceland. The contact between the shields is indicated (pink line). A, the upper shield; B, the lower shield. Both are composed of numerous flow units, such as are indicated in Figs. 5 and 6.

The principal aim of this paper is to look at the different types of Holocene lava shields in Iceland, and the tectonic conditions necessary for their formation. In particular, we provide conceptual, analytical and numerical models whose aim is to explain, first, why most shields occurred in the early part of the Holocene (and subglacial tablemountains of similar volumes and composition just before that, at the end of Weischelian); second, why volumes of individual eruptions at this stage were, on average, much larger than in the later parts of the Holocene; and, third, why most of the shields occur at the margins, or in between, the volcanic systems rather than inside them. A second aim is to use the tectonic-modelling results to discuss briefly the hazards associated with potential lava-shield eruptions in Iceland.

2. Lava shields in Iceland

2.1 General

The shield volcanoes of Iceland are commonly referred to as compound lava shields (Walker, 1971; Jakobsson et al., 1978). This term is used here so as to avoid confusion with the much larger, polygenetic, shield volcanoes that occur in areas such as the islands of Hawaii and Galapagos. Polygenic shield volcanoes such as these are, in the terminology used here, central volcanoes. In this paper the term lava shield (or just shield) is thus used for all the monogenic picrite and olivine-tholeiite shields in Iceland.

Lava shields are known throughout the Tertiary and Pleistocene lava piles (Fig. 2), but they are most conspicuous in the Holocene parts of the Neovolcanic Zone (Figs. 1, 3). The Neovolcanic Zone is traditionally defined as that part of Iceland covered by rocks belonging to the Brunhes magnetic epoch, that is, younger than 0.8 Ma. To avoid confusion, it should be pointed out that the Neovolcanic Zone of Iceland corresponds to the magnetic plate boundary at mid-ocean ridges where, by contrast, the “neovolcanic zone” is of Holocene age (Macdonald, 1982).

The Neovolcanic Zone of Iceland consists of three main segments or subzones (Fig. 1): the West Volcanic Zone (WVZ), East Volcanic Zone (EVZ), and the North Volcanic Zone (NVZ). These are all rift zones, that is, subject to pull as the Eurasian and North-American plates move apart, except for the soutwestern part of the EVZ, which is regarded as a propagating rift south to the island of Surtsey, the southernmost part of the Vestmannaeyjar Volcanic System (Fig. 1).

Fig. 3. View east, the lava shield Skjaldbreidur in the West Volcanic Zone (located in Fig. 1). Photograph by O. Ingolfsson.

The Holocene lava shields were formed mainly during the early postglacial times when the volcanic production rate was at its peak (Rossi, 1996), and all are thought to be older than 3500 years (Thorarinsson et al., 1959). Nearly all the Holocene lava shields are located in West Volcanic Zone, mainly on the Reykjanes Peninsula, and in the North Volcanic Zone (Fig. 1). The only shield that is located outside the WVZ and the NVZ, and thus in the EVZ, is the one formed during the 1963-1967 eruption of Surtsey in the volcanic system of the Vestmannaeyjar Islands (Fig. 1).

2.2. Shield geometry and lava types

The geometry and structure of a lava shield is well illustrated by that of the best-known shield in Iceland, Skjaldbreidur, located in the central part of the WVZ (Figs. 1, 3). The general slope of the flanks of the shield are 5-8° (Fig. 4), but they are steeper near the central crater which is 300-400 m in diameter and 60 m deep. Skjaldbreidur is considered the largest Holocene shield in Iceland, with an estimated volume of 25 km^3 (M.T. Gudmundsson et al., 2001). This may, however, be an underestimate since a part of the lava-flow field of Skjaldbreidur is buried by younger fissure eruptions.

The pahoehoe lava flows that make up a typical lava shield are of two main types: shelly type and P-type (Rossi, 1997). The shelly type is mostly vesicular and forms flow units (lobes), many of which become hollow when the lava drains out below the solidified lava crust. The shelly type flow units are generally thin. For example, Rossi (1996) calculated the median thickness of 55 flow units in a single shield as about 0.2 m. There are three main varieties of this type. (1) The amoeboid variety (Swanson 1973) consists of fragile slabs that are easily broken under the foot. (2) The sheet-flow variety is originally highly vesicular, but on loosing gas it becomes fairly dense and forms thicker units. (3) The slab-pahoehoe variety (Kilburn, 1993) is characterised by buckled and overturned slabs and formed by stretching and tilting of the slabs during flow of the lava.

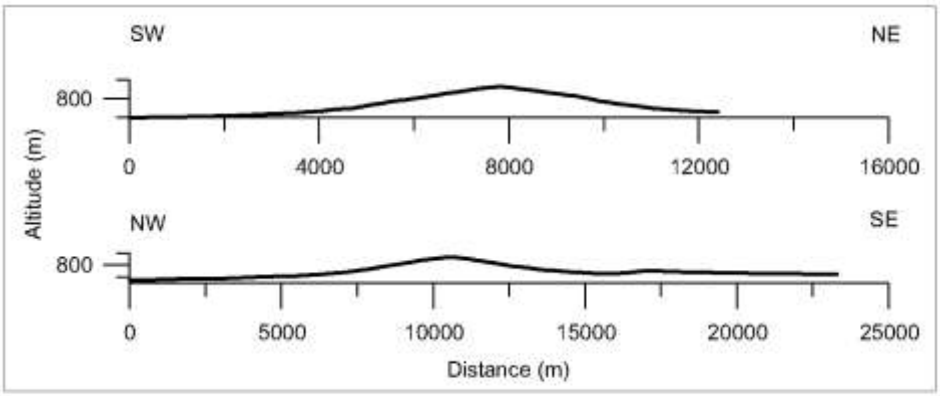

Fig. 4. Topographic profiles of the lava shield Skjaldbreidur in the West Volcanic Zone (located in Fig. 1).

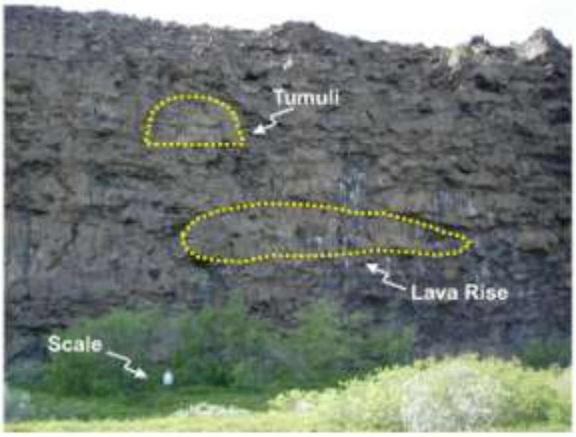

Fig. 5. Internal structure of the lava shield Theistareykjabunga in the North Volcanic Zone (located in Fig. 1) as seen in the 100-m-deep canyon of Asbyrgi. The main part of the vertical wall is composed of pahoehoe flow units, mostly 0.2-2 m thick, some of which have developed tumuli and lava rises (cf. Fig. 7). For further description of these structures see Rossi (1996, 1997) and Rossi and Gudmundsson (1996). View west, the person (indicated by an arrow) provides a scale.

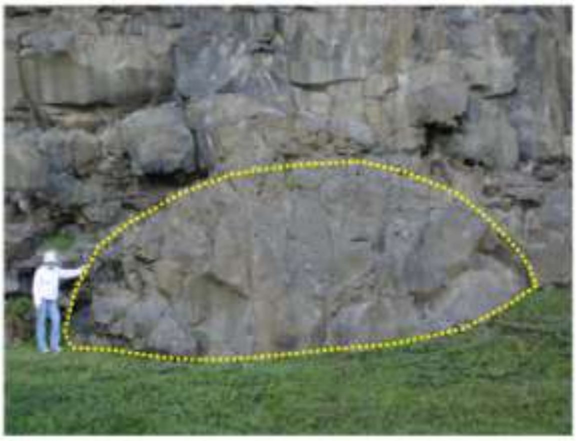

Fig. 6. Typical tumuli in the same lava shield and locality as in Fig. 5. This cross-section through a tumuli may be compared with the lava rise in Fig. 7. View west, the person provides a scale.

The P-type pahoehoe (Figs. 5, 6) is denser and thicker than the shelly type. For example, Rossi (1996) calculated the median thickness of 67 flow units of P-type lava in a single shield as about 0.6 m. The average density of the P-type lava is around 2400 kg m^{-3}, but may be as little as 1000 kg m^{-3} for some shelly-type lavas. The P-type is found in all lava flow fields that derive their lavas through lava tubes. The comparatively high density is partly due to long travel in the tubes during which the magma looses gas (Swanson 1973). Pipe vesicles are common in the P-type lavas, but absent from the shelly type (Walker, 1987).

A typical lava shield such as Skjaldbreidur (Figs. 3, 4) consists of two main morphological units: a central cone and a lava apron (Rossi, 1996, 1997). The cone is thought to be generally roughly symmetrical; the apron roughly a circular shape, though affected by topography (Rossi, 1996). The cone is made primarily of shelly-type pahoehoe lava flows. At greater distances from the central crater, however, the lavas change into P-type flows (Fig. 5). The lava aprons are mostly composed of P-type flows, delivered through an extensive network of lava tubes. The tube system commonly delivers magma from the upper part of the lava lake in the central crater of the shield. Some tubes, however, transport denser magma from the deeper parts of the lava lake or from parts of the conduit that supplies magma to the lava lake. As the shield develops there are occasional voluminous overflows from the central crater, as well as rootless outpourings on the flanks of the shield, that give rise to aa lava flows.

Fig. 7. View southwest, part of a lava rise in a Holocene pahoehoe lava flow at the north shore of Lake Myvatn in the North Volcanic Zone. The person provides a scale.

Among the most prominent features of a typical lava shield are tumuli and lava rises (Rossi, 1997; Rossi and Gudmundsson, 1996).

These features are commonly easily recognised in the walls of eroded shields (Figs. 5, 6), and characterise their surfaces (Fig. 7). On the Icelandic shields three types of tumuli are recognised: (1) lava-coated tumuli, (2) upper-slope tumuli, and (3) flow-lobe tumuli. Flow-lobe tumuli are common in the medial and distal parts of pahoehoe flow fields, particularly on large shields such as Skjaldbreidur, whereas lava-coated tumuli and upper-slope tumuli types are more common in the proximal parts of the flow fields. At the margins of the flow fields, flow-lobe tumuli gradually develop into the larger inflation structures named lava rises (Figs. 5, 7). Flow-lobe tumuli and lava rises are both generated by inflation of the lava crust as a result of magmatic overpressure in the lava core (Rossi and Gudmundsson, 1996).

2.3 Formation and composition

Icelandic lava shields are the result of single eruptions, some of which may have lasted for several years with periods of quiescence. It is likely that the formation of lava shields follows the following general pattern during each eruption. The eruption probably begins with lava being issued from a fissure. Subsequently, however, the eruption becomes concentrated at several vents, which may generate several overlapping, small shields. Gradually, the eruption becomes focused on the one vent that generates the main shield which buries the overlapping smaller shields (Gudmundsson, 2000). The main shield subsequently grows through lava flows (predominantly pahoehoe) that spread rapidly from a central crater and other vents on the flanks of the volcano. In this way the final shield is made up of layers of thin lava flow units (0.2 – 2m thick), as can be seen in exposures in some areas of Iceland (Fig. 5). The shield formation may stop at any stage during this scenario, so that some very small overlapping shields may, in fact, be generated in a single eruption from a segmented and offset volcanic fissure.

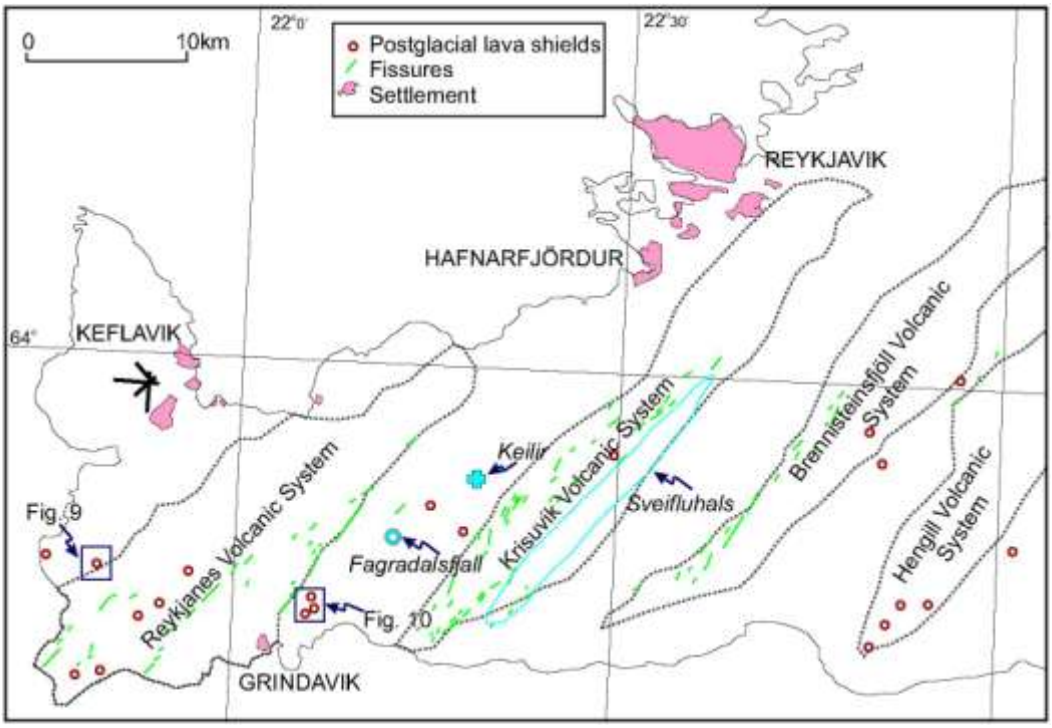

Fig. 8. Holocene volcanic systems, lava shields and volcanic fissures on the Reykjanes Peninsula (located in Fig. 1). Many lava shields occur near the margins of, or in between, the volcanic systems. Also located are the tablemountain Fagradalsfjall, the hyaloclastite ridge Sveifluhals, and the hyaloclastite cone Keilir as well as the capital of Iceland, Reykjavik, the Keflavik Airport, and the sites of Figures 9 and 10. Data from Jakobsson et al. (1978), Saemundsson and Einarsson (1980), and Gudmundsson (1986).

In Iceland the lava shields are mainly of two compositions, olivine tholeiites basalt and picrite basalt, both of which are primitive basalts. The type of magma is determined by the local pressure and temperature conditions, as well as the composition of the rock undergoing partial melting (Jakobsson et al., 1978). Rossi (1996) documented some 40 Holocene lava shields in Iceland. He made a detailed study of 31 Holocene lava shields and found them all to be monogenetic. Of these 24 were found to be of olivine tholeiite and 7 of picrite, indicating that the conditions for formation of olivine tholeiite shields are more commonly satisfied.

2.4 Shields on the Reykjanes Peninsula

The Reykjanes Peninsula which forms the southwestern part of the WVZ has the greatest density of Holocene lava shields in Iceland (Figs. 1, 8, 9, 10). The peninsula is also a part within the Neovolcanic Zone where the relationships in time and space between subglacial hyaloclastite mountains, lava shields, and volcanic fissures can be studied in great detail. Accordingly, in this study we paid particular attention to the volcanotectonic conditions on the Reykjanes Peninsula and how they are reflected in its Holocene volcanism.

The Holocene volcanism on the Reykjanes Peninsula is characterised by volcanic shields and volcanic fissures or crater rows (Figs. 8, 9, 10). The corresponding volcanic structures formed in subglacial eruptions are hyaloclastite ridges and tablemountains (Fig. 11). Hyaloclastite ridges form during subglacial fissure eruptions, as was confirmed during the 1996 Gjalp eruption in the Vatnajökull ice sheet (M.T. Gudmundsson et al., 1997). The mountain Sveifluhals is an example of a hyaloclastite ridge (Fig. 8). Short-lived subglacial fissure eruptions confined to a single conduit produce hyaloclastite cones, a well-known example being the mountains Thorbjörn (Fig. 10) and Keilir (Fig. 8). Examples of tablemountains, being the subglacial versions of lava shields, include Fagradalsfjall (Fig. 8).

The Reykjanes Peninsula provides an example of an apparent progression of volcanism related to the retreat of the ice caps (Fig. 11). This comes in the form of most of the tablemountains being created during the deglaciation periods, that is, when the ice was melting. The tablemountains are primarily located at the margins of, or in between, the present Holocene volcanic systems (Figs. 8, 11)..

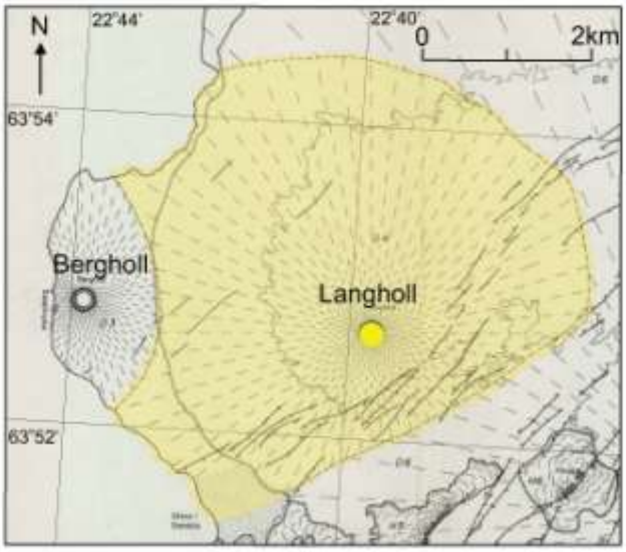

Fig. 9. Olivine-tholeiite lava shield Langholl at the westernmost part of the Reykjanes Peninsula (located in Fig. 8). Also indicated is a part of the olivine-tholeiite lava shield Bergholl. Based on a map by Jonsson (1978).

As the melting of the ice continues and the ice sheet retreats, subaerial lava shields form at the margins of the volcanic systems, whereas inside the volcanic systems themselves the main volcanic forms are Holocene volcanic fissures

Using the maps of Jonsson (1978) as a basis, we compiled a data set for the lava shields on the Reykjanes Peninsula comprising the height, volume and area of the lava shields, and the diameter of their central craters. The data set allows us to assess the relationships between different characteristics of the Holocene shields. The data set also provides an overview of the progression of the various features of the lava shields from the southwest tip of the Reykjanes Peninsula towards the east. This progression inland follows the on-land expression of the axis of the Mid-Atlantic Ridge.

The diameters of craters do not show any correlation with any other parameters. As for the volume and area of shields, there is essentially a linear correlation between these parameters, as is to be expected (Fig. 12). The small deviations from a linear correlation are because neither parameter can be determined with precision. Many of the shields are partly buried by later, partly overlapping shields (Figs. 9, 10), in which case the area cannot be measured with precision. Additionally, the thicknesses of the shield lavas are often difficult to determine so that the estimated volumes are not exact.

The progression of volcanism east along the Reykjanes Peninsula indicates that the large-volume shields tend to be in its eastern part (Fig. 13). However, the trend is certainly not very clear, and the small-volume shields occur with similar frequency in the east as in the west part of the peninsula. What is abundantly clear, however, is that the picrite shields form a class of their own, where the volumes and areas are generally much smaller than those of the olivine-tholeiite shields (Figs. 12 and 13).

2.5 Olivine-tholeiite lava shields

As indicated above, the composition of the lava shield affects their size and form. Olivine-tholeiite shields are larger (Figs. 12, 13) and have a median slope of 2.7^o, height of 60m and base diameter of 3.6km (Rossi, 1996). The size of an olivine-tholeiite shield can extend up to a vast structure such as, for example,

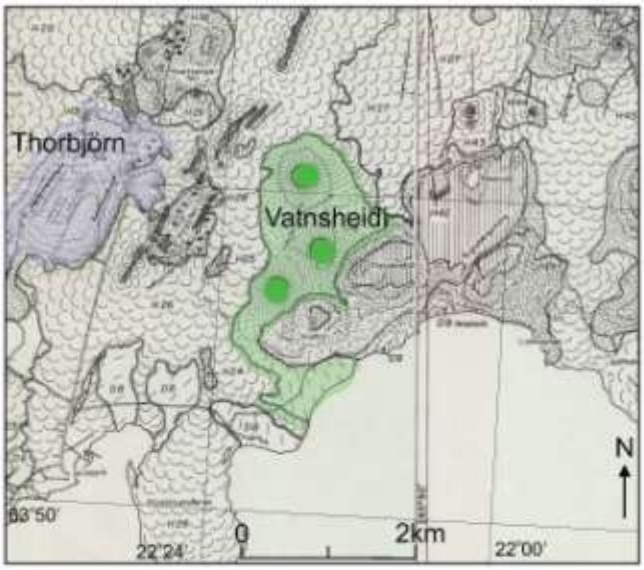

Fig. 10. Picrite lava shield complex Vatnsheidi, composed of three small shields, at the southern part of the Reykjanes Peninsula (located in Fig. 8). Also indicated is the hyaloclastite mountain Thorbjörn, dissected by a graben, and several crater rows (volcanic fissures) between Thorbjörn and Vatnsheidi and north of Vatnsheidi. Based on a map by Jonsson (1978).

Theistareykjabunga in the North Volcanic Zone and Skjaldbreidur in the West Volcanic Zone (Figs. 1, 3, 4). Skjaldbreidur is perhaps the best known "classic" Icelandic lava shield, with a volume of 25km^3, and Theistareykjabunga has a lava apron at least 30 km in diameter (Rossi, 1998).

There are more Holocene olivine tholeiite shields than picrite shields. The olivine-tholeiite shields are formed by magma that, although primitive, is slightly more evolved than the magma forming the picrite shields. A typical olivine-tholeiite shield on the Reykjanes Peninsula is Langholl (Fig. 9). Olivine-tholeiite magma is thought to originate from deep-seated magma reservoirs and to be transported to the surface without any delay time in a shallow crustal magma chamber (Fig. 11).

2.6 Picrite lava shields

The second type of lava shields are composed of picrite magma and are generally significantly smaller than the olivine tholeiite shields (Figs. 1, 10, 12, 13). The size difference of picrite shields and the olivine-tholeiite shields can be very pronounced. A good example of a cluster of small, overlapping picrite shields is the Vatnsheidi complex (Figs. 8, 10). The small sizes of the picrite shields (Figs. 12, 13) is presumably mainly governed by the high density of the magma. The high density of the magma implies small overpressure (driving pressure) for the magma when it reaches the surface (Gudmundsson and Brenner, 2005). Picrite shields have a steeper median slope than the olivine tholeiite shields, or some 5 – 6^o (Rossi, 1996).

Picrite magma is the primary magma from the mantle beneath the Reykjanes Peninsula (Jakobsson et al., 1978). During the waning stages of the last glaciation and the early postglacial period, this magma presumably accumulated under the volcanic zones of the Reykjanes Peninsula at exceptionally high rates (Gudmundsson, 1986). This high rate of accumulation is partly attributable to the decrease in lithostatic pressure as a result of the rapid melting of the glaciers, and partly to reservoir tensile stresses and increased porosity (section 3). Later in the postglacial period, the picrite magma was presumably trapped by the lighter olivine tholeiite magma that subsequently accumulated in the upper parts of the magma reservoir (Fig. 11; Gudmundsson, 1986). Many of the picrite shields are thought to have been buried by subsequent olivine tholeiite shields, and this accounts partly for their comparatively small number (Figs. 12, 13).

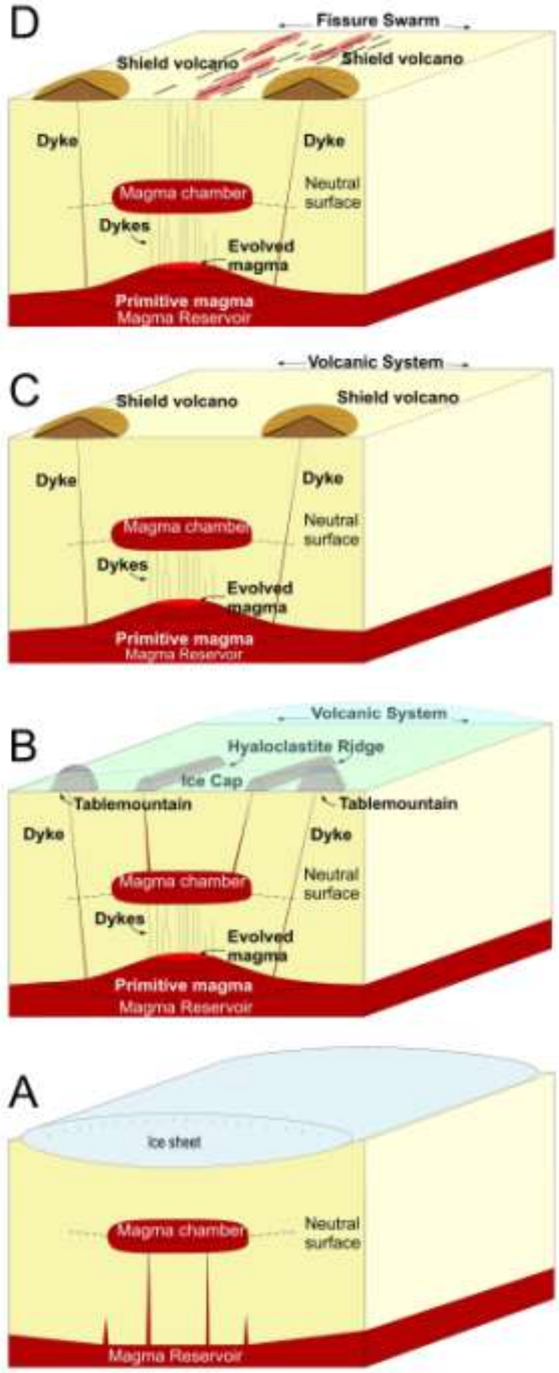

Fig. 11. Schematic illustration of the development of volcanism in the North and West Volcanic Zones from late Weischelian to early Holocene. (A) At the peak of the Weischelian the glacial load generates compressive stresses deep into the mantle so that porosity and reservoir size decreases. The downbending may, however, during parts of the Weischelian have opened up pathways for magma from the mantle, most of which would have become arrested at the neutral surface (the exact location of which depends on the mechanical layering of the crust, cf. Figs. 15, 16) and encouraged formation of sill-like shallow, crustal magma chambers. (B) Close to the end of the Weischelian, when much of the ice cap had melted, the size and melt content of the magma reservoir had increased and the stress situation made it possible for some eruptions to occur, particularly in the marginal parts, or in between, the volcanic systems. These eruptions gave rise to tablemountains (Fig. 8) and, when the ice had melted, lava shields. As the ice-induced bending stresses gradually relaxed, some fissure eruptions occurred, producing hyaloclastite ridges and cones (Fig. 8). In some parts of the volcanic zones, shallow magma chambers may have supplied magma to some fissure eruptions at this stage. (C) As the ice melted away completely, the first postglacial volcanoes to form were the lava shields, particularly at the margins of the volcanic systems (for the sake of clarity, the tablemountains and hyaloclastite ridges in B are omitted here). At this stage the reservoir was still large and deep and supplied hot, primitive magma (picrite and olivine tholeiite) to the shield eruptions. (D) When the remaining ice-induced stresses became completely relaxed because of plate pull (divergent plate movements), the lava shield activity died out and fissure eruptions inside the volcanic system became dominant. Some of the fissures were supplied with magma from the evolved tops of the deep-seated reservoirs, others from the shallow crustal magma chambers. As a result, the magma in the fissure eruptions tends to be considerably more evolved than in the lava shields.

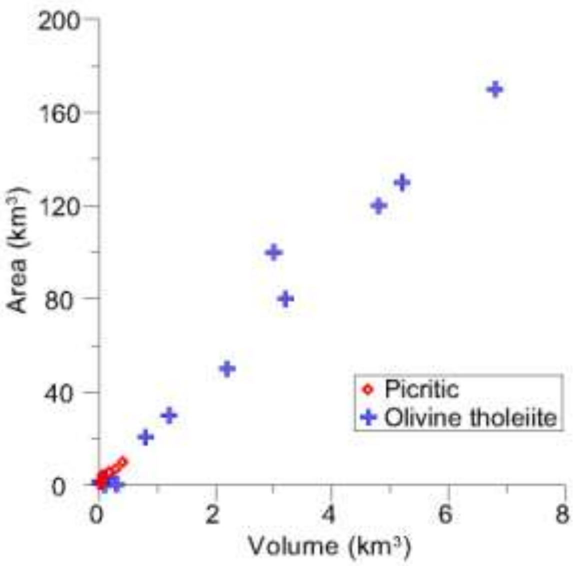

Fig. 12: Graph showing the correlation between volume and area of lava shields on the Reykjanes Peninsula. The graph shows a near-linear correlation, as expected, but also that the picrite lava shield volumes are an order-of-a-magnitude smaller than those of the olivine-tholeiite shields. Data from Jonsson (1978) and Jakobsson et al. (1978).

3. Effects of deglaciation on reservoir porosity

One of the remarkable features of the late-glacial stage (that is, the end part of the Weischelian or Devensian) and early Holocene volcanism in Iceland is that the erupted volumes per unit time are so large. There is good evidence that the volume production of most volcanic systems in Iceland peaked at the end of the Weicheselian and in the early Holocene and has since decreased markedly (Annertz et al., 1985; Gudmundsson, 1986; Vilmundardottir and Larsen, 1986; Jull and McKenzie, 1996; Bourgeois et al., 1998). It is not just that eruptions were more frequent in the volcanic systems in the late glacial and early Holocene, but the volumes of the individual eruptions were unusually large at that time compared with eruptions of the later part of the Holocene. For example, on the Reykjanes Peninsula the average volume of 26 lava shields is 1.11 km^3 whereas that of lavas from 101 volcanic fissure is only 0.11 km^3 (Gudmundsson, 1986). Since most of the shields formed earlier in Holocene than most of the fissures, it is clear that the average volume in individual eruptions has declined throughout the Holocene.

To explain this difference in average eruptive volume we have to consider two questions. First, do the magmas in the shield and fissure eruptions come from the same magma reservoirs? If they do, then the second question is what factors could have changed in the reservoir so as to change the average eruptive volume? While it is known that, in general, many fissure eruptions in the volcanic systems of Iceland derive their magmas from shallow crustal magma chambers (Fig. 11; Gudmundsson, 2006), many, perhaps most, of the volcanic systems on the Reykjanes Peninsula seem to lack shallow

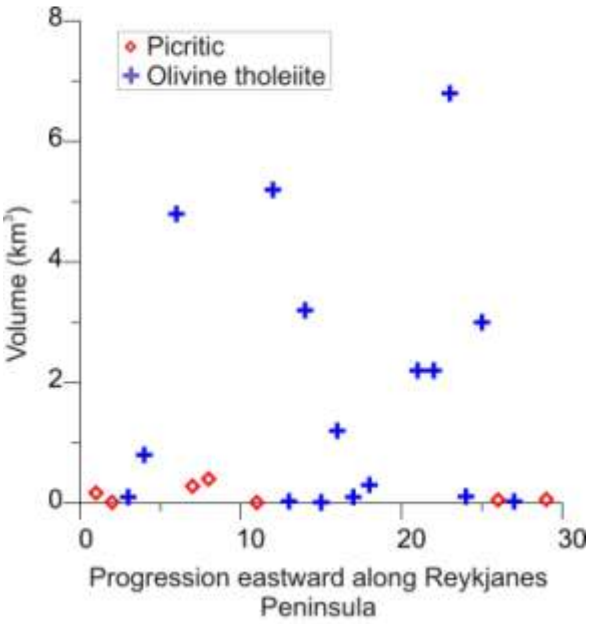

Fig. 13. Change in lava-shield volume eastward along the Reykjanes Peninsula. The results show very clearly the difference in volume between the picrite and the olivine-tholeiite shields, but also that the picrite-shield volumes are essentially constant. By contrast, the olivine-tholeiite shield volumes have a wide range but the maximum volumes seem to be somewhat larger in the eastern than in the western part of the peninsula. Data from Jonsson (1978) and Jakobsson et al. (1978).

magma chambers (Gudmundsson, 1986). And, generally, most of the fissure eruptions derived from crustal chambers would tend to be confined to the central volcano itself, and its surrounding, whereas the fissure eruptions in the parts of the volcanic system far away from the central volcano would be supplied with magma directly from the deeper reservoir. For the volcanic systems in general, and particularly for those on the Reykjanes Peninsula, there is thus little doubt that the volume of individual eruptions reached its peak at end of the Weischelian and in the early Holocene and has since declined.

The most likely explanation for the eruptive volume peak during this period is that the melt content in the reservoirs was at that time greater than during the subsequent part of Holocene. If a deep-seated reservoir associated with one of the volcanic systems on the Reykjanes Peninsula (Figs. 8, 11) has a total volume V_r, with melt or magma fraction (porosity) n, pore compressibility β_p, and magma compressibility β_m, then the total volume of magma V_r^e that flows out of the reservoir in a single magma flow is (Gudmundsson, 1987):

$$V_r^e = np_e(\beta_m + \beta_p)V_r \qquad (1)$$

where p_e is the magma excess pressure in the reservoir before it ruptures and magma begins to flow out of it. As used here, the magmatic excess pressure in the reservoir is the difference between the total magma pressure P_t in the reservoir at its time of rupture and the lithostatic stress in the host rock, that is, $p_e = P_t - p_l$ where p_l is the lithostatic stress or overburden pressure at the reservoir rupture site. The condition for reservoir rupture and dyke (or inclined sheet) initiation is (Gudmundsson, 2006):

$$p_l + p_e = \sigma_3 + T_0 \qquad (2)$$

where σ_3 is the minimum compressive or maximum tensile principal stress, and T_0 the local in situ tensile strength, at the rupture site. Following the tradition in geology, compressive stress is considered positive. Thus, for an absolute tension to occur, σ_3 must be negative. By contrast, the maximum compressive principal stress σ_1 is always positive.

In Eq. (2), the excess pressure at rupture, p_e, is approximately equal to the in situ tensile strength T_0 which normally ranges between 0.5-6 MPa (Schultz, 1995; Amadei and Stephansson, 1997) and is thus essentially a constant factor that is unlikely to explain the large difference in eruptive volumes from the reservoirs. In Eq. (2), the factor V_r^e is equal to the volume of a dyke (or an inclined sheet) plus, in case it is a feeder, the associated eruptive volume. For a given host rock and magma composition the pore

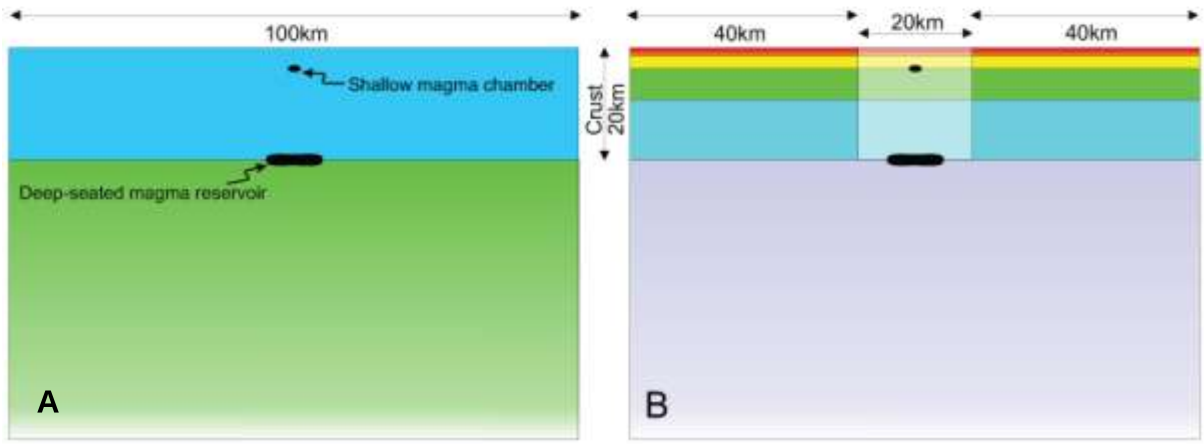

Fig. 14. General setup of the numerical models in Figs. 15 and 16 using the finite-element program Ansys. A 100-km-wide zone, that is, a typical volcanic zone and its close surroundings (Fig. 1) is subject to ice load along a part of (Fig. 16) or its entire (Fig. 15) width. (A) A shallow crustal magma chamber is located at 3 km depth and is 2 km wide whereas the deep-seated reservoir is at 20 km depth and is 10 km wide. (B) The crust is layered, the thicknesses and the mechanical properties being as indicated in Table 1. Also indicated in Table 1 is the ice load, which varies from a maximum in the ice-sheet centre to a minimum at the ice margins, based on rough estimates of the ice thickness (cf. Paterson, 1981). In Fig. 15 the entire 100-km-wide zone is subject to glacial load, but only the central 20-km-wide zone in Fig. 16.

compressibility β_p and the magma compressibility β_m are essentially constants. Since p_e is also close to being a constant, the only parameters in Eq. (1) that could differ between the peak period and with later parts of the Holocene are the melt or magma fraction (porosity) n and the total reservoir volume V_r. We made numerical models to check if these could have been affected by the glacial load, and concluded that, for certain boundary conditions, both would have been likely to reach their maximum during the late Weichselian.

We made several numerical models, using the Ansys finite-element program (www.ansys.com; Logan, 2002) to explore the potential effects of the glacial load on the porosity (and size) of the deep-seated magma reservoirs. The general structure of the models is as follows. The crust is 20 km thick and hosts two magma chambers; an upper shallow magma chamber and a deep-seated reservoir (Fig. 14A). In this two-dimensional model both chambers are elliptical (sill-like) in shape, and the pair is referred to as a double magma chamber (Gudmundsson, 2006). The upper chamber is at 3 km depth and is 2 km wide whereas the lower chamber is at 20 km depth and is 10 km wide. These figures are based on geological and geophysical data (Gudmundsson, 1987; 2000; 2006). In all the models, a tensile stress of 5 MPa is applied to simulate spreading-related plate pull within the volcanic zones. Since we are mainly concerned with the late Weischelian effects on the volcanic zones themselves, we focus on the ice load within the volcanic zones. These are generally mountainous areas and would retain glaciers longer than the lower-elevation surroundings; in fact, some of these areas contain glaciers today (Fig. 1).

	Depth	Volcanic System	
		Inside	**Outside**
Layer 1	0.5 km	2 GPa	4 GPa
Layer 2	1 km	0.5 GPa	1 GPa
Layer 3	2.2 km	14.5 GPa	29 GPa
Layer 4	5.6 km	25.5 GPa	51 GPa
Layer 5	10.7 km	34.5 GPa	67 GPa
Layer 6	150 km	100 GPa	100 GPa
Ice Load			
		Centre	**Edge**
Full Ice		18 MPa	1.5 MPa
20km Ice		4 MPa	1.5 MPa

Table. 1. Young's moduli inside and outside the volcanic systems (in giga-pascals), thicknesses (depths) of the crustal layers, and the ice-induced loading (in mega-pascals) for the numerical models (Figs. 14-16).

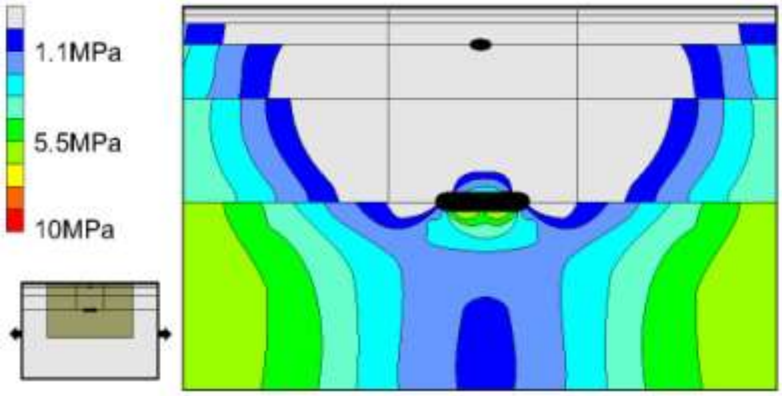

Fig. 15. Finite-element model showing the contours of the maximum principal tensile (minimum compressive) stress, σ_3, in mega-pascals generated as a result of loading by a 100-km-wide ice sheet (and a plate-pull tension of 5 MPa). The entire model is shown on the inset. Clearly, the ice-induced stresses in the crust are mainly compressive, so as to make dyke injections from either the shallow chamber or the deep-seated reservoir rather unlikely. The compressive stress regime extends partly into the top part of the upper mantle, meaning that at this stage the effective size and porosity of the deep-seated reservoir would tend to be small. We also made a model with a 400-km-wide ice sheet, and the results show that the ice-induced compressive regime extends to considerable depths in the upper mantle surrounding the deep-seated reservoir. Thus, during the greater part of each glacial period the conditions for magma accumulation and eruptions are unfavourable.

Thus, in the numerical models we explore the effects of an ice load generated, first, by a 100-km-broad ice sheet covering an entire volcanic zone and its surroundings, and, second, a 20-km-broad ice covering one volcanic system and its surroundings (Fig. 14B). These ice sheets are of similar dimensions and locations to some of those that occur today in Iceland (Fig. 1). Since it is well known from observations and theoretical studies that an ice-sheet thickness varies from its centre to its margins (Paterson, 1981), we take the ice load as varying from a maximum at the ice-sheet centre to a minimum at its margins. Also, the crust is modelled as layered, with the layers having different thicknesses and mechanical properties, in particular different Young's moduli (stiffnesses). A Holocene volcanic system contains the youngest rocks, which normally have a high porosity and a great number of tectonic fractures. Generally, the greater the number of fractures and the higher the porosity, the lower will be the effective Young's modulus of the rock. Consequently, we assume that the layers within a Holocene volcanic system have lower Young's moduli than the adjacent layers outside the system, as is, indeed, supported by geophysical measurements (Gudmundsson, 1988). The data on the ice load and the crustal layering are summarised in Table 1.

The results for the 100-km-wide ice sheet show that ice-induced compressive stresses dominate throughout the entire crust of the volcanic system and penetrate also partly into the upper mantle surrounding the lower reservoir (Fig. 15). Clearly, these stress conditions are unfavourable for volcanic eruptions. But they are also unfavourable for porosity development and magma accumulation in the deep-seated reservoir. This follows because the ice-induced compressive stress penetrates down to the reservoir and partly into the surrounding mantle. Also, those tensile stresses that develop in the surrounding mantle are mostly very low and thus unlikely to open up, or generate, new fractures, that is, new fracture porosity.

By contrast, the 20-km-wide ice sheet induces tensile stresses in a large region surrounding the deep-seated reservoir (Fig. 16). That there is also some induced tensile stress around the shallow chamber is clear, but for the

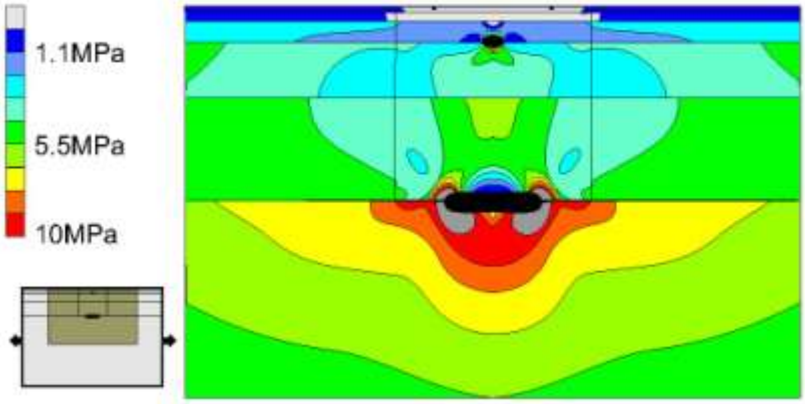

Fig. 16. Finite-element model showing the contours of the maximum principal tensile (minimum compressive) stress, σ_3, in mega-pascals generated as a results of loading by a 20-km-wide ice sheet (and a plate-pull tension of 5 MPa). The entire model is shown on the inset. Clearly, the ice-induced stresses in the crust are mainly tensile, both in the lower crust as well as in the uppermost part of the mantle. These stress conditions encourage increased porosity and magma accumulation in the deep-seated reservoir, so as to make its effective size larger. Thus, eruptions from the reservoir, at this stage or somewhat later, would tap of large volumes of comparatively hot and primitive magma, thereby explaining the composition and size of typical lava shields.

present purpose it is the tensile stress around the lower reservoir that is of importance. We propose that the induced tensile stress around the deep-seated reservoir increased the fracture-related porosity in a large region at the contact between the crust and upper mantle beneath the volcanic system during the late-glacial (late Weischelian) period, and that this porosity increase also increased the effective size of the reservoir.

Thus, the numerical model in Fig. 16 indicates that during the late-glacial period ice-induced tensile stresses increased the porosity (melt fracture) and effective size of the deep-seated reservoirs underlying the volcanic systems in Iceland in general, and those of the Reykjanes Peninsula in particular. That is, during this period both the melt fraction n and the total reservoir volume V_r were unusually large. It then follows from Eq. (1) that the total volume of magma V_r^e that could flow out of the reservoir in a single magma flow, including the eruptive volume, was unusually large.

The large V_r^e, we propose, explains the large-volume eruptions at the end of the Weischelian and during the early Holocene, since the factors n and V_r started to decrease only when the ice had melted away completely. Thus, comparatively large melt fractions and reservoirs were maintained into the early Holocene, but then gradually decreased as melt or magma was tapped off the reservoirs and the fracture-related porosity decreased. This model also explains the common occurrence of very primitive magmas in this same period (late-glacial and early Holocene) since, from Fig. 16, it is clear that the reservoir extends to much greater depths in the mantle, and thus to melts at higher temperature and with a more primitive composition, than during the later part of the Holocene.

4. Effects of deglaciation on volcanism

At the maximum glaciation the frequency of volcanic eruptions was presumably much lower than that during the interglacial periods. For example, during the peak glaciation of the Weischelian the entire Iceland was covered with an ice sheet that may have been as thick as 2-3 km. At that state the compressive regime would have extended deep into the upper mantle and thus prevented much accumulation of magma in the deeper reservoirs. Even when the glaciers had started to melt, but were still covering the highland parts of the volcanic zones, the compressive regime extended down to the deep-seated reservoirs and partly into the surrounding upper mantle (Fig. 15).

This conclusion is supported by the observation that most of the tablemountains and hyaloclastite ridges in Iceland appear to have formed during the waning stages rather than the peaks of the glacial periods; in particular, the most conspicuous hyaloclastite mountains seen in the volcanic zones of Iceland today (Fig. 1) were presumably formed towards the end of the Weischelian (Gudmundsson, 1986; 2000; Bourgeois et al., 1998). During the greater part of each glacial period, any magmatism is likely to have generated intrusions and encouraged crustal magma-chamber formation, rather than volcanic eruptions, as is indicated in the conceptual model illustrated in Fig. 11A.

In terms of this model (Fig. 11A), any magma that was able to rise through a dyke into the crust during the major part of a glacial period became arrested at the neutral surface, where the change of tectonic stress from tensile (mostly relative, not absolute, tension) to compressive. The exact location of the neutral surface in the crust and upper mantle depends on mechanical layering and properties and thicknesses of those layers, but any surface or discontinuity where there is horizontal compression above a region of horizontal tension would tend to arrest dykes and change them into sills.

That dykes change into sills is commonly observed at the location of such stress changes (Gudmundsson, 2006; Gudmundsson and Philipp, 2006). If the sills thus formed received fresh magma through repeated dyke injections frequently enough, they had a change of developing into shallow magma chambers (Gudmundsson, 1990). Magma-chamber development from sills is in fact well established, such as at mid-ocean ridges (Mutter et al., 1995), as well as in Iceland (Gudmundsson, 2006). Depending on the exact layering and the size and thickness of the ice sheet at any particular time during an ice period, ice-induced tensile stresses in the lower brittle crust and the compressive stresses in the upper brittle crust may have encouraged the development of sill-like shallow chambers.

The gradual relaxation of the ice-induced compressive stresses in the upper part of the crust on approaching the end of the Weischelian would have increased the likelihood of dyke-fed eruptions. In addition, as the ice load decreased, and at the same time there was extra fracture-related porosity in the reservoirs (Fig. 16), magma accumulation in the reservoirs increased. When the ice melted completely, the reservoirs gradually acquired their present “dome-shaped” form (Fig. 11) where density stratification of the magma could take place.

The first volcanoes to form during a glacial retreat are likely to have been tablemountains (Fig. 11B). This follows because the upper crust may then still have been subject to some horizontal compression (Fig. 16), particularly in areas far away from shallow chambers. Thus, volcanic fissures that formed in the initial stage of an eruption, when the overpressure of the magma is at its peak, were unlikely to remain open as such for long but rather to form some sort of pipe-like conduits at one or more localities along the original fissure. At this stage the magma is likely to come from the outer, more primitive parts of the deep-seated magma reservoir rather than from its central part or a shallow magma chamber (Fig. 11B). Thus, in case an eruption is possible, it would tend to supply magma to the surface at the margins of, or in between, the volcanic systems. Also, except for the very primitive, high-density picrite magmas, the erupted volumes tend to be large, since the sources are deep-seated reservoirs that, at this stage, were larger and with greater melt content than during the later part of Holocene.

When doming of the crustal segment, because of magma accumulation in the reservoir at the bottom of the crust, starts to have an effect on the stresses in the upper part of the crust, the tensile stresses tend to concentrate not at the centre but rather at and above the lateral ends (margins) of the shallow magma chamber. At this stage the shallow chamber has not received additional input of new magma so that it is in a lithostatic equilibrium with the surrounding rock and the only loading is due to the doming pressure from the deeper reservoir. Numerical models for such loading conditions show quite clearly that the main tensile (and, in fact, shear) stresses occur above the lateral ends of the shallow chamber (Gudmundsson and Nilsen, 2006). Thus, if any dykes are injected from the shallow chamber at this stage, they would tend to propagate upwards from its lateral ends and reach the surface at the margin of the associated volcanic system as hyaloclastite ridges (Fig. 11B). Since the dykes are derived from a shallow chamber that has not been replenished with primitive magma for a long time, the issued magma would normally be somewhat more evolved than that which forms the tablemountains.

When the glacier has completely melted away, the lava shields take over from the tablemountains (Fig. 11C). These tend to be of similar volumes and compositions as the tablemountains, the primary difference being that the shields form in subaerial rather than subglacial eruptions. Then high-density picrite magma is for a while able to rise from the deep magma reservoir to the surface forming lava shields (Fig. 11C). The initial stress change from the ice retreat creates a pressure gradient allowing the picrite magma to reach the surface. The olivine tholeiite magma is more evolved than the picrite and gradually becomes the dominant magma to reach the surface at the margins of the volcanic systems.

Subsequently, when the plate-pull tensile stress has relaxed the doming-generated compressive stress in the lower part of the crust above the deep reservoir fissure eruptions, tapping both the (then smaller, and with less melt content) deeper reservoir as well as the shallow crustal magma chamber, start to form near the centre of the volcanic system and the associated central volcano.

5. Implications for volcanic hazards

Hazards can be defined as extreme geological events capable of causing disaster. Such events and their effects are fundamentally determined by their location, magnitude, timing and frequency (Alexander, 2000). More specifically, a volcanic hazard is the probability of a potentially dangerous volcanic process to occur; it refers to the conditions whereby dangerous volcanic processes might occur. Volcanic risk, however, is the probability of loss of human lives, property and productive capacity in a region subject to volcanic hazard. Thus, hazards have to do with the probability of occurrence of a certain type of event, in this case a volcanic eruption, whereas risk has to do with the probable effects of that type of event on human society and its infrastructure.

The lava shields in Iceland are essentially confined to the NVZ and, in particular, to the WVZ (Fig. 1). In case a shield eruption were to occur again in these areas, the main possible hazardous location would be in the vicinity of the capital, Reykjavik (Fig. 8). Since shields have commonly formed at the margins of or outside the volcanic systems themselves (Figs. 1, 8), then if the same conditions would apply, we might expect a shield eruption on the Reykjanes Peninsula to occur, say, at the northern margins of, or in between, the volcanic systems of Krisuvik and Brennisteinsfjöll, in which case the Reykjavik Metropolitan Area would be at risk. This area has a population of about 190,000, compared with the total population of Iceland of just over 300,000. Reykjavik not only represents the main area of population within Iceland but is also a centre of economic activity, infrastructure, business and commerce. Further to this the location of Keflavik airport on the Reykjanes Peninsula, the main international airport in Iceland, and the town of Grindavik also pose areas of risk (Fig. 8).

The main hazard associated with lava-shield formation are the lava-flow fields themselves. Given the proportion of olivine tholeiite shields to picrite shields in the Holocene, the former are more likely to be the form if a shield eruption were to occur in the near future. From this it can be deduced that a resulting lava-flow field would be mainly composed of pahoehoe flow units. The field of a comparatively large olivine-tholeiite shield is largely formed through an extensive network of lava tubes, and also from rootless outpourings on the flanks of the shield (Rossi, 1996).

As regards the risks involved, lava flows in their nature are rarely life threatening, due to their characteristically low velocities and, for pahoehoe flows, comparatively low effusion rates. Rossi (1996) deduced that the dominant pahoehoe morphology of the Icelandic shields indicated a low effusion rate at the vent, probably of 5-15m^3/s or less. Their damage to property is through burning, crushing and burial of structures, and their volume can be sufficient to inundate tens of kilometres of usable land (McGuire, 1998).

Lava flows are naturally governed by topography and as a result of this their courses can be predicted. Also, previous eruptions, both in Iceland and in other countries, have shown that it is possible to guide and slow down lava flows. In the 1991 – 1993 eruption of Etna, Italian scientists succeeded in holding back the flow front of the lava for several months with an artificially constructed earth barrier. They also blocked the main feeder tube of the lava near to its source, cutting off the supply to the flow front. This caused thickening rather than lengthening as the magma subsequently travelled over the surface of the flow field (Barberi et al., 1993).

Similarly, in Iceland during the 1973 eruption of Eldfell in the Vestmannaeyjar the lava flow threatened to block the natural harbour entrance and thus threatened one of the largest fishing ports of the country. The front of the lava was cooled by means of spraying water onto it, and was thus slowed as the flow backed up onto the cooled front (Williams, 1997). The Vestmannaeyjar Volcanic System (Fig. 1) is also the home to the only historical shield eruption in Iceland, that on top of the island of Surtsey that formed in 1963 – 1967.

Due to the imprecise nature of the hazard involved in a possible future shield eruption, the creation of a specific hazard map for the Reykjavik area and the Reykjanes Peninsula is not possible. It would be possible, however, in the event of a shield eruption to predict the course of the lava flow. The prevention of too much damage to some areas could also be achieved through this.

An associated hazard of lava flows is the emission of gas, which can accumulate in low-lying areas, and in basements and cellars. This was the case with the 1973 Heimaey eruption, as concentrations of gases rose continually throughout the early months of the eruption (Williams and Moore, 1976). Basalts are known to have the largest potential for dissolved sulphur and emission of SO_2 and H_2S during an eruption (Self et al., 1998). There are documented cases from the 1783 Laki Fissure eruption of respiratory problems, and damage to crops and livestock from the gases (Grattan, 1998).

A shield eruption, or a volcanic eruption of any kind, in the vicinity of the Reykjavik area would also pollute the groundwater. In fact, the groundwater reservoirs of Reykjavik are mainly located in the Holocene lava flows of the northern part of the Krisuvik Volcanic System (Fig. 8), so that any eruption in that area would be likely to pollute the aquifers and, perhaps, make them unusable for the foreseeable future.

Our models, however, indicate that the probability of lava-shield eruptions is not very high in Iceland in general, and on the Reykjanes Peninsula in particular, during the present tectonic conditions in the WVZ and the NVZ, where nearly all the Holocene shields are located (Fig. 1). The models, however, do not really explain the lava shield that formed on top of the island of Surtsey in the 1963-67 eruption of the southernmost part of the Vestmannaeyjar Volcanic System in the EVZ (Fig. 1). Thus even if the probability of a lava-shield eruption may be small, as our models indicate, they may still form under some, as yet unknown and unexplored, tectonic conditions.

6. Discussion and conclusions

One of the surprising observations about the volcanism in Iceland is the frequency and size of lava shield eruptions in early Holocene, particularly in the West and North Volcanic Zones (Fig. 1). Just preceding these large lava shield eruptions was the formation of numerous tablemountains at roughly the same localities. The tablemountains that are most conspicuous in the volcanic zones today were presumably mostly formed during the waning stages of the last glaciation, that is, close to the end of the Weischelian. Thus, the late glacial was characterised by the formation of comparatively large, subglacial tablemountains followed, when the ice had completely melted away, by equally large, or larger, volcanic shields (Fig. 11).

Both the lava shields and the tablemountains are primarily composed of primitive basaltic magma, picrite and olivine tholeiite. The primitive composition, as well the large volumes and the location of these volcanoes mainly at the margins of, or in between, the volcanic systems (Figs. 8 and 11), indicates that they were supplied with magma from deep-seated reservoirs rather than from shallow crustal magma chambers (Fig. 11). Where data are available they indicate that the volumes of the lava shields are, on average, an order of a magnitude larger than those of lavas from volcanic fissures in the same areas of the volcanic zones (Gudmundsson, 1986).

In this paper we have addressed several basic questions as to the Holocene volcanic shields (and very similar questions apply to the somewhat earlier formed, subglacial tablemountains), namely:

- Why are the shields mostly located at the margins of, or in between, the volcanic systems?
- Why are the shields, on average, of much greater volumes than the volcanic fissures in the same parts of the volcanic zones?
- Why is the shield magma much more primitive than that of the nearby volcanic fissures?
- Why did essentially all the shields form in the early part of the Holocene?
- What is the probability of a new shield eruption in the areas where they are most common, and what are the associated hazards and risks?

Our answers to these questions, and the main conclusions of the paper, may be summarised as follows. (1) The lava shields tend to form at the margins of, or in between volcanic systems, because the stress field in the early Holocene did not allow eruptions in the central parts of the volcanic systems while allowing eruptions in the marginal parts (Fig. 11; Gudmundsson, 1986). This stress field, partly related to the bending of the crust above source reservoirs, lasted only for a while, and was later relaxed as a consequence of divergent plate separation and associated plate pull.

(2) The large volumes of the lava shields (and tablemountains) in relation to the volumes of the lavas generated by volcanic fissures in the same areas of the volcanic zones (Gudmundsson, 1986) can be explained as follows. During late-glacial (near the end of the Weischelian) the glacial load increased the fracture porosity and, thereby, the sizes of the deep-seated reservoirs, extending them laterally as well as vertically into deeper parts of the upper mantle (Fig. 16). Consequently, when the lava shields (and, somewhat earlier, the tablemountains) were erupted, much more melt or magma was available to feed a single eruption than later in Holocene. And, because of the greater vertical extent of the reservoir at this time, this magma tended to be hotter and more primitive than when the reservoir had diminished in size during the later parts of the Holocene. Thus, the numerical model in Fig. 16 together with Eqs. 1 and 2 largely explain the large volumes and primitive compositions of the lava shields.

(3) The primitive composition of the lava shields themselves is partly accounted for in point (2). However, it should also be stressed that the lavas issued by later-formed volcanic fissures tend to be more evolved for two main reasons. First, those of the volcanic fissures that come from the deep-seated reservoirs, such as most of the fissures on the Reykjanes Peninsula (Fig. 8), derive their magmas from the central, that is, the top parts of the reservoirs where the most evolved magma tends to accumulate (Fig. 11). This follows because all the fissures occur within the volcanic systems, and commonly close to their central parts. Second, many volcanic fissures, particularly those providing comparatively small eruptions within central

volcanoes, derive their magmas from shallow, crustal magma chambers, and these, as a rule, contain more evolved magmas than the deep-seated source reservoirs (Gudmundsson, 2000).

(4) The lava shields formed almost exclusively in the early part of the Holocene because at that time the state of stress in the crust was favourable for their formation (Fig. 11). Thus, during this early part of Holocene, dykes could propagate to the surface at the margins of, or in between, the volcanic systems, whereas later the local stresses favoured dyke emplacement inside, and close to the centres, of the volcanic systems. Since the dykes injected from the marginal parts of the reservoirs tend to tap off deeper magma sources, and because in the early Holocene the reservoirs were larger and with more primitive magma, the picrite and olivine-tholeiite shields are mostly confined to this period.

(5) If the models presented in this paper reflect reasonably well the volcanotectonic conditions that must be satisfied for lava shields to form, then their probability of formation in the WVZ and the NVZ is currently low. Consequently, the hazards and risks associated with lava-shield eruption in these zones are low. What we have not addressed in this paper is the condition for lava-shield formation in off-rift zones, such as in the southernmost part of the EVZ (Fig. 1). These conditions may be, and are likely to be, different from those discussed in connection with deglaciation in this paper if only for the reason that the Surtsey lava shield formed only a few decades ago, long after all tectonic effects of the Weischelian had disappeared. These conditions should be explored, but are best left for analysis in a future work.

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12 Appendix II

Mechanical interaction between active volcanoes in Iceland

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Abstract

We test the possibility of mechanical interaction between eight central volcanoes in the central part of the active Iceland rift zone. The average distance between the volcanoes is 30 km; all are thought to have shallow magma chambers, and many contain collapse calderas. We analyzed many finite-element models with the volcanoes subject to a tensile stress of 5 MPa (equal to the maximum in situ tensile strength of the crust) in a direction parallel to the spreading vector, N105°E. The results show zones between many nearby volcanoes where the tensile stresses exceed the in situ tensile strength of the crust. The results indicate that mechanical interaction between volcanoes in a pair, such as simultaneous dike emplacement, seismogenic faulting, and deformation, may be common in this part of Iceland, in agreement with observations.

Keywords: central volcano, clusters, mechanical interaction, inclusion, hole, dyke sharing

1. Introduction

Many active central volcanoes (composite volcanoes, stratovolcanoes, calderas) occur in clusters referred to as fields. When assessing volcanic hazards and risks, it is important to know in what way, if any, volcanoes within a given field interact mechanically. For example, do deformation and earthquakes in one volcano trigger deformation and earthquakes in a nearby volcano? Alternatively, are dikes occasionally "shared" between volcanoes? That is, do simultaneously formed dike segments (propagating laterally, vertically or both) eventually link up two volcanoes so as to allow magma transport between them? Some data support interaction between closely spaced volcanoes, such as the calderas of the Galapagos Islands (Amelung et al., 2000). The mechanism by which such an interaction occurs is, however, not well understood. Studies on the correlation between closely to moderately spaced volcanoes in other countries (e.g. Bebbington and Lei, 1996) has revealed no substantial link. However, a characteristic of the situation in Iceland, is the divergent plate boundary and thus the dominant tensile stress.

Ideas about possible mechanical interaction between active volcanoes in Iceland, and between volcanoes and seismic zones, have been around for some time. For example, some authors have proposed lateral dike propagation from one central volcano to another (Blake, 1984; McGarvie, 1984). Others have suggested stress transfer between volcanic and seismic zones (Gudmundsson, 2000; Gudmundsson and Brenner, 2003). There is, indeed, some evidence that volcanic activity in the East Volcanic Zone is in harmony with earthquake activity in the South Iceland Seismic Zone (Larsen et al., 1998).

The principal aim of this paper is to test the likelihood of mechanical interaction between active central volcanoes through numerical models. An ideal place to test such an interaction is where there are many active volcanoes with well-documented histories within a comparatively small volcanic field. We selected that part of Iceland which has the greatest number of active central volcanoes per unit area (Figures 1, 2). This is also the part above the

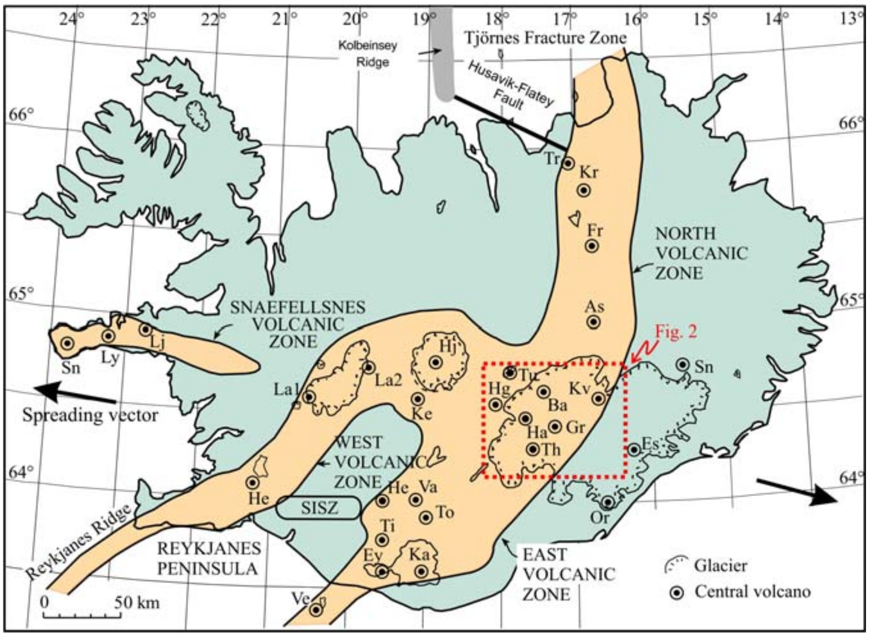

Figure 1. Volcanotectonic map of Iceland showing the tectonic framework of the 8 central volcanoes used in this study. The central volcanoes indicated are as follows: Tr= Theystareykir, Kr = Krafla, Fr = Fremri-Namur, As = Askja, Kv = Kverkfjöll, Th = Thordarhyrna, Gr = Grimsvötn, Ha = Hamarinn, Ba = Bardarbunga, Tu = Tungnafellsjökull (including Vonarskard), Hg = Hagöngur, Ka = Katla, Ey = Eyjafjallajökull, Ve = Vestmannaeyjar, Ti = Tindfjallajökull, Va = Vatnafjöll, To = Torfajökull, He = Hekla, Hj = Hofsjökull, Ke = Kerlingafjöll, La1&2 = Langjokull, He = Hengill, Sn = Snaefellsjökull, Ly = Lysuskard, and Lj = Ljosufjöll, Or = Öraefajökull, Es = Esjufjöll, and Sn = Snaefell. SISZ = South Iceland Seismic Zone. Modified from Gudmundsson (2000).

centre of the Iceland Mantle Plume (Wolfe et al., 1997; Gudmundsson, 2000) and where there have been many eruptions in recent decades (Larsen et al., 1998; Gudmundsson and Högnadottir, 2007; Thordarson and Larsen, 2007).

2. Volcanotectonic framework

Volcanism in Iceland is mostly confined to one major zone (Figure 1). This zone is traditionally referred to as the Neovolcanic Zone and is covered by rocks belonging to the Brunhes magnetic epoch (< 0.8 Ma). The Neovolcanic Zone has three main segments or subzones, namely the North Volcanic Zone (NVZ), the East Volcanic Zone (EVZ), and the West Volcanic Zone (WVZ); some authors divide the last one further into the WVZ and the Reykjanes Peninsula (RP). These subzones form the divergent plate boundary (rift zone), except for the portion of the EVZ that lies southwest of the volcano Torfajökull (Figure 1) which is a propagating rift. There are, in addition, several off-rift central volcanoes in Iceland. These include the volcanoes of the Snaefellsnes Volcanic Zone, an old Tertiary zone that became reactivated at 2 Ma, and the central volcanoes Öraefajökull, Esjufjöll, and Snaefell, close to or inside the eastern part of the Vatnajökull Ice Cap.

Since all the central volcanoes considered in this study are within the divergent plate boundary (Figures 1, 2), they are all subject to tensile stress pulling the plates apart. The tensile stress may be absolute (the minimum principal compressive stress σ_3 being negative) close to the surface, but at depths of a kilometer or more the tensile stress is relative, that is, a reduction in the compressive stress (Gudmundsson, 2006). In the Iceland area the average spreading rate is about 1.8 cm/yr in the direction N105°E, that is, parallel with the spreading vector (DeMetz et al., 1990, 1994).

A rift-zone central volcano is normally supplied with magma from a shallow crustal magma chamber which, in turn, has its source at the bottom of the crust. Such a pair is referred to

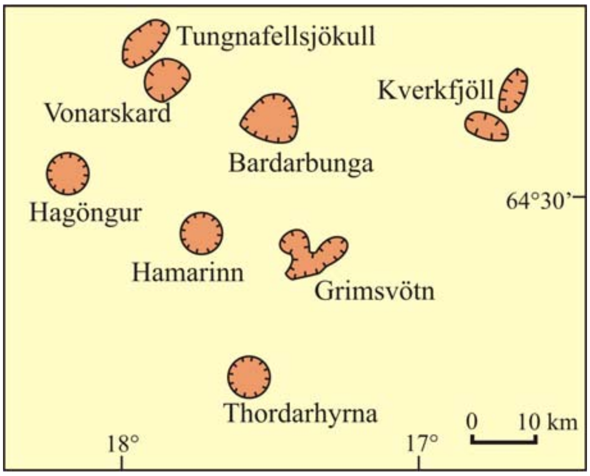

Figure 2. Map of the 8 volcanoes modeled in this study. Data from Johannesson and Saemundsson (1998).

as a double magma chamber (Gudmundsson, 2006). The deeper chamber, referred to as a reservoir, can be inferred from general mechanical considerations as well as from petrological and geodetic data. Thus, lava shields and many other monogenetic basalt volcanoes, most of which are of primitive composition (olivine-tholeiite) and reach volumes of 25 km^3 in a single eruption (Thordarson and Larsen, 2007), must come from deep-seated and large reservoirs of primitive magma. Recently, geodetic data from several rift-zone central volcanoes also indicate the existence of large, deep-seated reservoirs perhaps underlying entire volcanic systems (Sturkell et al., 2006).

There is abundant geodetic (Sturkell et al., 2006) and petrologic (Sigmarsson and Steinthorsson, 2007; Thordarsson and Larsen, 2007) data indicating that shallow, crustal chambers exist under many and perhaps most active central volcanoes. Also, general mechanical considerations indicate that all central volcanoes containing collapse calderas have, or had at the time of caldera formation, shallow chambers. And many extinct chambers occur in the roots of deeply eroded Pleistocene and Tertiary central volcanoes where they are represented by plutons of gabbro, granophyre, or both (Gudmundsson, 2006).

To test the possibility of a mechanical interaction between central volcanoes, we selected the part of the Neovolcanic Zone in Iceland that is above the centre of the Iceland Plume, where the East Volcanic Zone meets the North Volcanic Zone, and has the highest density of active central volcanoes (Figures 1, 2). These volcanoes are Tungnafellsjökull, Vonarskard, Hagöngur, Bardarbunga, Hamarinn, Grimsvötn, Thordarhyrna, and Kverkfjöll. All these volcanoes are presumably supplied with magma from shallow chambers.

Some of the volcanoes are among the most active in Iceland. During historical time (the past 1100 years), the Grimsvötn Volcano (Figures 1, 2) has erupted about 70 times (Thordarson and Larsen, 2007). This makes Grimsvötn by far the most active central volcano in Iceland, in terms of eruption frequency in historical time. During the same period, Bardarbunga (Figures 1, 2) has erupted at least 23 times (Thordarson and Larsen, 2007). This makes Bardarbunga, together with the Hekla Volcano (Figure 1), the second most active central volcano in Iceland in terms of eruption frequency (Thordarson and Larsen, 2007). There have been as many as five conincident eruptions in more than one of these volcanoes (Larsen et al., 1998). A number of authors (Einarsson et al., 1997; Sigmarsson et al., 2000) have documented a connection between the 1996 and 1998 eruptions within this field, on Gjalp (part of the Grimsvötn volcanic system) and Grimsvötn volcanoes. Evidence from isotope composition and seismicity indicate "dike sharing" between these two, as well as propagating fractures between them.

Gravity measurements indicate a partially molten magma chamber beneath many of these volcanoes (Gudmundsson and Högnadottir, 2007). Magma chambers in Iceland are known to be at depths between 1.5 and 3 km to the roof of the chamber (Gudmundsson, 2006). Five of the volcanoes are located beneath the ice sheet of Vatnajökull, but have been studied in detail by geophysical methods (Gudmundsson and Högnadottir, 2007). Most and perhaps all these volcanoes have crustal magma chambers and many contain collapse calderas. One volcano, Kverkfjöll, has a double caldera, and another one, Grimsvötn, a triple caldera (Johannesson and Saemundsson, 1998; Gudmundsson and

Högnadottir, 2007). We here regard Tungnafellsjökull and Vonarskard as two central volcanoes, each with a caldera (Johannesson and Saemundsson, 1998). Since the calderas are very close some authors consider them a double caldera of a single central volcano, referred to as Vonarskard (Gudmundsson and Högnadottir, 2007). In our mechanical analysis, however, it is immaterial whether this volcano is regarded as one or two. Excluding double and triple calderas, the average distance between the nearby volcanoes is about 30 km (Figure 2).

An active collapse caldera together with a shallow magma chamber makes these volcanoes act mechanically as holes (cavities in three dimensions) in the elastic crustal plate. All such holes concentrate stresses (Gudmundsson, 2006), so that to understand the potential of mechanical interaction between the volcanoes, they can, to a first approximation, be modeled as holes in an elastic plate subject to plate pull.

3. Modeling

For the selected 8 volcanoes (Figure 2), we made many numerical models using the finite-element program Ansys (www.ansys.com; Logan, 2002). To make the analysis as simple as possible, our models are two-dimensional and each volcano is modeled as a hole. We have also made some three-dimensional models, with the magma chambers modeled as oblate spheroids, to compare with the two-dimensional models. Models were made with magma chambers all at the same depth, and at varying depths; the results are basically very similar. We therefore conclude that the two-dimensional models give as accurate results regarding the stress fields between the volcanoes as are needed for the present purpose of analyzing their mechanical interaction.

In all the models we used a homogeneous, isotropic crust with a Young's modulus of 20 GPa and a Poisson's ratio of 0.25, both values being typical for the uppermost part of the crust in Iceland (Gudmundsson, 2006). In all the models, we applied a tensile stress of 5 MPa (equal to the maximum in situ tensile strength of typical solid rocks) in a direction parallel with

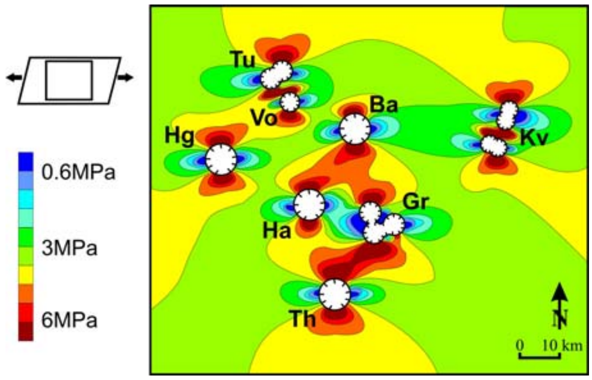

Figure 3. Finite-element model of the tensile stresses, in mega-pascals, around the 8 volcanoes in Figure 2. A homogeneous, isotopic crust with a Young's modulus of 20 GPa and a Poisson's ration of 0.25 is used, under a force of 5 MPa tension. The spreading vector (applied tension) is depicted.

the spreading vector in this part of Iceland, that is, in the direction N105°E. The location and geometry of each caldera are taken from a general geological map of the area (Figure 2).

Using these boundary conditions as a basis, as well as the geometries of the calderas (and inferred shallow magma chambers), we made models that belong to 4 main classes. In the first class, all the calderas have the same diameter (8 km), but where double calderas exist they are modeled as such. In the second class, double calderas are modeled as single and all the calderas have the same size. In the third class, the double calderas are shown as single but with a diameter of 12 km (instead of the standard 8 km). And in the fourth class, the geometries of the calderas are not so much idealized but rather modeled as presented on the map.

4. Mechanical interaction

The basic results of all the model classes are the same: there is strong mechanical interaction in terms of zones of stress concentration between all the nearby central volcanoes. Here we show results from the fourth class of models where the geometry is most similar to that inferred for the real volcanoes (Figure 2). There are clear tensile-stress concentration zones between the nearby volcanoes (Figure 3). The existence of these

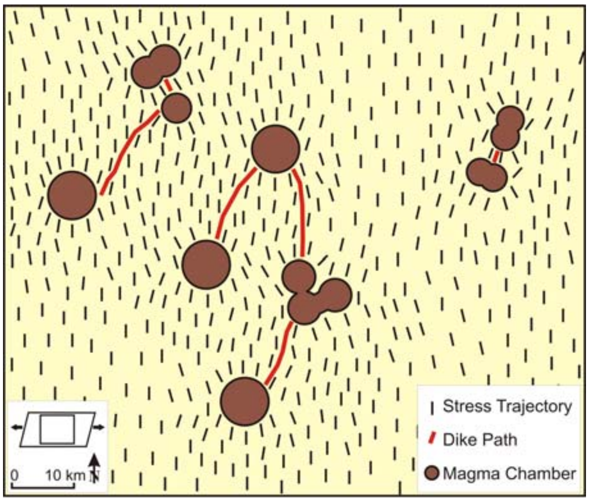

Figure 4. Trajectories (trends) of the maximum principal compressive stress, σ_1, around the 8 volcanoes in Figure 2. Ideal dikes propagate parallel with σ_1. Potential shared dikes between some volcanoes are indicated (very schematically) by thick, red lines.

zones indicates that unrest in one volcano can trigger unrest in a nearby volcano. This is further supported by the fact that between the same volcanoes there are zones of shear-stress concentration (not shown here), suggesting that seismogenic faults may propagate or be shared between the volcanoes. Here, however, the focus is on the zones of tensile stress concentration because they are likely to encourage the emplacement of shared dikes.

For a dike to be shared between two volcanoes, it is necessary that there be tensile-stress concentration zone between the volcanoes, but it is not sufficient. In addition, the directions of the principal stresses, that is, the stress trajectories in the high-stress zone between the volcanoes must be of suitable orientation.

Since most dikes are extension fractures (Gudmundsson, 2006), it follows that they generally follow the trajectories of the maximum principal compressive stress, σ_1. We have plotted the σ_1 – trajectories between all the volcanoes for the model in Figure 3. The results (Figure 4) show that for the volcanoes that are very close, such as Tungnafellsjökull and Vonarskard, and of course the double caldera in Kverfjöll, the conditions for dike sharing between the volcanoes is presumably commonly satisfied during unrest periods.

Interestingly, the conditions for dike sharing between some of the central volcanoes that are comparatively far from each other are also satisfied (Figure 4). For example, there is not only a zone of tensile stress between Vonarskard and Hagöngur (Figure 3), but the σ_1 – trajectories would generally encourage dike sharing between these central volcanoes. Zones of higher tensile stress also exist between the volcano pairs Bardarbunga and Hamarinn and Bardarbunga and Grimsvötn (Figure 3). The σ_1 – trajectories indicate that dike sharing between both volcano pairs is possible (Figure 4). However, the magnitudes of the tensile stresses in these zones (Figure 3) appear to be more favorable for dike sharing between Bardarbunga and Hamarinn. The highest tensile stresses between central volcanoes that are comparatively far apart occur in a zone between Grimsvötn and Thordarhyrna (Figure 3). The tensile stresses in this zone exceed the common in situ tensile strength of rocks, 0.5-6 MPa (Haimson and Rummel, 1982; Schultz, 1995). Also, the σ_1 – trajectories between these volcanoes are generally favorable for dike sharing.

5. Discussion and conclusion

The numerical models presented above indicate that there is likely to be mechanical interaction between closely spaced central volcanoes in a rift-zone volcanic field such as the one in Figure 2. This interaction is primarily through stress concentrations between nearby volcanoes so as to favor dike sharing during unrest periods. There is considerable observation support for this model prediction. We mention the following points.

First, Gudmundsson and Högnadottir (2007) show that there are ridges in the gravity field between the central volcano pairs Vonarskard-Hagöngur, Bardarbunga-Hamarinn, and Grimsvötn-Thordarhyrna (Figure 2). They interpret the gravity ridges to be dense dike swarms, indicating that the central volcano pairs are connected by dikes. Such a connection would normally imply at least some dike sharing between the two volcanoes constituting the each pair.

Second, some authors have suggested that during the 1996 Gjalp eruption in the Vatnajökull Ice Sheet (Gudmundsson et al., 1997) there was lateral propagation of a dike from the Bardarbunga Volcano to the Grimsvötn Volcano (Einarsson et al., 1997). Although the results are not conclusive in that the earthquake data generally support such a dike propagation while the petrological data speak against it (Sigmarsson et al., 2000), the possibility exists. Our mechanical models (Figures 3,4) indicate that such a propagation or dike sharing is entirely possible between Bardarbunga and Grimsvötn and even more so between Grimsvötn and Thordarhyrna (Figures 3, 4).

Third, many authors have suggested magma mixing through lateral propagation of dikes between volcanoes in the EVZ (Blake, 1984; McGarvie, 1984). These ideas are only suggestive and primarily based on petrological considerations. They need to be refined and tested. One way to do so is to provide a detailed mechanical analysis of the feasibility of lateral propagation or other type of dike sharing between far-away central volcanoes.

In conclusion, closely spaced central volcanoes in a rift zone subject to plate pull may develop stress fields giving rise to strong mechanical interactions between some of the volcanoes. Zones of shear-stress concentration connecting a volcano pair may encourage simultaneous earthquake activity in the volcanoes constituting the pair. Also, zones of tensile-stress concentration connecting such a pair may promote either lateral propagation or other type of dike sharing between the two volcanoes.

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13 Appendix III

Volcanoes as elastic inclusions: their effects on the propagation of dykes, volcanic fissures, and volcanic zones in Iceland

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Abstract
Mechanically, many volcanoes may be regarded as elastic inclusions, either softer (with a lower Young's modulus) or stiffer (with a higher Young's modulus) than the host-rock matrix. For example, many central volcanoes (stratovolcanoes, composite volcanoes) are composed of rocks that are softer than the crustal segments that host them. This is particularly clear in Iceland where central volcanoes are mostly made of soft rocks such as rhyolite, pyroclastics, hyaloclastites, and sediments whereas the host rock is primarily stiff basaltic lava flows. Most active central volcanoes also contain fluid magma chambers, and many have collapse calderas. Fluid magma chambers are best modelled as cavities (in three dimensions) or holes (in two dimensions), entire calderas as holes, and the ring faults themselves, which commonly include soft materials such as breccias, as soft inclusions. Many hyaloclastite (basaltic breccias) mountains partly buried in the basaltic lava pile also function as soft inclusions. Modelling volcanoes as soft inclusions or holes, we present three main numerical results. The first, using the hole model, shows the mechanical interaction between all the active central volcanoes in Iceland and, in particular, those forming the two main clusters at the north and south end of the East Volcanic Zone (EVZ). The strong indication of mechanical interaction through shared dykes and faults in the northern cluster of the EVZ is supported by observations. The second model, using a soft inclusion, shows that the Torfajökull central volcano, which contains the largest active caldera in Iceland, suppresses the spreading-generated tensile stress in its surroundings. We propose that this partly explains why the proper rift zone northeast of Torfajökull has not managed to propagate through the volcano. Apparently, Torfajökull tends to slow down the rate of southwest propagation of the rift-zone part of the EVZ. The third model, again using a soft inclusion, indicates how the lateral propagation of two segments of the 1783 Laki fissure became arrested in the slopes of the hyaloclastite mountain Laki.

Keywords: central volcanoes, volcanic fissures, caldera, crustal stresses, dyke arrest.

1. Introduction

An elastic inclusion may be defined as a body with material properties that contrast with those of the surrounding (hosting) material, commonly referred to as the matrix. While the term "inclusion" in this sense is well established in the classical elasticity and rock mechanics literature (Savin, 1961; Jaeger and Cook, 2007), the more recent literature on micromechanics refers to elastic inclusions, as defined above, as "inhomogeneties" (Nemat-Nasser and Hori, 1999). Here, however, we use the term "inclusion" as meaning a material body hosted by a larger body with different elastic properties (Gudmundsson, 2006).

A principal measure of the difference in elastic properties between the inclusion and its matrix is Young's modulus, commonly referred to as "stiffness". It is the ratio between stress and strain in the one-dimensional Hooke's law, and is thus indicated by the slope of the stress-strain curve (Hudson and Harrison, 1997; Jaeger and Cook, 2007). The inclusion can be either stiffer or softer than its matrix; if the inclusion contains only fluid, its Young's modulus is zero and it may be regarded as a cavity (a three-dimensional structure) or a hole (a two-dimensional structure).

A volcano, including its magma chamber, may be regarded mechanically as an inclusion. An active volcano, whether polygenetic (stratovolcano, collapse caldera, central volcano) or monogenetic (lava shield, crater cone, hyaloclastite mountain), has normally elastic properties that differ from those of its host rock (say a lava pile) and therefore acts as an inclusion. Similarly, an active magma chamber is either totally or partly fluid, and thus acts as a fluid-filled cavity (for a two-dimensional structure, a hole) or, when partly solidified, as a soft inclusion. When the chamber is completely solidified it becomes a pluton whose elastic properties normally differ from those of the host rock; thus the pluton acts also as an inclusion. Inclusions and cavities in an elastic body change the stress fields in their vicinity and, commonly, act as stress raisers; that is, they concentrate (increase) the magnitudes of the stresses in the host rocks immediately surrounding them.

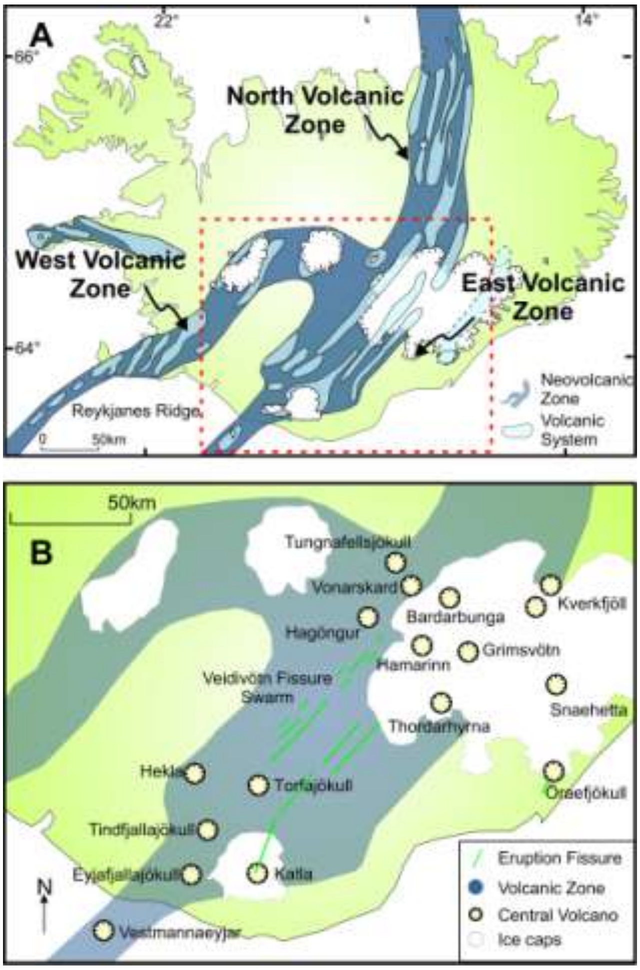

Fig. 1: A: Map of Iceland showing the Neovolcanic Zone, the volcanic systems, and the main ice sheets. The three main segments of the Neovolcanic Zone, the North Volcanic Zone, the East Volcanic Zone, and the West Volcanic Zone are marked. Also shown is the outline of Fig. 1B. B: Map of the East Volcanic Zone showing the central volcanoes and some of the main Holocene volcanic fissures.

Using mechanical models as a basis, we have recently suggested that some central volcanoes in close proximity form clusters that interact mechanically (Gudmundsson and Andrew, 2007). Our models were made particularly for a cluster of volcanoes in central Iceland, where there is abundant evidence of interactions such as shared dykes and volcanic fissures. However, geodetic measurements in other volcanic areas also indicate mechanical interactions between volcanoes in close proximity, such as on the Galapagos Islands (Amelung et al., 1997).

Here we extend our studies of the mechanical effects of, and interaction between, volcanoes in Iceland in several ways. First, we now expand the studies so as to include all the main central volcanoes in the East Volcanic Zone (EVZ) of Iceland (Fig. 1). For this part we focus on the magma chambers of the central volcanoes, and thus model them as fluid-filled holes. Second, we analyse the effects of a large collapse caldera, the Torfajökull central volcano, modelled as a soft inclusion, on the propagation of the entire EVZ (Fig. 1). This volcanic zone is a proper rift zone northeast of the Torfajökull central volcano, but a flank zone lacking in rift-zone characteristics southwest of the volcano. We present numerical models as to how the volcano may have, temporarily at least, arrested the southwest propagation of the EVZ. Third, we study the effects of the volcanoes themselves, at or close to the surface, on the propagation and arrest of laterally propagating volcanic fissures. For this we focus on the effects of the hyaloclastite mountain Laki, modelled as soft inclusion, in arresting the lateral propagation of volcanic fissures.

2. Volcanotectonic background

Volcanism in Iceland is mainly confined to the Neovolcanic Zone, defined as that part of Iceland containing rocks belonging to the Brunhes magnetic epoch (<0.8Ma). The Neovolcanic Zone consists of three subzones: the North Volcanic Zone (NVZ), the West Volcanic Zone (WVZ), and the East Volcanic Zone (EVZ). The main geological features of the Neovolcanic Zone are the volcanic systems, of which there are about 30 (Fig. 1). A volcanic system normally features a central volcano or a fissure (dyke) swarm or both, and has a typical lifetime of 0.5-1.5Ma (Saemundsson, 1978, 1979; Jakobsson et al 1978; Jakobsson, 1979). When a central volcano is present within a volcanic system, it acts as a focal point for the eruptive activity, and is normally the largest edifice within the system (Thordarson and Larsen, 2007). Of these volcanic systems, 19 contain 23 central volcanoes, indicating that some systems hold more than one central volcano (Thordarson and Larsen, 2007).A volcano is defined as central when it exhibits the following characteristics: it erupts frequently, it extrudes basaltic, intermediate and acid lavas and other eruptive materials, it is associated with and fed by a shallow crustal magma chamber, and it is often associated with a collapse caldera (Gudmundsson, 1995). A central volcano which forms a part of a rift-zone volcanic system is also associated with well-defined swarms of tension fractures, normal faults and volcanic fissures, and with dykes at deeper crustal levels (Gudmundsson, 2000).

The presence of shallow crustal magma chambers beneath many active central volcanoes in Iceland has been confirmed, by both petrologic (Thordarson and Larsen, 2007; Sigmarsson and Steinthorsson, 2007) and geodetic (Sturkell et al, 2006) data. Also, in many deeply eroded palaeorift zones, such as in the Tertiary lava pile of East and Northwest Iceland, there are many plutons at depths of 1.5-2 km below the original surfaces of the rift zones. These plutons, of gabbro and granophyre, show clear evidence of having acted as shallow chambers supplying magma to the associated central volcanoes (Gudmundsson, 1995, 2006). Many of the central volcanoes contain one or more collapse calderas in conjunction with the shallow magma chamber.

Within the Neovolcanic Zone, the volcanoes are not evenly distributed. This is particularly clear as regards central volcanoes in the EVZ (Figs. 1, 2). Clearly, the central volcanoes form two clusters. One is located inside and in the vicinity of Vatnajökull ice cap at the northeastern end of the EVZ; the other in the southwestern part of the EVZ. In between these two clusters the main volcanoes are crater rows, such as the 1783 Laki volcanic fissure (Fig. 7). There are various other volcanic types that are unevenly distributed within the Neovolcanic Zone. A good example is the presence of Holocene lava shields in the NVZ and WVZ, whilst there are almost none (the exception being the small shield on Surtsey in the Vestmannaeyjar) in the EVZ (Rossi, 1996; Rossi, 1997; Andrew and Gudmundsson, 2007). By contrast the EVZ, at present, is the location of all the really large Holocene fissure eruptions, such as the Eldgja eruption in 934AD and the 1783 eruption of Lakagigar (Thordarson and Self, 1993; Thordarson and Larsen, 2007).

The volcanic activity among the volcanic zones is also variable. Thus, the EVZ has a greater volcanic activity in historical time (the past 1100 years) than either the WVZ or the NVZ, in terms of both volume of materials erupted and frequency of eruptions (Thordarson and Larsen, 2007). Of the volcanic systems in the EVZ (Fig. 1) all but one, Tindfjöll, have been active in historical time, and 4 of these (Grimsvötn, Veidivötn, Hekla and Katla) are the most active in the whole country. There have been 172 verified volcanic events within historical time in Iceland, and 80% have been within the EVZ, the majority concentrated on the aforementioned systems. Also, while the recurrence period at most central volcanoes in the Neovolcanic Zone is in the order of several centuries, for the 4 most active central volcanoes within the EVZ the recurrence period is in the order of years to decades (Thordarson and Larsen, 2007).

The volcanotectonic framework of the EVZ differs from the WVZ and the NVZ in several other aspects. First, the mantle plume is located under its northeastern part (beneath the Vatnajökull ice cap) and has clearly large effects on the geochemistry of erupted materials and the volcanotectonic activity in that part. In particular, this arguably accounts for the high eruption frequency of several volcanoes in this part. Second, the EVZ is the youngest segment of the Neovolcanic Zone. The oldest parts of the EVZ are about 3 Ma, whereas the NVZ and the WVZ have been active for many million years (Saemundsson, 1979). Third, the EVZ is thought to be rifting in its northern half, down to the central volcano Torfajökull. To the southwest of Torfajökull, the EVZ lacks volcanotectonic elements typical of rift zones, such as normal faults, grabens and long volcanic fissures, and that part of it is regarded as a flank zone slowly propagating southwest, with the Vestmannaeyjar volcanic system at its tip (Fig. 1). This propagation through an existing crust has accounted for the construction of the entire EVZ. The WVZ forms an overlapping configuration with the EVZ, which may eventually take up all the rifting in South Iceland whereby the WVZ becomes extinct (Gudmundsson, 2007; Thordarson and Larsen, 2007).

The difference between the rift zone proper and the propagating rift zone is not only related to the volcanotectonic elements discussed above but can also be traced through a compositional change in eruptive products along its length (Jakobsson, 1972; Meyer et al, 1985). Thus, the rift-zone part of the EVZ is characterised by tholeiite basalts which, on entering the non-rifting southwestern part in the south change to primarily alkaline basalts. The EVZ, particularly its southwestern part, has also a thicker crust than the normal rift-zone segments of the Neovolcanic Zone (Allen et al., 2002).

3. Modelling

In order to examine the effects of the volcanoes as inclusions and cavities on the volcanotectonic activity in their vicinities, two different types of models were run. The first type models the magma chambers associated with the central volcanoes in the two main clusters in the EVZ (Figs. 1, 2) as holes. This approximation is justified because when the chambers are filled with fluids (mainly magma) that are initially in lithostatic equilibrium with

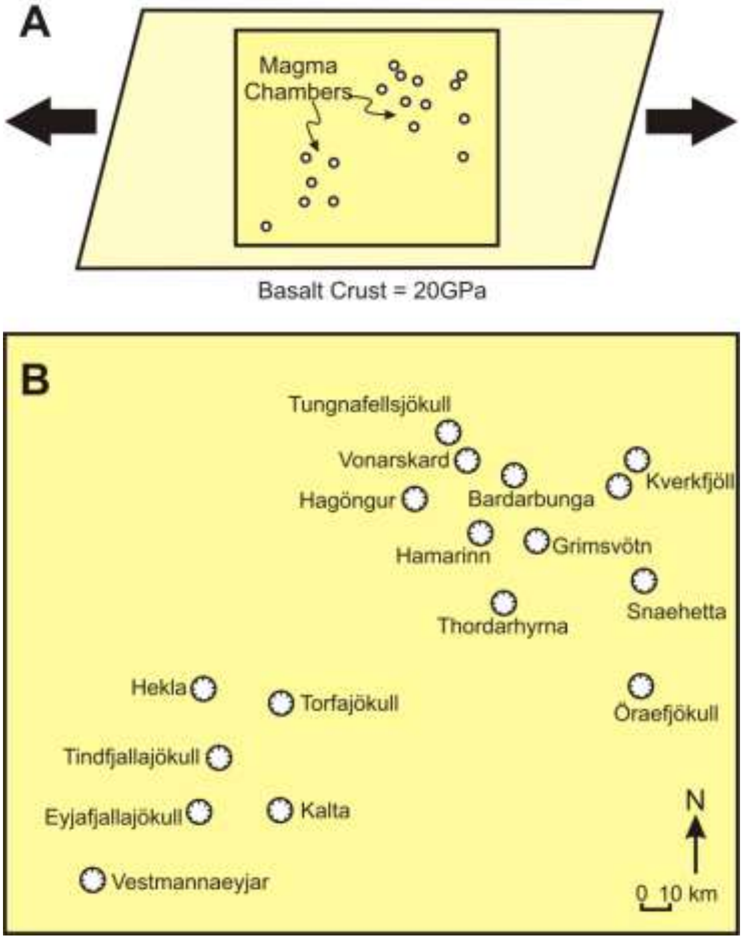

Fig. 2: A: Model outline showing the location of the central volcanoes of the East Volcanic Zone (modelled as holes) in the basaltic crust. Note the direction of the tension applied to the models demonstrated by the arrows; this is a tensile stress of 5MPa parallel with the spreading vector in this part of Iceland. The inset box shows the area shown in the model results (Fig. 4). B: Map showing the locations and names of the modelled central volcanoes, based on Johannesson and Saemundsson (1998).

the host rock, a hole model is appropriate. This follows particularly since the magma chambers are located in a brittle lithosphere that is very wide compared with its thickness (Fig. 2A).

The second type of models considers the volcanoes themselves, rather than their magma chambers, in which case a fluid-filled cavity or a hole is not an appropriate model. For this second type, we take into consideration the different elastic properties of the volcanoes and their host rocks, and model the volcanoes as elastic inclusions with the host rock as a matrix. Here the focus is on the effects of two types of volcanoes. One is the monogenetic hyaloclastite mountain, as exemplified by the hyaloclastite mountain Laki (Fig. 7). We are interested in assessing the effects of this mountain, modelled as soft inclusion, in arresting the lateral propagation of Laki volcanic fissures. The other type is the central volcano with a collapse caldera, here exemplified by the central volcano, Torfajökull, which has the largest collapse caldera within the Neovolcanic Zone (Fig. 7). Here we are interested in the effects that this volcano and its caldera, modelled as soft inclusions, have on the propagation of the EVZ. We present numerical models as to how the volcano may have, temporarily at least, arrested the southwest propagation of the EVZ.

All the models are numerical and use the finite-element program Ansys (www.ansys.com; Logan, 2002). Numerical models were used as they allow an examination of the local stress fields surrounding the volcanoes modelled as holes and inclusions. The host rock, that is, the matrix, is modelled as a homogeneous isotropic crust with a Young's modulus of 20GPa, and a Poisson's ratio of 0.25, as being typical for the uppermost crust in Iceland (Gudmundsson, 2006). A tensile strength of 5MPa, equal to the tensile strength of typical solid rocks (Haimson and Rummel, 1982; Schultz, 1995; Amadei and Stephansson, 1997), was applied to all models, parallel to the spreading vector N105^{o}E in the south part of Iceland (DeMetz et al., 1990, 1994).

4. Mechanical interaction

The first series of models examines the mechanical interaction between central volcanoes, focusing on their magma chambers modelled two-dimensionally as holes within the elastic crust. Some previous models using three-dimensional cavities for the chambers yielded stress-concentration zones between chambers that were similar to those obtained from hole models (Gudmundsson and Andrew, 2007). For this reason, and those discussed above, we regard the two-dimensional hole models as adequate for the present analysis.

All volcanoes were modelled as 8km diameter holes. For the volcanoes Kverkfjöll and Vonarskard, two 8km diameter holes were used to represent their double calderas. The locations of the volcanoes were taken directly from a general geological map of Iceland (Johannesson and Saemundsson, 1998) (Fig. 2). Many models were run; some concentrated on whole segments of the Neovolcanic Zone, such as the EVZ which contains two clusters of volcanoes, and others included the entire Neovolcanic Zone with all its central volcanoes. Series of models were also run using the true ring-fault geometries for the calderas where they were known. The use of circles to represent the central volcanoes yielded essentially the same interaction as when the actual ring-fault geometries were used (Gudmundsson and Andrew, 2007; Gudmundsson et al., 2008). Thus for the present purpose the circles may be regarded as sufficiently accurate presentation of the central

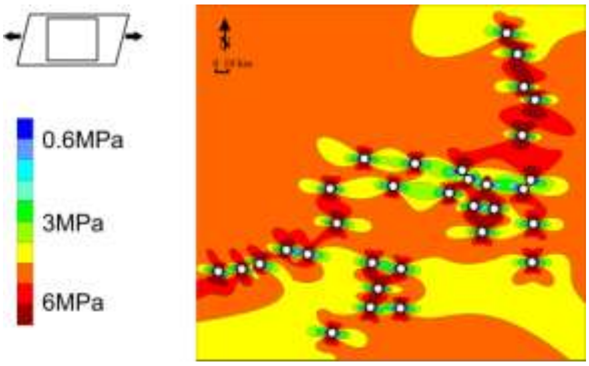

Fig. 3: Numerical model of the tensile stresses surrounding all the central volcanoes of the Neovolcanic Zone. The stress magnitudes are shown in mega-pascals. The properties of this model are the same as those shown in Fig. 2. There are stress-concentration zones and, therefore, likely mechanical interaction between volcanoes in all the volcanic zones. In this paper, however, the focus is on the interactions between volcanoes in the East Volcanic Zone.

volcanoes and the associated ring faults, and are thus used in this paper. In the models, some volcanic systems were given central volcanoes, where they are strictly not present. This has been done in order to make the models more complete, and to represent activity within all the branches of the Neovolcanic Zone. This is, for example, the case for the Vestmannaeyjar Volcanic System where, at present, there is no clear central volcano but where one is apparently developing (Mattson and Hoskuldsson, 2003).

The models show that mechanical interaction is a common feature of the central volcanoes throughout the Neovolcanic Zone (Fig. 3), but the patterns in the EVZ are of particular interest. The models of the EVZ highlight two clear clusters of central volcanoes, one at the northern end of the volcanic zone, and a second one in its southern part. The space in between these two clusters is characterised by the earlier mentioned large volcanic fissures, such as Eldgja and Lakagigar (Fig. 7). Both of the clusters of volcanoes show mechanical interactions among the volcanoes that constitute them (Fig. 4). In the numerical models, the interaction between the volcanoes is indicated by zones of high tensile stress between the volcanoes, indicating the possibility of an activity in one volcano triggering unrest in another volcano in the same cluster. A simultaneous activity in two or more volcanoes

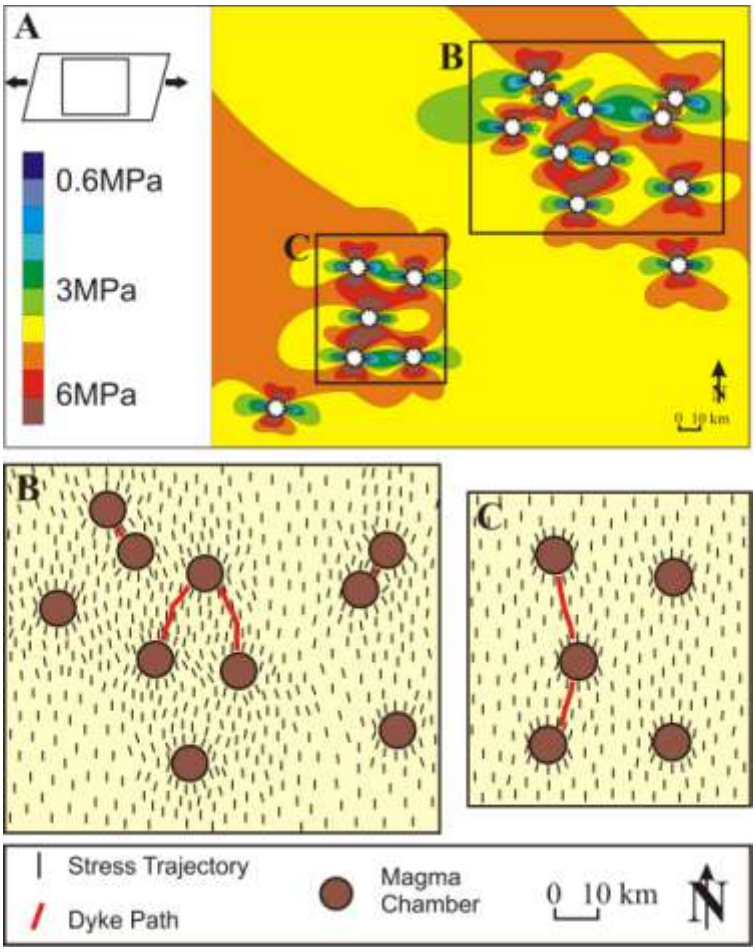

Fig. 4: A: Numerical model of the tensile stresses surrounding the central volcanoes in the East Volcanic Zone. Stresses are shown in mega-pascals. Boxes mark the areas shown in sections B and C of the figure. B: Numerical models of the stress trajectories (trends) of the maximum principal compressive stress (σ_1) around the volcanoes in the northern cluster. Thick red lines depict possible normal fault and dyke propagation paths between the volcanoes. C: Numerical models as in section B for the southern cluster of volcanoes. The properties of all of these models are given in Fig. 2.

could result in shared seismic swarms or dyke injections.

In order for two volcanoes to "share" a dyke, for example, there must be a tensile-stress concentration zone connecting the volcanoes. In addition to this, the stress trajectories for the maximum principal compressive stress σ_1 must be of an orientation allowing a dyke path between the two volcanoes (Figs. 4B, C). The stresses that allow this to occur show certain characteristics highlighted in these models. The greatest potential for dyke sharing is when the zones of high-tensile stress concentration, as well as the orientation of σ_1, trend roughly NE-SW in the EVZ and N-S in the NVZ (Figs. 3, 4). These trends coincide with those of all the major structural elements in the EVZ and the NVZ, including volcanic fissures, tension fractures, normal faults, and the volcanic systems themselves, as has long been recognised (Thorarinsson, 1966) The volcanic zones themselves also have roughly these trends that, although oblique to the direction of the spreading vector, are roughly perpendicular to the direction of local minimum compressive (maximum tensile) principal stress, σ_3.

The cluster at the northern end of the EVZ (and partly at the southern end of the NVZ) is located mainly over the mantle plume, which may, to a certain extent, explain the clustering (Einarsson et al, 1997; Larsen et al, 1998; Gudmundsson and Andrew, 2007; Gudmundsson and Högnadottir, 2007;). There is a clear

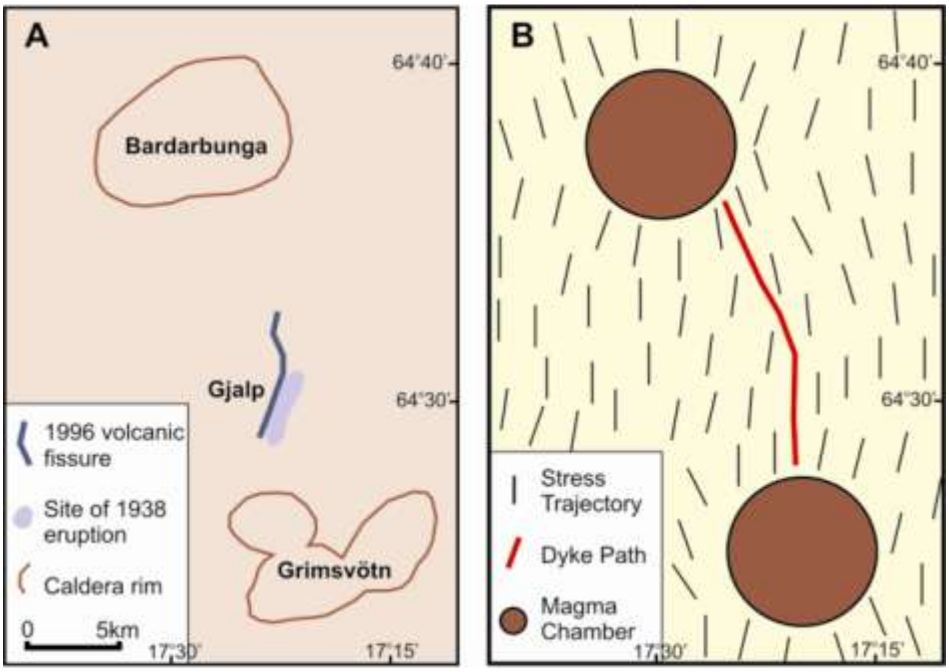

Fig. 5: A: Map showing the location of the 1996 Gjalp eruption, and another eruption in 1938 between the Grimsvötn and Bardarbunga volcanoes. Map adapted from Gudmundsson et al. (2002). B: Part of the numerical model of the trajectories of the maximum principal compressive stress (σ_1) around the central volcanoes Grimsvötn and Bardarbunga (taken from Fig. 4B). The thick red line depicts the possible path between the volcanoes, and allows comparison with the Gjalp eruption.

mechanical interaction between the volcanoes in a very close proximity, such as between Vonarskard and Hagöngur, Bardarbunga and Hamarinn, and Bardarbunga and Grimsvötn (Fig. 4). But there is also a mechanical interaction between Thordarhyrna and Grimsvötn, even if these volcanoes are comparatively far apart (Figs. 3, 4). The reason for this clear interaction between Thordarhyrna and Grimsvötn is that a line connecting them has a clear northeast trend and thus coincides with the trend of the volcanic zone itself and the most favourable orientation for the generation of stress-concentration zones between volcanoes in the EVZ.

The high level of volcanic activity in the northern cluster of the EVZ, under the Vatnajökull ice cap, is in direct contrast with the activity on either side of it. Eruption frequencies in the northern cluster are higher than in the adjacent parts of the NVZ and the EVZ (Larsen et al, 1998). Eruptions in rift-zone volcanic systems are directly linked to rifting evens, that is, rupture of the plate boundary and rapid local plate movements (temporary excess spreading rates), and are commonly confined to a single system at any one time. However, in the northern cluster of the EVZ, near concurrent or even simultaneous activity on two or more systems is known to have happened (Larsen et al., 1998; Thordarson and Larsen, 2007), with as many as five coincident eruptions in two or more of these volcanoes (Larsen et al., 1998). The cluster has a pattern of activity possibly linked to either a magma flux from the mantle, or controlled by an interaction of the local tectonics and the mechanical properties of the plates. The latter of these indicates that the activity could then be determined by the spreading rate, and the crustal strength and thickness (Larsen et al., 1998).

Further to the frequency of activity, many studies on volcanoes within this cluster discuss the case of the 1996 Gjalp eruption between the volcanoes Bardarbunga and Grimsvötn as a lateral propagation of a dyke (Fig. 5; Einarsson et al., 1997). Whilst petrological data, however, disagree with this idea (Sigmarsson et al., 2000), our models indicate that the stress-concentration zone between Bardarbunga and Grimsvötn encourages shared dykes, irrespective of whether they are primarily injected laterally or vertically. Further evidence for the favourable stress field for mechanical interaction between these volcanoes, and the potential for sharing dykes, comes from studies of the hyaloclastite ridge associated with the 1996 Gjalp eruption (Gudmundsson et al., 2002, Gudmundsson et al., 2004; Gudmundsson et al., 1997) (Fig. 5). Linear extrusive (ridges) and intrusive (dyke swarms) connections between central volcanoes in the area are not limited to the one located between Bardarbunga and Grimsvötn. Gravimetric studies by Gudmundsson and Högnadottir (2007) show dyke swarms extending southwest from Vonarskard towards Hagöngur, from Bardarbunga to Hamarinn, and from Grimsvötn to Thordarhyrna (Fig. 6). The main difference between these three dyke swarms and the ridge between Bardarbunga and Grimsvötn is the structural trend: whereas the ridge between Bardarbunga and Grimsvötn trends roughly north-south (Fig. 5), the three dyke swarms trend northeast and are thus parallel with the trend of the EVZ (Fig. 1) and some of the tensile stress-concentration zones between the volcanoes (Fig. 4).The gravimetric study of Gudmundsson and Högnadottir (2007) also highlighted some other characteristics of the central volcanoes under Vatnajökull that are relevant to the present study. They note that the gravity anomalies associated with the larger volcanoes are significantly stronger than those associated with the smaller volcanoes such as Hagöngur, Hamarinn and Thordarhyrna. Each of these small volcanoes shares a link (a hyaloclastite ridge) with a larger central volcano within the same volcanic system, lies in the southern part of the system, and lacks a collapse caldera. Gudmundsson and Högnadottir (2007) explain these characteristics in terms of magma

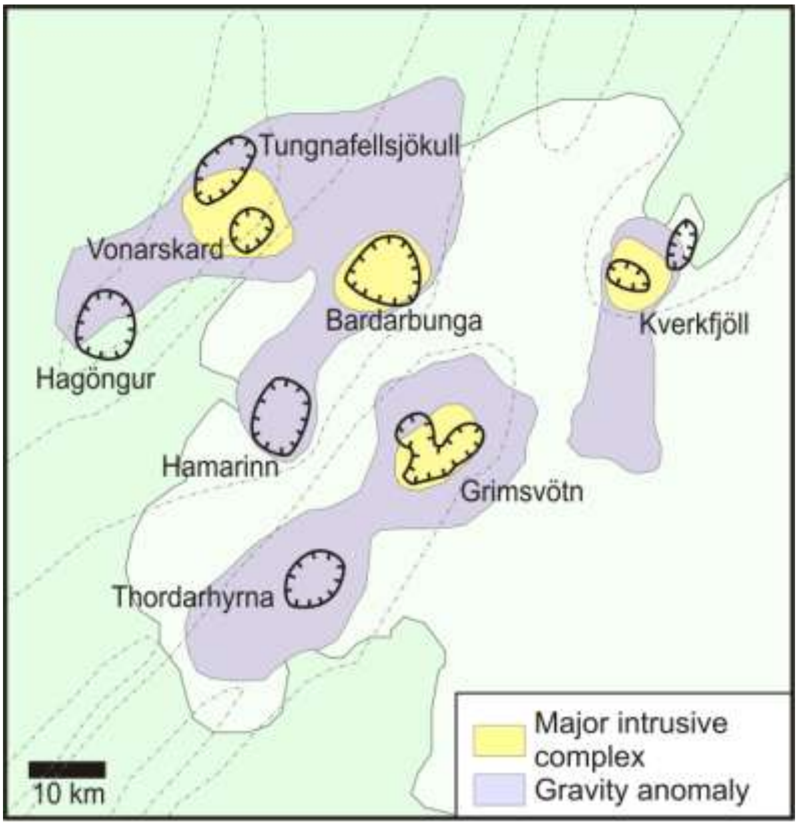

Fig. 6: Map outlining the dyke and sheet swarms (marked as intrusive complexes) underneath and between the central volcanoes in the northern cluster of the East Volcanic Zone. Vatnajökull ice sheet is shown in white, the volcanic systems marked by dashed grey lines. Map adapted from Gudmundsson and Högnadottir (2007).

supply; they argue that two central volcanoes in one volcanic system may indicate a high magma supply rate for at least part of the lifetime of the system. One of the volcanoes becomes dominant and serves as the main conduit for magma in the system. The dominant volcano creates a shallow magma chamber (and thus the potential for a caldera), whilst the second volcano receives only a sporadic supply of magma, commonly through dykes shared with the dominant volcano. Alternatively, both central volcanoes may be supplied with magma, either directly or through a shallow magma chamber, from reservoirs in the lower crust or at the crust-mantle boundary (Gudmundsson, 2000). The presence of two central volcanoes within the same volcanic system may be a reflection of the hydrodynamic situation in the mantle below the volcanic system, and clearly indicates a mature system.

The dominance of some central volcanoes can also be detected in the southern cluster of the EVZ (Figs. 1, 2) For this cluster, the numerical models show a clear stress-concentration zone, and thus a likely mechanical interaction, between Eyjafjallajökull and Tindfjallajökull, and then extending north to Hekla (Fig. 4). The central volcano in Myrdalsjökull, Katla, has a stress-concentration zone linking it to only one of its neighbours, Eyjafjallajökull (Fig. 4A). However, this zone trends east-west, and thus at a high angle to the general trend of the EVZ and the most likely dyke-sharing direction. This is confirmed by the trajectories of σ_1 which trend north to north-northeast and are thus unfavourable for dyke sharing in an east-west direction (Fig. 4C). This would suggest that Katla has no mechanical links as regards dyke sharing with other central volcanoes in this cluster. Similarly, Vestmannaeyjar is apparently too far from the other central volcanoes in this cluster to form mechanical links with them (Fig. 4).

The numerical models indicate, and the available data supports, that central volcanoes and mature volcanic systems predominantly interact mechanically with volcanoes that are close to them. Immature or young volcanic systems that lack fully developed central volcanoes and are unfavourably located interact less, or (depending on the distance to nearby volcanoes) not at all, for the main reason that they do not yet possess the necessary shallow magma chamber. Thus, the great distance from other volcanic systems (and their central volcanoes) is presumably the main reason for the lack of mechanical interaction between Vestmannaeyjar and other volcanoes in the cluster. By contrast, the unfavourable location of Katla in Myrdalsjökull may be the main reason for its lack of mechanical interaction with the nearby central volcanoes.

5. Mechanical effects of Torfajökull

It has been mentioned here that the space between the two central volcano clusters of the EVZ is primarily occupied by large volcanic fissures (and normal faults) (Figs. 1, 7). The volcanic fissures run in a NE-SW direction, the same as most shared dykes and dyke swarms (Fig. 6). However, many of the fissures, and most of the normal faults, extend only southwest to the location of the Torfajökull volcano; those few that extend further to the southwest, such as Eldgja, tend to change direction at about the latitude of Torfajökull (Fig. 7). While there may thus be some general tectonic changes within the EVZ at this latitude, it appears that Torfajökull itself has large mechanical effects on the volcanic fissures (and associated dykes). To test these effects, we made several numerical models with Torfajökull acting as a soft elastic inclusion.

It is known that if an elastic material contains an inclusion of a different material and subject to a loading then the inclusion may develop a local stress that differs from that of the surrounding material. The stress within the inclusion will normally be higher or lower than that in the surrounding material if the inclusion material is stiffer (has a higher Young's modulus) or softer (has a lower Young's modulus) than the surrounding material, respectively (Savin, 1961; Nemat-Nasser and Hori, 1999; Jaeger et al., 2007). If a crack approaches an inclusion, the local stress field

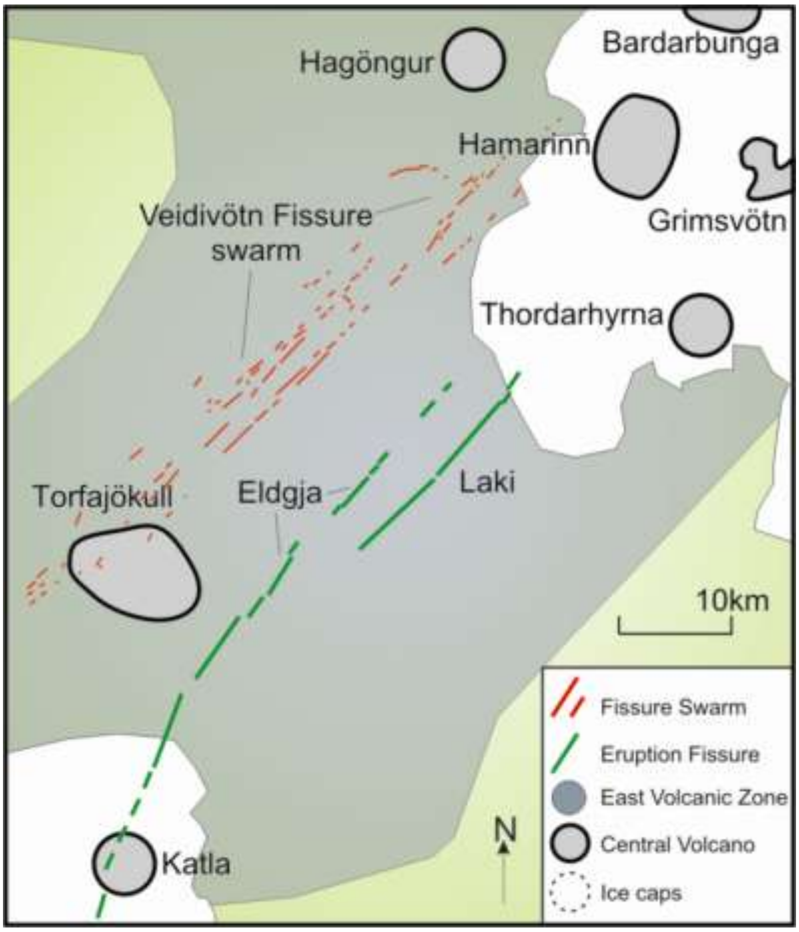

Fig. 7: Map showing the location of the large fissures and some central volcanoes from the East Volcanic Zone. The Veidivötn fissure swarm passes partly through the Torfajökull central volcano but with a very much reduced intensity. Map adapted from Johannesson and Saemundsson (1998).

associated with the inclusion (generated through shielding and amplification of the surrounding stress field) controls the behaviour and propagation path of the crack (e.g., Li and Chudnovsky, 1993). For example, the crack propagation rate may accelerate when approaching a soft inclusion, and decelerate when approaching a stiff inclusion. Further to this, the crack propagation velocity inside a soft inclusion may decrease because of the shielding, whilst it may increase inside a stiff inclusion because of stress amplification (Li and Chudnovsky, 1993).

Here we apply the theory of elastic inclusions and their effects on nearby cracks to the central volcano of Torfajökull, the associated volcanic fissures, and the EVZ. Torfajökull is located at the northern end of the southern cluster of central volcanoes (Figs. 2, 7) and is known to have experienced magma mixing with large volcanic fissures of the Veidivötn swarm, a part of the Bardarbunga volcanic system (Fig. 7; McGarvie, 1984; Mork, 1984; Larsen, 1984). Torfajökull has produced more rhyolitic rocks and obsidian lava flows than any other active volcano in Iceland. It holds the largest surface volume of rhyolite within the Neovolcanic Zone, and also contains the largest active collapse caldera in Iceland, with a maximum diameter of about 18 km.

We modelled Torfajökull as a soft, elastic inclusion with a Young's modulus of 10GPa (a value appropriate for surface rhyolite), with a thin ring of still softer material, namely with a Young's modulus of 0.1GPa, corresponding to the ring fault of the caldera (Fig. 8). The surrounding host rock has a Young's modulus of 20GPa. Two series of models were run with the

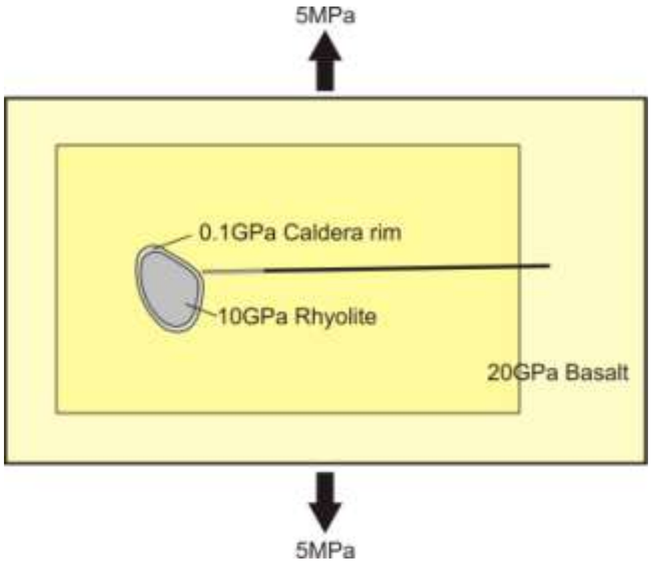

Fig. 8: Model outline showing Torfajökull and two stages in the lateral propagation of the main volcanic fissure of Veidivötn (shown in black and grey) towards the volcano. The ring fault forming the caldera rim has a Young's modulus of 0.1 GPa, the rhyolite inside the volcano 10 GPa, and the basaltic lava pile outside the volcano 20GPa, these values being typical for these rocks in Iceland (Gudmundsson, 2006). Note the direction of the tension applied to the models demonstrated by arrows. This tensile stress is parallel with the spreading vector and has a magnitude of 5MPa. The volcanic fissure (feeder dyke) in the model is loaded with 10MPa internal magma overpressure. The inset box shows the area shown in the model results (Fig. 9).

main Veidivötn volcanic fissure both approaching and almost touching the Torfajökull caldera (Fig. 8). All measurements and alignments of both the caldera and fissure were taken directly from a geological map of the area (Johannesson and Saemundsson, 1998).

Different Young's moduli were tested for the volcano inclusion, using values appropriate for hyaloclastite in addition to those suitable for rhyolite (both rock types occur within the volcano). Also, some models were run without the ring fault properties, to see how much effect these had on the local stress field. The volcanic fissure itself is modelled as a fluid-driven fracture subject to an internal magmatic overpressure of 10 MPa, as is reasonable for feeder-dykes close to or at the surface (Gudmundsson, 1995, 2006). In addition there is, during the rifting event associated with the propagation of the volcanic fissure, an applied 5 MPa regional tensile stress. This value is similar to the maximum in situ tensile strength of rocks in Iceland and is a reasonable maximum tensile stress to be expected during a rifting event (Gudmundsson, 2000, 2006).

The results of the numerical models clearly show that, as expected, Torfajökull acts as a crack arrester or "buffer" for the approaching fissure, suppressing the tensile stresses related to the spreading vector in this part of Iceland (Fig. 9). Since the volcano, and particularly the ring fault, acts as a soft inclusion, they suppress the regional tensile stress generated at the tip of the propagating fissure in such a way as to make it difficult for the fissure to enter and cross the Torfajökull volcano. This effect of Torfajökull on the main Veidivötn volcanic fissure is general in the sense that the volcano is expected to have similar effects on all volcanic fissures and normal faults that meet with it. If follows that Torfajökull may thus slow down the overall change of the EVZ into a proper rift zone. That is to say, by suppressing the spreading-related tensile stresses in this part of the EVZ, Torfajökull makes it more difficult for the part of the EVZ located southwest of the volcano to become a proper rift zone. The volcano and its caldera do not halt the propagation completely (Fig. 9B), but allow a minimal portion of the fissures to pass through. Some of the few NE-trending fissures southwest of Torfajökull may have propagated through the volcano, whereas in general they indicate that some spreading-related tensile stress

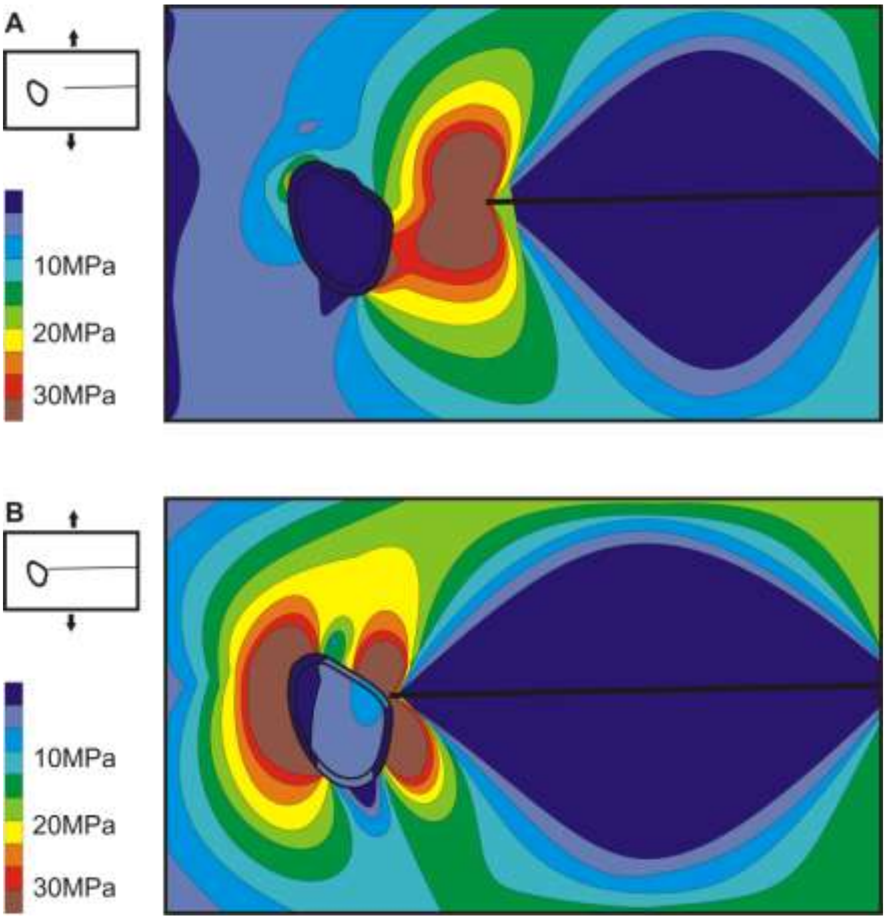

Fig. 9: Numerical model showing the tensile stress in mega-pascals around the volcanic fissure approaching (A) and almost coming into contact with (B) the caldera of Torfajökull. The properties of the models are given in Fig. 7.

concentrates in this part. Some of the fissures that enter the volcano from the northeast may be responsible for magma mixing (Fig. 7; McGravie, 1984; Mork, 1984; Larsen, 1984).

6. Mechanical effects of the Laki mountain

The effects of volcanoes in suppressing spreading-related tensile stresses and thus acting as buffers to the propagation of volcanic fissures can also be seen elsewhere within the EVZ, though on a smaller scale. Many Pleistocene hyaloclastite mountains occur in the EVZ and, as a rule, hyaloclastite has a comparatively low Young's modulus. When a hyaloclastite mountain is partly buried within much stiffer (mostly basaltic) lava flows, the mountain acts as a soft, elastic inclusion. One such mountain is the Laki mountain that significantly affected the propagation of the 1783 Laki volcanic fissure (Figs. 10-12).

Our field observations indicated that the 1783 Laki volcanic fissure and associated graben faults had difficulty in passing through the Laki mountain (Fig. 12), so we decided to run numerical models to test the mechanical effects of the Laki mountain (as a soft inclusion) on the Laki volcanic fissure. The model boundary conditions are based on our own field studies, as well as previous studies (Thordarson and Self, 1993) and geological maps and aerial photographs of the area. The Laki volcanic fissure was modelled in two progressions, approaching and almost touching the Laki hyaloclastite mountain (Fig. 10). The volcanic fissure has a magmatic overpressure of 10 MPa and is, during the rifting event, also subject to a regional tensile stress of 5 MPa. These are the same values as used in the previous numerical

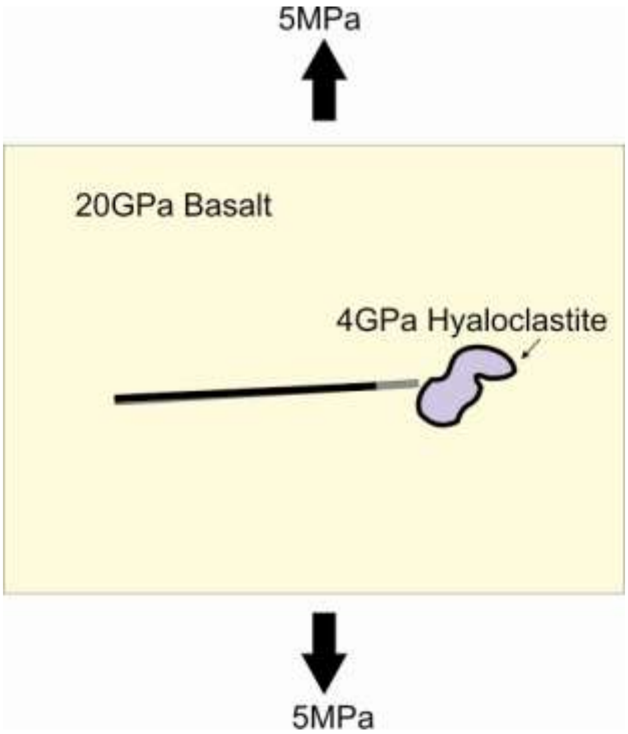

Fig. 10: Model outline showing the Laki mountain and two stages in the lateral propagation of the 1783 Laki volcanic fissures (shown in black and grey). The hyaloclastite Laki mountain has a Young's modulus of 4 GPa, whereas the basaltic lava pile hosting the volcano has a Young's modulus of 20 GPa (Gudmundsson, 2006). Note the direction of the tension applied to the models demonstrated by arrows; this tensile stress is parallel with the spreading vector and has a magnitude of 5MPa. The volcanic fissure in the model is loaded with 10MPa internal magmatic overpressure.

model (Fig. 8) and are used for the same reason. In the models, the Laki mountain itself has a Young's modulus of 4GPa, which is a reasonable value for young hyaloclastite at the surface (Gudmundsson, 2006; Gudmundsson and Andrew, 2007). In the model, the soft, elastic inclusion has the same geometry as the Laki mountain itself (Fig. 10).

The results of the models show that the volcanic fissure is hindered but not stopped on entering the Laki mountain. This is clearly indicated by the magnitudes of the induced tensile stresses at the tip of the fissure close to the mountain (Fig. 11). Clearly, the Laki mountain suppresses the tensile stresses to such an extent that it would have been difficult for the volcanic fissure, propagating in a NE-SW direction, to penetrate the mountain. Our field studies support this conclusion for they show that the volcanic fissure only goes partly into the mountain, along its northwestern slope, but does not reach its top. In fact, the main segment of the volcanic fissure close to the mountain ends clearly after propagating for a short distance into the mountain (Fig. 12). More specifically, detailed studies on the 1783 Laki eruption show that the different segments of the volcanic fissure opened in a succession from southwest to northeast towards the Vatnajökull ice cap (e.g. Thordarson and Self, 1993). As the fissure segments approach the Laki mountain, evidence from the field and aerial photographs show very clearly how their propagation is halted as they enter the mountain (Fig. 12).

7. Discussion and conclusions

There is clear clustering of the central volcanoes at the northern and southern ends of the EVZ (Figs. 1, 2). The cluster at the northern end can be explained, partly at least, as a concentration of highly active volcanoes over the mantle plume (e.g. Larsen et al, 1998; Gudmundsson and Högnadottir, 2007). The cluster in the southern part may be partly related to the changes in stress regime at this location, as

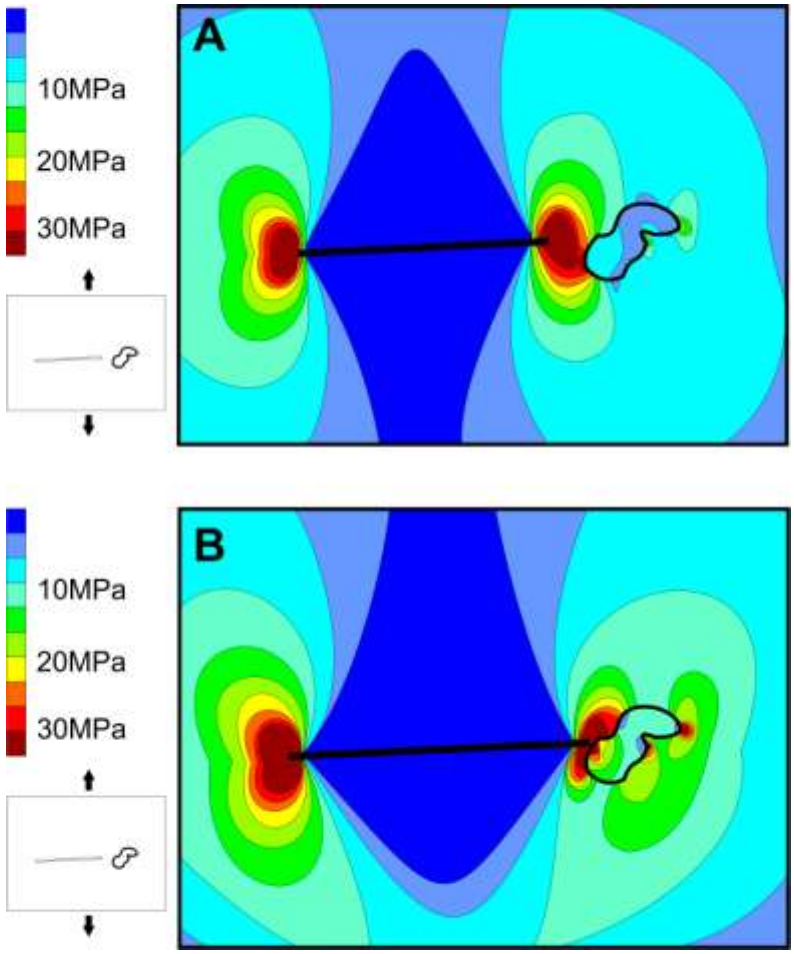

Fig. 11: Numerical model of the tensile stresses, in mega-pascals, around the 1783 Laki fissure approaching (A) and almost coming into contact with (B) the Laki mountain. The properties and configuration of the model are given in Fig. 10.

the rift zone propagates to the southwest through an older crust.

There is considerable observational evidence for mechanical interaction between some of the volcanoes within each cluster, but no evidence for interaction between volcanoes in different clusters. The evidence for mechanical interaction between volcanoes in the northern cluster includes the 1996 Gjalp eruption and earlier eruptions in that region (Fig. 5), dense dyke swarms connecting volcanoes (Fig. 6), and other data summarised by Gudmundsson and Andrew (2007). There is, at the moment, not so clear observational evidence for a mechanical interaction between the volcanoes in the southern cluster, but our numerical models indicate that such an interaction is likely between some of the volcanoes in that cluster (Fig. 4).

The results of the numerical models indicate that the local stress fields surrounding each of the volcanoes, modelled as elastic inclusions, have a very limited sphere of influence and, therefore, interaction. Not only does this apply to distance in that volcanoes that are comparatively far apart do not develop stress-concentration zones between them, but also to location. This latter point is illustrated by the volcano pairs Bardarbunga-Kverkfjöll in the northern cluster and Eyjafjallajökull-Katla in the southern cluster (Figs. 1-4). Both these pairs develop tensile stress-concentration zones in the numerical models (Figs. 3, 4A), so that mechanical interaction should be possible. However, the stress trajectories of σ_1 are clearly unfavourable to the sharing of dykes or normal faults (and thus seismic swarms). Thus, for mechanical interaction that involves dyke emplacement and faulting, it is not enough to have zones of high stress concentration between

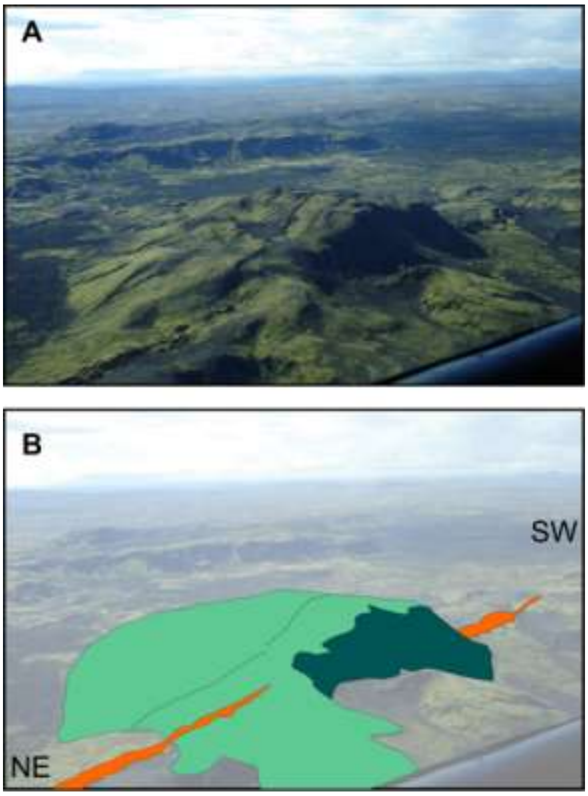

Fig. 12: A: View southeast, an aerial photograph of the Laki mountain. Segments of the 1783 Laki fissure, shown in orange, can be seen entering the mountain, where they become arrested. Photograph by R. Andrew.

nearby volcanoes; the orientation of the principal stresses must also be suitable.

The numerical models presented in the paper illustrate two opposite effects that volcanoes, modelled as inclusions, can have on the propagation of volcanic fissures and the associated feeder dykes. One effect is illustrated by the tensile stress-concentration zones between some of the nearby volcanoes within the same cluster, primarily those with fluid (mainly basaltic magma) shallow crustal magma chambers, as is discussed above. When the trajectories of σ_1 within the stress-concentration zones are favourable, that is, primarily trending northeast, then the propagation and sharing of dykes and seismogenic normal faults between nearby volcanoes in a cluster are encouraged. The other effect, however, is suppression of tensile stress close to, and inside, volcanoes that are primarily composed of soft (low Young's modulus) materials and that do not have a major active (fluid) shallow magma chamber. This effect leads to "buffering", that is, the arrest or slowing down of the propagation of normal faults and volcanic fissures.

The buffering effect is illustrated by the central volcano Torfajökull and the hyaloclastite Laki mountain. The Laki mountain is likely to have formed in a single or a few eruptions, like most hyaloclastite mountains do, and unlikely ever to have developed a shallow magma chamber. Torfajökull, by contrast, is a large central volcano with a collapse caldera that must have had a large shallow crustal magma chamber at some stages during its development. Presently, and presumably for some considerable time, however, there is little or no evidence for a major active (fluid) magma chamber beneath Torfajökull. Several seismic experiments carried out in recent years have failed to detect any large fluid body beneath the volcano (Soosalu and Einarsson, 2004). Thus, for a considerable time there has presumably been no major fluid magma chamber active beneath the volcano, meaning that the one that once existed is now partly or entirely solidified. The volcano and its ring fault are thus modelled as a soft inclusion.

Because of the low Young's modulus of both the Laki mountain and the eruptive rocks and the ring fault of Torfajökull, and the partly or entirely solidified magma bodies beneath the latter volcano, tensile stresses associated with rifting events and magma-driven volcanic fissures and dykes tend to be suppressed on entering either of these volcanoes (Figs. 8-12). For the Laki mountain, this arrest is of a very local nature, whereas for Torfajökull affects a considerable part of the EVZ. We argue here that the buffering effects of Torfajökull may have contributed to the slowing down the rate of the change of the part of the EVZ part southwest of Torfajökull into a proper rift zone.

In conclusion, the present models explain why some volcanoes in the clusters may act in harmony and encourage shared dykes and faults. They also explain the tendency of other volcanoes to arrest the lateral propagation of faults, dykes, and volcanic fissures and, thereby, in the case of a large volcano such as Torfajökull, slow down the rate of change of a part of the EVZ into a proper rift zone. The models do not, however, explain the occurrence of the large volcanic fissures between the two volcano clusters of the EVZ. The presence of the large fissures could be partly due to the regional stress in the area, and the relative youth of the EVZ. Evidence of large eruptions, with voluminous lava flows, can be seen in the Tertiary lava pile of Iceland. The presence of these large volcanic fissures could be part of the natural progression of the EVZ towards maturity and in becoming the main rift zone of Iceland. The mechanical reasons for the large fissures of the EVZ need to be explored and explained. We believe, however, that the effect of Torfajökull as a soft inclusion goes some way to explain the different volcanotectonic activity within the EVZ.

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14 Appendix IV

Effects of dyke emplacement and plate pull on mechanical interaction between volcanic systems and central volcanoes in Iceland

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Abstract.

The surface expressions of most Holocene rift-zone volcanic systems in Iceland are 40-150-km-long and 5-20-km-wide swarms of tension fractures, normal faults and basalt volcanoes each of which extends from a central volcano (a composite volcano or a caldera). Below the Holocene surface, the swarms are mainly composed of subvertical dykes and normal faults except within the central volcanoes where the main tectonic elements are inclined sheets. The inclined sheets are mostly 0.5 m thick, whereas the regional dykes are commonly 3-6 m thick and occasionally as thick as 50-60 m. The two principal ways by which rift-zone volcanic systems become loaded are (1) the magmatic overpressure induced by dykes, and (2) the plate "pull" associated with extension in the direction of the spreading vector (105°). This paper shows that both loading conditions give rise to mechanical interaction between volcanic systems in general, and their central volcanoes in particular.

Here we show that the magmatic overpressure of a regional dyke may reach tens of mega-pascals. We model the effects of simultaneous dyke injections, each dyke with an overpressure of 10 MPa, in the echelon systems on the Reykjanes Peninsula. The results indicate N-trending zones of high shear stress between the nearby ends of the volcanic systems, favouring strike-slip faulting. Geometrically similar shear-stress zones develop between the volcanic systems on the peninsula when acted on by a plate pull of 5 MPa in a direction parallel with the spreading vector. The results agree with the observation that there are many N-trending strike-slip faults on the Reykjanes Peninsula. When the same plate pull is applied to a cluster of 8 central volcanoes in Central Iceland, zones of high tensile stress develop between many of the volcanoes. These highly stresses zones encourage mechanical interaction between the volcanoes, such as simultaneous dyke emplacement and seismogenic faulting, as is supported by observational data.

Keywords: volcanic system, dyke emplacement, central volcano, mechanical interaction

1. Introduction

The zone of active volcanism in Iceland is covered by rocks of the Brunhes magnetic epoch (< 0.8 Ma) and is referred to as the Neovolcanic

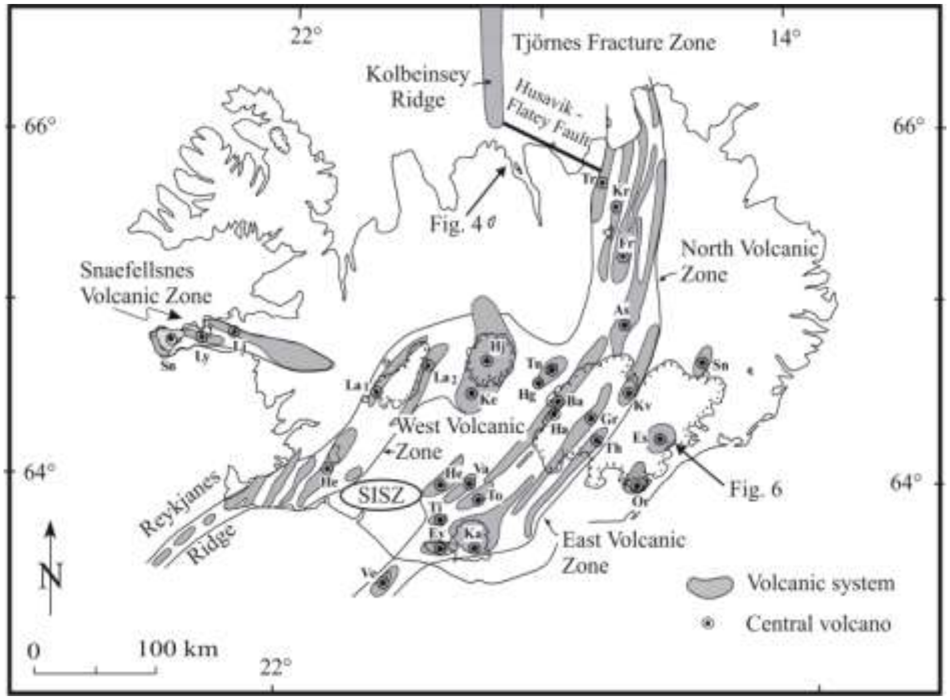

Fig. 1. General volcanotectonic map of Iceland showing the Neovolcanic Zone, the ocean ridge discontinuities, and the main Holocene volcanic systems and central volcanoes. The part of the Neovolcanic Zones referred to as the rift zone comprises the North Volcanic Zone, the West Volcanic Zone, and the East Volcanic Zone to the south tips of the volcanic systems of Bardarbunga and Grimsvötn. The two main ocean-ridge discontinuities are the Husavik-Flatey Fault of the Tjörnes Fracture Zone and the South Iceland Seismic Zone (SISZ).The central volcanoes and associated volcanic systems indicated are as follows: Tr = Theystareykir, Kr = Krafla, Fr = Fremri-Namur, As = Askja, Kv = Kverkfjöll, Th = Thordarhyrna, Gr = Grimsvötn, Ha = Hamarinn, Ba = Bardarbunga, Tu = Tungnafellsjökull, Hg = Hagöngur, Ka = Katla, Ey = Eyjafjallajökull, Ve = Vestmannaeyjar, Ti = Tindfjallajökull, Va = Vatnafjöll, To= Torfajökull, He = Hekla, Hj = Hofsjökull, Ke = Kerlingafjöll, La1&2 = Langjökull, He = Hengill. The other systems on the Reykjanes Peninsula (from east to west) are Brennisteinsfjöll, Trölladyngja and Reykjanes. Off-coast are Eldey, Geirfuglasker and Eldeyjarbodi. In the Snaefellsnes Volcanic Zone are the systems Sn = Snaefellsjökull, Ly = Lysuskard, and Lj = Ljosufjöll. Outside the East Volcanic Zone are the systems Or = Oraefajökull, Es = Esjufjöll, and Sn = Snaefell. Data from Jakobsson (1978a, b), Jakobsson et al. (1978), Saemundsson (1978), Johannesson and Saemundsson (1998), and other sources.

Zone (Fig. 1). The Neovolcanic Zone is composed of three main segments or subzones. These are the North Volcanic Zone (NVZ), the East Volcanic Zone (EVZ), and the West Volcanic Zone (WVZ). Some authors divide the WVZ further into the WVZ proper and the Reykjanes Peninsula (RP). The NVZ, the EVZ, and the WVZ form the divergent plate boundary, that is, they constitute the rift zone in Iceland, with the exception that the part of the EVZ that lies to the southwest of the central volcano Torfajökull is a propagating rift (Fig. 1).

The main segments or subzones are composed of volcanic systems (Figs. 1, 2). In the rift zone, these are essentially giant swarms of fractures and basalt volcanoes such as crater rows (volcanic fissures) at the surface, but dykes and some normal faults at deeper crustal levels (Gudmundsson, 2000a; Thordarson and Larsen, 2007). Most systems are 40-150 km long, 5-20 km wide, and develop a central volcano. The central volcanoes are either stratovolcanoes forming topographic highs or, more commonly, collapse calderas. The volcanic systems outside the rift zone are somewhat different in that they

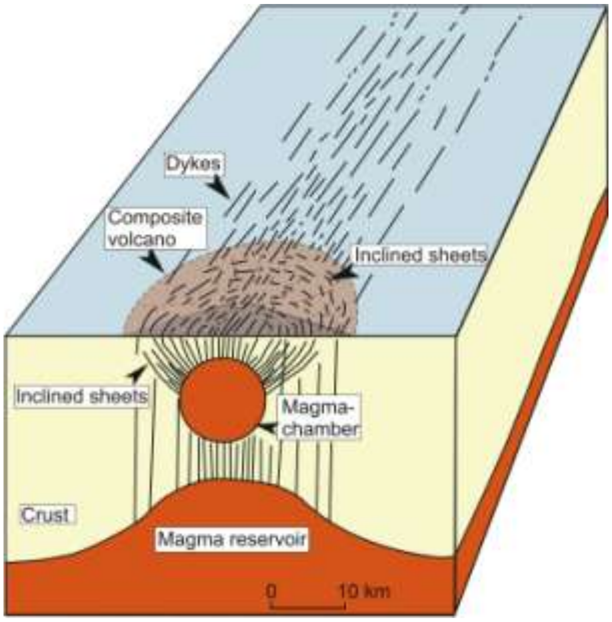

Fig. 2. Schematic internal structure of (half) a rift-zone volcanic system in Iceland. An elongate magma reservoir underlies the system in the lower crust or at the crust-mantle boundary. The reservoir supplies magma directly to many shield volcanoes and fissure eruptions in the parts of the volcanic system outside the associated central volcano, as well as to the shallow magma chamber beneath the central volcano. In inactive and partly eroded volcanic systems, the central volcano (here half of the former volcano, shown in brown, is outlined) is characterised by a dense swarm of thin inclined sheets (Fig. 6) and, when deeply eroded, plutons (uppermost part of an extinct magma chamber) in its core. The parts outside the composite volcano are characterised by a swarm of subvertical, thick regional dykes (Figs. 3-5).

lack well-defined fissure swarms (Fig. 2). In this paper the focus is on rift-zone volcanic systems.

Most volcanic eruptions in Iceland are supplied with magma through magma-driven fractures. When the magma in a fracture solidifies, the fracture is, depending on its attitude and relation to the host rock, referred to as an inclined sheet, a dyke, or a sill. Sills, however, rarely feed eruptions but may be important in initiation and development of shallow magma chambers (Gudmundsson, 1990). Thus, most sheets that supply magma to eruptions are either inclined, hence the term "inclined sheet", or close to vertical and thus proper dykes. In this paper, however, we use the word "dyke" in a generic sense, unless specified otherwise, including fluid and solidified inclined sheets and proper dykes.

There are two principal ways by which the rift-zone volcanic systems are loaded so as to give rise to stress interaction between them. One is plate pull, that is, the tensile stress pulling the plates apart and generating extension across the divergent plate boundary. While the tensile stress may be absolute (the minimum principal compressive stress σ_3 being negative) close to the surface, at depths of a kilometre or more the tensile stress is relative, that is, reduction in compressive stress (Gudmundsson, 2006). Close to the surface the plate-pull generated absolute tensile stress cannot exceed the in situ tensile strength, so that the magnitude of this loading is normally less than about 5-6 MPa (Haimson and Rummel, 1982; Schultz, 1995; Amadei and Stephansson, 1997). In Iceland the average spreading rate is about 1.8 cm/year in the direction of the spreading vector, that is, N105°E (DeMetz et al., 1990, 1994).

The other principal loading source is the magmatic overpressure of the dykes that become emplaced within the volcanic systems. A magma reservoir becomes ruptured and injects a dyke when the total magma pressure in the reservoir becomes equal to the minimum principal compressive stress σ_3 plus the tensile strength of the host rock (Jaeger and Cook, 1979; Gudmundsson, 2006). If, in addition, the magma is less dense than the host rock, as is commonly the case for dykes injected from a deep-seated reservoir (Fig. 2), buoyancy increases the overpressure in the dyke on its path towards the surface. At its maximum, the magma overpressure in the dyke may reach tens of mega-pascals (Delaney and Pollard, 1981; Pollard and Segall, 1987; Ray et al., 2007; this paper) and gives rise to compressive stress in the surrounding host rock that modifies the stress fields in and around the nearby volcanic systems.

This paper has two principal aims. The first is to present some general results on the

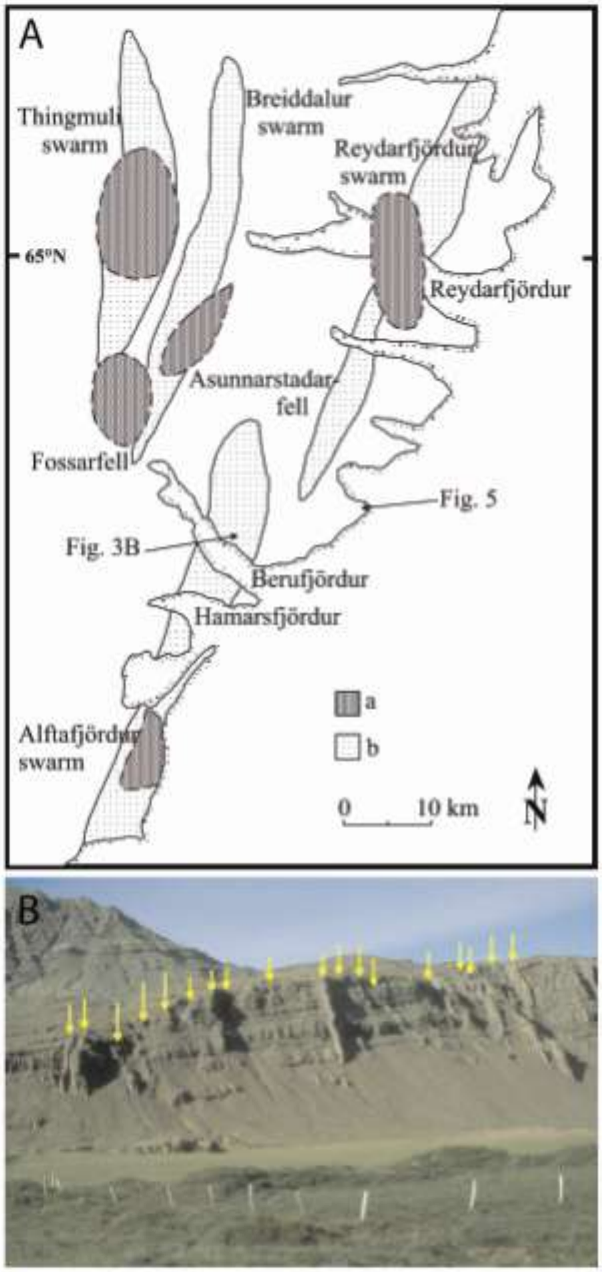

Fig. 3. A) Central volcanoes and local sheet swarms (a) and associated regional dyke swarms (b) in East Iceland. Data from Walker (1959, 1960, 1963, 1974) and Gudmundsson (1995). B) View north, regional dykes from the Alftafjördur Dyke Swarm as seen in a cliff section on the north shore of the fjord Berufjördur (located in Fig. 3A). The dykes in this part of the swarm identified with arrows are mostly 5-10 m thick whereas the arithmetic average thickness of dykes in this part of the swarm is 5.5 m. The basaltic lava pile dissected by the dykes dips 6-8°W.

geometries and mechanics of emplacement of dykes in the rift-zone volcanic systems of Iceland. Systematic studies of dyke swarms in Iceland were initiated by Walker (1959, 1960, 1963, 1965, 1974), so that this aspect of the paper is a fitting tribute to Walker's outstanding work. The focus is on dyke emplacement and overpressure, and how the overpressure temporarily alters the local stress field around the emplaced dyke.

The second aim is to analyse the stress interaction between rift-zone volcanic systems. The two main loading conditions considered are compressive stresses generated by overpressured dykes and (absolute and relative) tensile stresses generated by plate pull. The focus is on how stress interaction between volcanic systems and central volcanoes may contribute to two tectonic processes, namely (1) dyke sharing between central volcanoes and (2) seismogenic faulting in stress-concentration zones connecting nearby volcanic systems.

2. Volcanic systems

The structure, general characteristics, and historical (past 1100 years) eruption frequencies of the volcanic systems of Iceland have recently been discussed in detail (Thordarson and Larsen, 2007). Consequently, here we present only a brief summary of their main characteristics, focusing on the aspects most relevant for stress interactions.

The systems are defined partly through petrological data (Jakobsson 1979a,b; Jakobsson et al., 1978) and partly through volcanotectonic data (Saemundsson 1978, 1979). Using these data, some 30 Holocene volcanic systems have been defined within the volcanic zones (Fig. 1). Of these 30 systems, some 20 have a fissure (and a dyke) swarm and belong to the rift zone. Of these 20, 12 have a clear central volcano. Another 8 systems have what may be regarded as a "central domain" where the eruption frequency is highest (often with a high-temperature geothermal field) but not a clear-cut central volcano (Thordarson and Larsen, 2007). The remaining 10 volcanic systems are in the off-rift zones, have no fissure swarms (but central volcanoes), and are thus largely beyond the scope of the present paper.

Defining a volcanic system as active if it has erupted during the Holocene, the exact number and boundaries of active volcanic systems are not quite clear (Fig. 1). This is partly because the Holocene is a very short time when considering that the systems are commonly active for 0.5-1 Ma. Thus, some active systems may, by chance, have erupted only once or twice during the Holocene, in which case it is not always clear if those eruption sites constitute a part of a separate system or are just the extension of a nearby system. How the geometric outlines of the systems are drawn thus depends on which fissures are thought to belong to which volcanic systems. Several different geometries of the volcanic systems have been proposed; here we follow mainly Johannesson and Saemundsson (1998) and Gudmundsson (2000a), which are also generally very similar to the geometries shown by Thordarson and Larsen (2007).

The Holocene systems can only be studied at the surface. However, analogous, extinct systems occur in the deeply eroded Tertiary and Pleistocene lava pile. Some 15 such systems have been mapped and as many as 55 identified (Walker 1960, 1963, 1965, 1974, Saemundsson 1979). The extinct systems are very helpful for understanding the likely three-dimensional structure of the active systems (Fig. 2). The extinct rift-zone volcanic systems are characterised by dyke swarms (Fig. 3), such as were initially mapped by Walker (1959, 1960, 1963, 1974). Subsequently, many more dyke swarms have been mapped, and some of those mapped by Walker, remapped (Gudmundsson, 1995; Klausen, 2004, 2006).

3. Dyke swarms

In volcanic systems with central volcanoes, there are two main subswarms of dykes: local sheet swarms and regional dyke swarms (Figs. 2, 3A). The regional dykes are mostly subvertical and subparallel; the mode thickness is 1-2 m and the arithmetic average thickness 4-6 m in the Tertiary swarms and 1-2 m in the Pleistocene swarms (Fig. 3B). However, some dykes in the regional swarms are as thick as 50-60 m (Fig. 4), whereas others are just a few

Fig. 4. View north, a 54-m-thick multiple, basaltic dyke in North Iceland (located in Fig. 1). The multiple dyke is composed of at least three separate dykes. The general attitude of the dyke is N16°E, 80°E, the thick, vertical arrows indicating its outer margins.

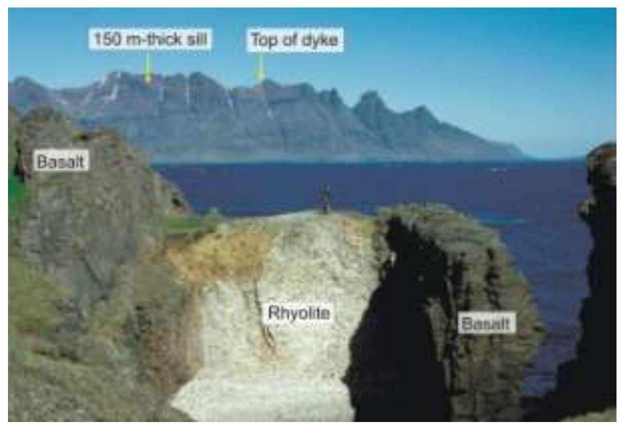

Fig. 5. View northeast, the composite dyke at Streitishvarf in East Iceland (located in Fig. 3A). The outer parts of the dyke are of basalt, the central part of rhyolite. The total thickness of the dyke is 26 m, the rhyolite part being 13 m, the western basalt part 7.5 m and the eastern part 5 m. The dyke can be traced for about 14 km, its strike changing from about N20°E to about N14°E towards its northern end. Similarly, the dyke dip changes along its path, from about 74°E to vertical.

Fig. 6. View northwest, part of a swarm of inclined sheets in Southeast Iceland (located in Fig. 1). The entire cliff is composed of (mostly basaltic) sheets dipping towards a shallow crustal chamber (not exposed). The person standing close to the lower, central margin of the photograph provides a scale.

centimetres. Most dykes are basaltic, but some are acid or composite (Fig. 5). Generally, it is difficult to trace the dyke over a long distance, but some dykes are known to reach lengths of at least 10-20 km (Fig. 5; Gudmundsson, 1995).

The inclined sheets occur in swarms normally confined to central volcanoes (Gudmundsson, 1995; Klausen, 2004, 2006). The swarms are normally very dense (Fig. 6), particularly close to the extinct shallow magma chambers (exposed as mafic and/or felsic plutons) where 80-100% of the rock may consist of inclined sheets. The swarms are normally circular or slightly elliptical in plan view, several kilometres in radius, and contain anywhere between several thousand to tens of thousands of sheets. Although the sheet thickness ranges up to 14 m, most are 0.5 m thick or less, and thus much thinner than the regional dykes.

The number of dykes and sheets decreases rapidly with elevation in the crust, meaning that most dykes and sheets never reached the surface to feed eruptions but rather became arrested (Gudmundsson, 2006). The dykes become arrested in various ways, but perhaps the most common is to see them arrested at layer contacts. In particular, when the contacts are between rock layers of contrasting mechanical properties, dykes tend to become either offset across the contact or arrested.

4. Dyke emplacement and overpressure

A magma chamber or a reservoir ruptures and injects a dyke when the following condition is satisfied (Gudmundsson, 1990, 2006):

$$p_l + p_e = \sigma_3 + T_0 \tag{1}$$

where p_l is the lithostatic stress (or pressure) at the depth of the chamber. The other symbols are defined as follows: $p_e = P_t - p_l$ is the excess magmatic pressure, that is, the difference between the total magma pressure, P_t, in the chamber at the time of its rupture and the lithostatic stress or pressure; σ_3, as defined above, is the minimum principal stress (compressive stress regarded as positive); and T_0 is the in situ tensile strength in the roof of the chamber.

The condition of dyke injection as presented by Eq. (1) refers to the local σ_3 and T_0 of the host rock of the chamber. This means that dyke injection occurs when the condition of Eq. (1) is reached at any point along the walls of the chamber, irrespective of the shape of the chamber or its depth below the surface. Thus, stress concentration effects due to the shape of the magma chamber, as well as to irregularities at its boundary, are included in the local magnitude of σ_3. This condition can be reached either by increasing p_e, decreasing σ_3, or both (Gudmundsson, 2006). Since magma chambers are long-lived fluid-filled structures which normally rupture infrequently in comparison with their lifetime, it may be assumed that a chamber is in lithostatic equilibrium with its host rock except during unrest periods. Denoting the maximum compressive principal stress by σ_1, it follows that for most of the time along the contact between the host rock and a fluid chamber, in Eq. (1)

$p_l = \sigma_3 (= \sigma_1)$ and $p_e = 0$. It is only during short-term unrest periods that $p_e > 0$, either because of added volume of magma (absolute increase in fluid pressure) or reduction in σ_3, that the condition of Eq. (1) may be satisfied, resulting in rupture and dyke injection.

When the injected dyke and its source chamber are hosted by rocks that behave as elastic and the magma flow is vertical, the vertical co-ordinate z being positive upwards, then the overpressure available to drive the propagation of the dyke P_0 is given by (Gudmundsson, 1990):

$$P_0 = -(\rho_r - \rho_m) g z + p_e \qquad (2)$$

where ρ_r is the density of the host rock. The other symbols are the density of the magma ρ_m, the acceleration due to gravity g, and the excess pressure p_e, defined above. A minus sign is used since g is positive downwards but the co-ordinate z is positive upwards. The overpressure P_0 is also commonly referred to as driving pressure or net pressure; here overpressure will be used.

Equation (2) applies to the case where the state of stress in the host rock is isotropic, so that all the principal stresses are equal. When, however, the state of stress is anisotropic, so that the principal stresses have different magnitudes (as is common during unrest periods), then, for the two-dimensional case examined here, the term $\sigma_d = \sigma_1 - \sigma_3$ must be added to Eq. (2) to yield:

$$P_0 = -(\rho_r - \rho_m) g z + p_e + \sigma_d \qquad (3)$$

where all the symbols are as defined above.

One way to estimate the overpressure in a dyke is to use Eqs. (2, 3). Then the overpressure in a hypothetical dyke is calculated using information about magma and host rock density, depth to the magma chamber, excess pressure in the chamber at the time of rupture, and the stress difference in the crust. As an example, consider the feeder dyke to the Laki 1783 eruption (Thordarson and Self, 1993; Thordarson and Larsen, 2007). The Laki eruption is most likely to have originated from a deep-seated magma reservoir which, according to current ideas, would be in the lower crust or at the boundary between the crust and the upper mantle. Thus, the depth could be 15-20 km (Fig. 2; Gudmundsson, 2006). Here we use a 15-km-thick crust with an average density of 2900 kg m^{-3} (Gudmundsson, 2006) and take the average density of the tholeiitic magma similar to that which erupted in the Laki eruption as 2650 kg m^{-3} (Williams and McBirney 1979). The magmatic excess pressure in the reservoir is regarded as equal to a typical in situ tensile strength, or 3 MPa (Haimson and Rummel, 1982; Schultz, 1995). Using these values, Eq. (2) yields the magmatic overpressure of the Laki feeder dyke close to the surface as about 39 MPa. Close to the surface, the term $\sigma_d = \sigma_1 - \sigma_3$ in Eq. (3) is positive and would thus add to the overpressure of the feeder dyke, by at least several mega-pascals.

Another method for estimating the magmatic overpressure in a dyke is to use the aspect (length/thickness) ratio of the dyke modelled as a two-dimensional through-the-thickness crack subjected to internal fluid pressure (Sneddon and Lowengrub, 1969; Delaney and Pollard, 1981; Pollard and Segall, 1987; Gudmundsson, 2000b; Ray et al., 2007). If the strike dimension L of the dyke is less than its dip dimension, and its maximum thickness is b_{max} (Figs. 4, 5), then the magmatic overpressure P_0 can be estimated crudely from the following equation (Gudmundsson, 2000b):

$$P_0 = \frac{b_{\max} E}{2L(1-\nu^2)} \qquad (4)$$

Thus, the aspect (L/b_{max}) ratio is a crude measure of the magmatic overpressure at the time of dyke emplacement, provided the appropriate elastic moduli of the host rock are known. Aspect ratios of exposed dykes are normally not well constrained because even if the thickness is reasonably well known from direct measurements, the length or strike dimension is less well known (Delaney and Pollard, 1981; Pollard and Segall, 1987; Gudmundsson, 1995; Ray et al., 2007). This follows, first, because the lateral ends of a dyke are normally uncertain and, second, because most dykes are discontinuous in lateral sections so that it is often unclear which segments belong to a single dyke during a single eruption. Even for feeder dykes, the associated volcanic fissures are normally discontinuous and commonly the segments are not all active at the same time (Thordarson and Self, 1993; Thordarson and Larsen, 2007).

In a study of regional Tertiary basaltic dykes, 3-22 km long and exposed at 1-1.5 km depth below the original surface of the lava pile, Gudmundsson (2000a) determined the aspect ratios of 16 dykes as being from 300 to 1500. For a Young's modulus of 20 GPa and Poisson's ratio of 0.25, as are appropriate for this crustal depth (Gudmundsson, 2006), Eq. (4) gives, for an aspect ratio of 1500, an overpressure of 7.4 MPa, and for an aspect ratio of 300, an overpressure of 37 MPa. The latter value compares very well with the overpressure value obtained from Eqs. 2-3 for a typical basaltic dyke. The results, although crude, indicate that during regional basaltic dyke emplacement in a volcanic rift-zone system, the magmatic overpressure associated with the dyke may be from several mega-pascals to several tens of mega-pascals.

5. Stress transfer due to dyke emplacement

Field studies indicate that most injected dykes do not reach the surface to feed eruptions but rather become arrested in some crustal layers at depth (Gudmundsson and Philipp, 2006). This general conclusion is supported by seismological data such as from Hawaii where most episodes of accelerated seismic event rate, inferred to be associated with dyke propagation, do not result in eruption (Chastin and Main, 2003). Rough calculations indicate that in the rift-zone volcanic systems of Iceland, typically only about 10% of the dykes injected from a deep-seated magma reservoir would reach the surface (Gudmundsson, 2006). From the calculations above it follows, however, that all dykes injected into volcanic systems cause loading within those systems. More specifically, the magmatic overpressure associated with any injected dyke generates compressive stresses in a direction perpendicular to the dyke. Since the magmatic overpressure can range from several mega-pascals to several tens of mega-pascals, it is clear that the local stresses within a volcanic system following a dyke injection may for years or decades be largely controlled by the loading due to the dyke.

Such an effect was observed in the Krafla Volcanic System in North Iceland following a 9-year dyke emplacement episode associated with the Krafla Fires 1975-1984 (Björnsson, 1985; Einarsson, 1991; Saemundsson, 1991; Thordarson and Larsen, 2007). A feeder dyke was generated in at least 9 injections, reaching a maximum thickness of about 9 m (Tryggvason, 1984; Gudmundsson, 1995; Sigmundsson, 2006; Sturkell et al., 2006). The magmatic overpressure associated with the dyke resulting in compressive stresses and excess spreading rates out to a distance of tens of kilometres to either side of the Krafla Volcanic Systems (Foulger et al., 1992). Similarly, during a recent rifting episode in the East African Rift Valley, overpressured dykes are thought to be partly responsible for pushing the plates apart (Wright et al., 2006).

To test stress transfer and mechanical interaction between volcanic systems resulting from dyke injection, it is best to consider systems that form geometrical patterns where such effects would be maximised. The westernmost systems in the WVZ form such a pattern (Fig. 1). While the volcanic systems in this part of the rift zone are often regarded as four separate systems (Fig. 1), a more detailed analysis indicates that they may perhaps more accurately be regarded as five systems (Fig. 7). For the present analysis, the results are essentially the same whether the systems are regarded as four or five; in the numerical models we regard the systems as five.

To test the stress transfer and mechanical interaction between the volcanic systems on the Reykjanes Peninsula due to overpressured dykes being injected into the systems, we made many numerical models using the finite-element program Ansys (www.ansys.com; Logan, 2002). We model the part of the volcanic systems where the dykes are injected as two dimensional, elliptical holes, subject to magmatic overpressure of 10 MPa as the only loading. Conceptually, this does not necessarily mean that any single dyke extends along the entire length of a volcanic system. Rather, it means that a few dykes were injected into the system over such a short period of time, say tens or hundreds of years, that the region outside the dyke-injection zone is subject to pressure over the same period of time as if the pressure were generated by a single dyke. It is well known from fracture mechanics that if collinear cracks (in this case dykes) are closely spaced their mechanical effect is as if they were a single crack (Sneddon and Lowengrub, 1969). We model the dykes and the associated parts of the volcanic systems as through-the-thickness cracks (Eq. 4), in which case a two-dimensional approxiation is appropriate (Sneddon and Lowengrub, 1969; Gudmundsson, 2000b). To simplify the models, we ignore the mechanical layering (Gudmundsson and Philipp, 2006) and use a constant average static Young's modulus for the crust of 20 GPa and a constant Poisson's ratio of 0.25. These values are typical for the uppermost 3-4 km of the crust in the volcanic zones (Gudmundsson, 2006).

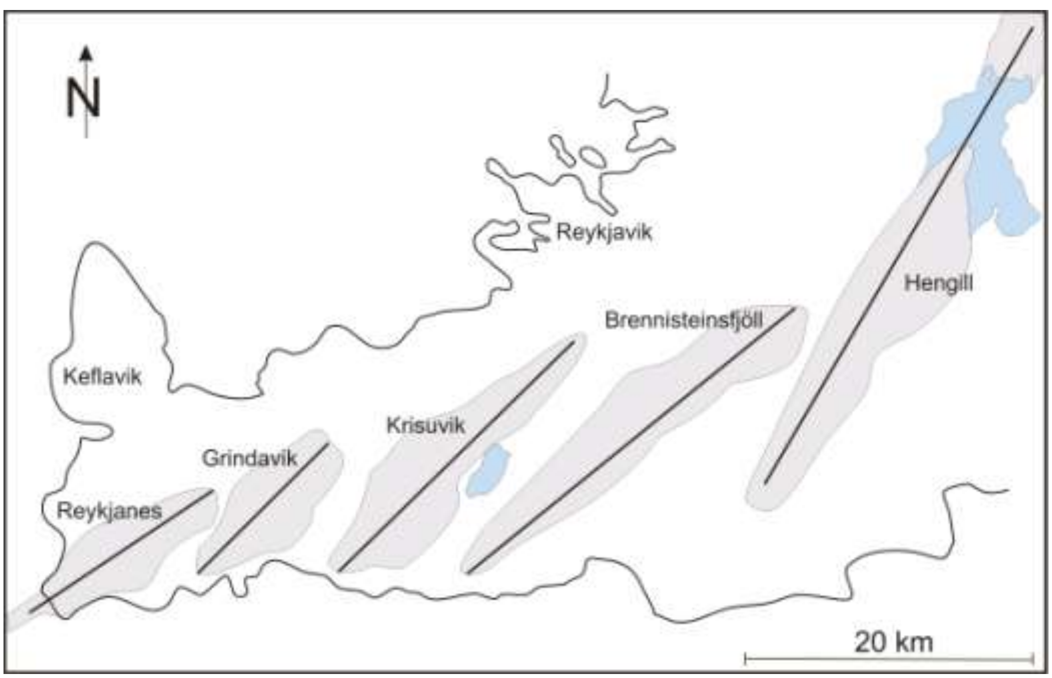

Fig. 7. Volcanic systems on the Reykjanes Peninsula (located in Fig. 1), here regarded as five, can be approximated as elliptical holes (or inclusions). This geometry is used as a basis for the numerical models presented in Figs. 8 and 9.

The typical results are presented in Fig. (8). They show, first of all, that dyke injections into two or more nearby volcanic systems generate zones of high shear stress between the nearby tips of the systems (Fig. 8B). Faults form or slip when the local shear stress satisfies the condition for failure as represented by the Navier-Coulomb criterion, the von Mises criterion, and other similar criteria (Jaeger and Cook, 1979). Here the zones are represented by the von Mises (octahedral) shear stress, the criterion being that yielding or failure occurs when the octahedral shear stresses reach a value characteristic of the material (Jaeger and Cook, 1979). We refer to the plotted stresses as von Mises shear stress or simply shear stress.

The shear-stress zones between the volcanic systems trend north-south and could be responsible for part of the seismogenic faulting on the Reykjanes Peninsula. As presented here, all the volcanic systems are supposed to have had recent dyke injections with 10 MPa overpressure. This scenario is perhaps not the most likely, although

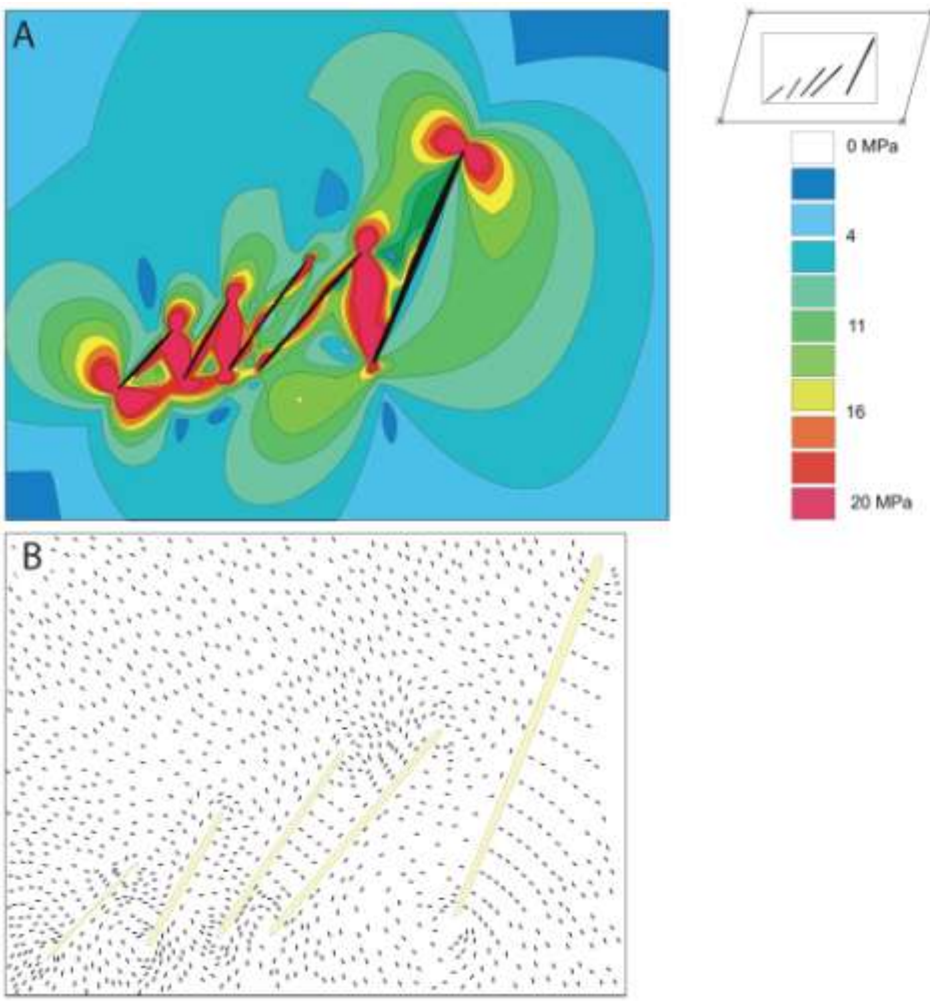

Fig. 8. Numerical model results showing the stress fields that develop around the volcanic systems on the Reykjanes Peninsula (located in Fig. 7) when loaded by simultaneously injected dykes, each dyke with an overpressure of 10 MPa. A) North-south trending zones of high von Mises shear stress, given in mega-pascals, develop between the nearby ends of the volcanic systems. B) Trends of the directions (trajectories) of the maximum principal compressive stress, σ_1.

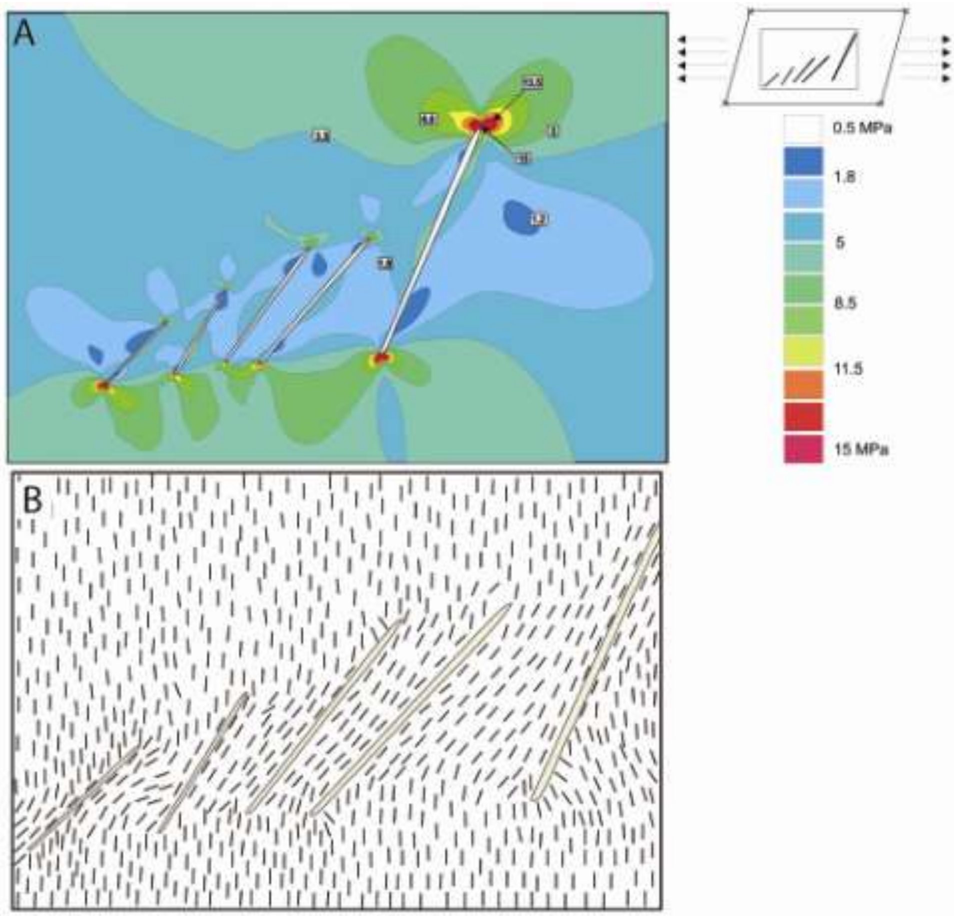

Fig. 9. Numerical model results showing the stress fields that develop around the volcanic systems on the Reykjanes Peninsula (located in Fig. 8) when subject to plate pull of 5 MPa in the direction of the spreading vector, 105°. A) Von Mises shear stress in mega-pascals. B) Trends of the directions (trajectories) of the maximum principal compressive stress, σ_1.

dyke injections into the crust are of course much more frequent than eruptions (Gudmundsson, 2006), so that even if there are no known eruptions in these volcanic systems for the past several hundred years, many or all the systems may have been subject to one or more dyke injections during this period. The main point is, however, that if any of the nearby systems would experience dyke injections within a time window of, say, tens of years shear stresses would develop between their nearby ends as indicated (Fig. 8A).

The stress trajectories (Fig. 8B) show that if faults would form in the area, some would tend to be north-trending, dextral strike slip, as are commonly observed (Clifton and Kattenhorn, 2006). Depending on changes in loading conditions with time, some northerly trending faults, however, could be sinistral. N-trending sinistral faults have not been reported from the Reykjanes Peninsula. It is known, however, from the South Iceland Seismic Zone that some N-trending strike-slip faults are sinistral while other faults show evidence or both sinistral and dextral movements (Gudmundsson, 2007). Although careful studies looking for N-trending Holocene sinistral faults have not so far been carried out in the Reykjanes Peninsula, the data available today indicate that they may be rare, and that N-trending dextral faults are more common (Clifton and Kattenhorn, 2006). To analyse further the mechanical interaction between the volcanic systems, and its potential for generating dextral and sinistral faults, we also explored a different type of loading, namely plate pull.

6. Stress transfer due to plate pull

The Reykjanes Peninsula

To test the stress transfer and mechanical interaction between the volcanic systems on the Reykjanes Peninsula as a result of plate pull, we again made numerical models using the Ansys program. We used the same crustal propertites, Young's modulus of 20 GPa and a Poisson's ratio of 0.25, as in the models above. The only loading used in the present models is plate pull, that is, tensile stress of 5 MPa in a direction parallel with the spreading vector, 105°.

The results (Fig. 9) show that while the von Mises shear stress concentration between the volcanic systems is not very high, the trend of the maximum principal compressive stress, σ_1 , is generally northeast-southwest and would favour the formation or reactivation of north-trending dextral strike-slip faults, as are observed (Fig. 10). Furthermore, there is a zone of shear-stress concentration trending roughly east-northeast. This zone coincides approximately with the plate boundary on the Reykjanes Peninsula, as defined by earthquake swarms (Klein et al., 1977). Also, many strike-slip earthquakes have this trend (Vogfjörd et al., 2005).

Thus, because of their mechanical interaction during dyke injection and plate pull, the volcanic systems on the Reykjanes Peninsula give rise to local stress fields that encourage strike-slip faulting. These results are as yet only suggestive; they need to be worked out in more detail, refined, and tested on data from this part of Iceland.

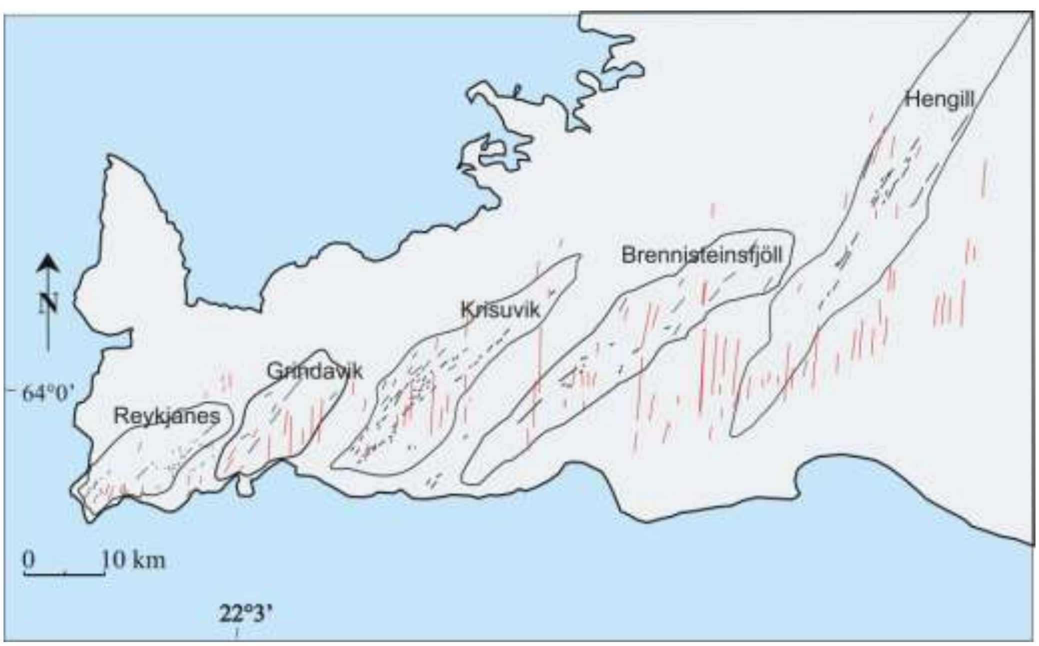

Fig. 10. Volcanic systems on the Reykjanes Peninsula (cf. Figs. 1, 7), with some of the tectonic and volcanic fissures indicated as well as the N-S trending dextral strike-slip faults (indicated by straight, red lines). Data on volcanic systems from Jakobsson (1978a,b), Jakobsson et al. (1978), and Johannesson and Saemundsson (1998), and other sources. Data on the N-S trending strikes-slip faults from Clifton and Kattenhorn (2006).

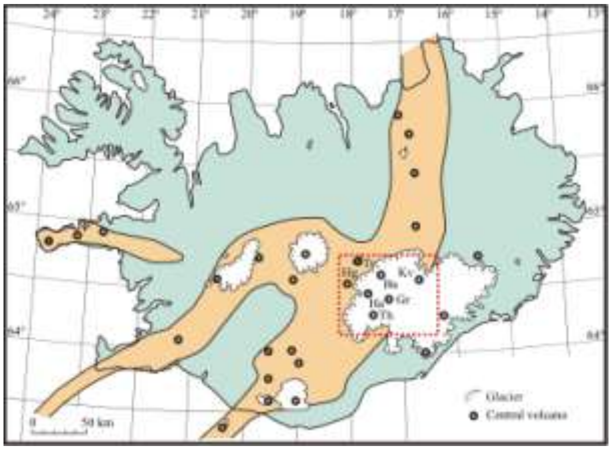

Fig. 11. Location of the central volcanoes in Central Iceland that are modelled in Figs. 13 and 14. The central volcanoes indicated are the same, and with the same abbreviations, as in Fig. 1.

Central Iceland

We also studied the mechanical interaction between the central volcanoes of the volcanic systems. We selected the part of Iceland with the greatest number of active central volcanoes per unit area (Figs. 11, 12) and which is located above the centre of the Iceland Mantle Plume (Wolfe et al., 1997; Gudmundsson, 2000a). The selected central volcanoes are: Tungnafellsjökull, Vonarskard, Hagöngur, Bardarbunga, Hamarinn, Grimsvötn, Thordarhyrna, and Kverkfjöll. Many of these volcanoes are highly active. For example, during historical time (the past 1100 years) the Grimsvötn Volcano has erupted about 70 times and Bardarbunga at least 23 times (Thordarson and Larsen, 2007). These data suggest that in terms of eruption frequencies, Grimsvötn is the most active central volcano in Iceland, seconded by Bardarbunga. Geophysical studies indicate that many of the central volcanoes contain collapse calderas and most, or all, are supplied with magma from a partially molten shallow magma chamber (Gudmundsson and Högnadottir, 2007). These results agree with general geological studies of deeply eroded, extinct central volcanoes in Iceland which indicate that they normally contain shallow magma chambers (exposed as plutons) with roofs at the depths of 1.5-3 km (Gudmundsson, 2006). As for the collapse calderas, while most have single calderas, Kverkfjöll has a double caldera and Grimsvötn a triple caldera (Johannesson and Saemundsson, 1998; Gudmundsson and Högnadottir, 2007). In the models below, Tungnafellsjökull and Vonarskard are taken as a single large central volcano, Tungnafellsjökull, and the two calderas associated with the Kverkfjöll Volcano are modelled as one large caldera. Thus, in contrast with the model by Gudmundsson and Andrew (2007), we here model the volcanoes in this field not as 8 but rather as 7, with two of the volcanoes, Tungnafellsjökull and Kverkfjöll, being 12 km in diameter rather than the typical diameter of 8 km (Figs. 13, 14). The average distance between the nearby volcanoes is about 30 km (Fig. 13).

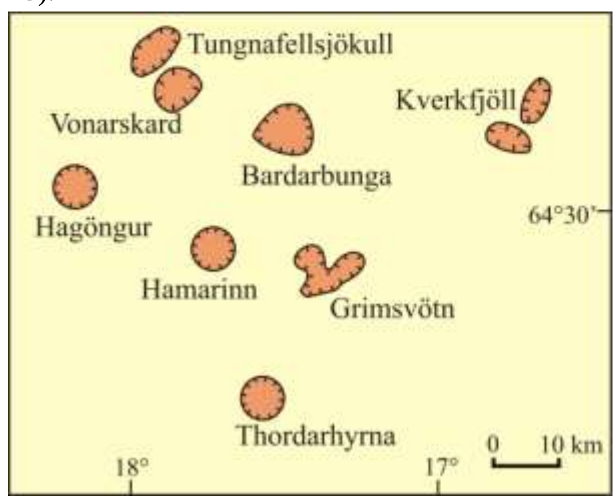

Fig. 12. Detailed map of the modelled central volcanoes in Central Iceland. In the numerical models in Figs. 13 and 14, the central volcanoes Tungnafellsjökull and Vonarskard are treated as a single, large volcano (as many authors do), and the double caldera of Kverkfjöll is taken as a single, large caldera. Data from Johannesson and Saemundsson (1998).

We assume that each volcano has a shallow magma chamber (and, in addition, most have an active collapse caldera) and model it as a hole (a cavity in three dimensions) in the elastic crustal plate. The models are two-dimensional, but we have also made some three-dimensional models, with the magma chambers modelled as

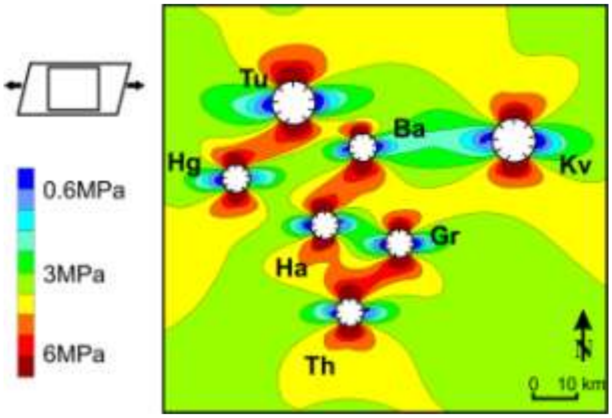

Fig. 13. Numerical model of the tensile stresses, in mega-pascals, around the 7 volcanoes in Central Iceland (Figs. 11 and 12). The volcanic field hosting the volcanoes is regarded as being composed of a homogeneous and isotropic crust with a Young's modulus of 20 GPa and a Poisson's ratio of 0.25. The only loading is tensile stress of 5MPa in the direction of the spreading vector (as indicated). Tu = Tungnafellsjökull, Hg = Hagöngur, Ba = Bardarbunga, Ha = Hamarinn, Th = Thordarhyrna, Gr = Grimsvötn, and Kv = Kverkfjöll.

oblate spheroids. The results are basically very similar, indicating that the two-dimensional models give as accurate results of stress fields between the volcanoes as are needed for the present purpose of analysing their mechanical interaction.

Since all the volcanoes are located within the rift zone (Figs. 11, 12), they are subject to plate pull. Accordingly, we model the volcanoes as holes subject to tensile stress of 5 MPa, which is similar to the commonly estimated maximum in situ tensile strength (Haimson and Rummel, 1982; Schultz, 1995), again using Ansys. The host rock is assumed homogeneous and isotropic with a Young's modulus of 20 GPa and a Poisson's ratio of 0.25, both values being typical for the uppermost part of the crust in Iceland (Gudmundsson, 2006). The applied tensile stress of 5 MPa is parallel with the spreading vector in this part of Iceland, that is, in the direction of N105°E.

The results indicate a strong mechanical interaction in terms of zones of stress concentration, particularly tensile stress, between all the nearby central volcanoes (Fig. 13). The existence of these zones indicates that unrest in one volcano can trigger unrest in a nearby volcano, in particular they may share the emplacement of dykes. Here dykes are referred to as "shared" when simultaneously formed dyke segments (propagating laterally, vertically or both) eventually link up two volcanoes so as to allow magma transport between them. For a dyke to be shared between two volcanoes, it is necessary that, in addition of a tensile-stress zone connecting the pair of volcanoes, the directions of the principal stresses, the stress trajectories, in the high-stress zone between the volcanoes be suitable.

Dykes generally follow the trajectories of the maximum principal compressive stress, σ_1 (Gudmundsson, 2006). A plot of the σ_1 – trajectories between the 7 volcanoes (Fig. 14) shows that the conditions for dyke sharing are satisfied between many volcanoes forming pairs. For example, there are zones of high tensile stress with suitable trends of σ_1 – trajectories for dyke sharing between the volcano pairs Tungnafellsjökull and Hagöngur, Bardarbunga and Hamarinn, and Bardarbunga and Grimsvötn, and Grimsvötn and Thordarhyrna (Figs. 13, 14).

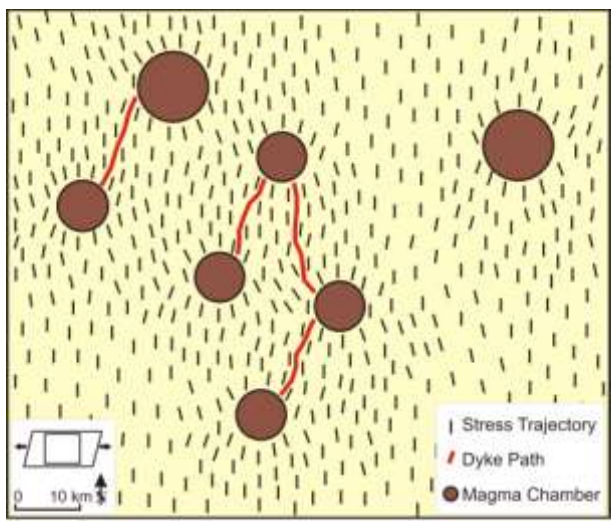

Fig. 14. Trend (trajectories) of the maximum principal compressive stress, σ_1, around the 7 volcanoes in Fig. 13. Ideal dykes would propagate parallel with these trends. Some possible shared dykes between volcanoes in a pair are indicated (schematically) by thick, red lines.

7. Discussion

The results presented here indicate that mechanical interaction between volcanic systems can have large effects on the associated local stress fields. These interactions are best described as stress concentration zones or stress transfer between systems. In terms of the models presented here, the stress-transfer zones arise from two different loading conditions, and two different geometrical constraints. The loading conditions are either tensile stresses related to plate pull operating parallel with the spreading vector or magmatic pressure generated by injected dykes. The geometrical constraints relate to the shapes of the modelled structures. In two dimensions, a volcanic system can be modelled as an elongated elliptical hole or an inclusion, whereas many central volcanoes are best modelled as circular or slightly elliptic holes (Figs. 7-14).

The main difference between the two loading conditions is that while plate pull operates continuously and has a small magnitude, the pressure due to dykes operates discontinuously and its magnitude may be quite large. Thus, the plate pull stress close to the surface cannot exceed the tensile strength which, for the basaltic rocks in Iceland, is normally not greater than 5-6 MPa (Haimson and Rummel, 1982; Schultz, 1995). However, even with this comparatively small loading, there develop stress concentration zones between nearby central volcanoes, such as in Central Iceland, which may encourage simultaneous activity (earthquakes, dyke injections) in two or more volcanoes (Figs. 13, 14).

There is, indeed, evidence for simultaneous activity in the volcanoes of Central Iceland. For example, Larsen et al. (1998) found evidence for at least five simultaneous eruptions in two or more of these volcanoes in the past eight centuries. Also, Gudmundsson and Högnadottir (2007) interpret gravity ridges between some of these volcanoes as dense dyke swarms, suggesting that the volcano pairs are connected by dykes some of which may have been shared between the volcanoes. Also, some authors interpret earthquake data from the 1996 Gjalp eruption in the Vatnajökull Ice Sheet (Gudmundsson et al., 1997) as indicating dyke sharing between the central volcanoes Bardarbunga and Grimsvötn (Einarsson et al., 1997; Gudmundsson and Andrew, 2007). Similar results have also been obtained from other volcanic areas (Cailleau et al., 2007).

Plate pull can apparently also give rise to zones of shear stress that encourage the formation of, or slip on existing, strike-slip faults. In particular, as a result of plate pull there develop north-south trending zones of shear stress between the nearby ends of the volcanic systems on the Reykjanes Peninsula (Figs. 9, 10). The magnitudes of the shear stresses are not very high, but the trends of the stress trajectories in parts of the regions between the volcanic systems would encourage slip on north-trending dextral faults. Such faults are, indeed, common on the Reykjanes Peninsula (Clifton and Kattenhorn, 2006).

The plate-pull stress builds up gradually over many years. By contrast, the magmatic overpressure generated by a dyke injected into a volcanic system can easily reach tens of mega-pascals (Eq. 3) and the associated loading is, in geological context, essentially instantaneous. The overpressure that generates a dyke fracture (Eq. 4) results in displacement of the host rock to either side of the dyke. The resulting displacement can be up to tens of metres (Figs. 4, 5), although more commonly of the order of metres (Figs. 3B, 6). A vertical dyke emplaced in a volcanic system causes a temporary change in the local stress field around that system. And when dyke injections occur in closely spaced volcanic systems, the local stresses thus generated may encourage seismogenic faulting.

Emplacement of overpressured dykes in the closely spaced, and partly overlapping, volcanic systems on the Reykjanes Peninsula (Fig. 7) generates local shear-stress zones between the systems (Fig. 9). The stress fields are complex, but could encourage both dextral and sinistral strike-slip faults. In particular, some of the well-documented north-trending dextral strike-slip faults on the Reykjanes Peninsula may be related to shear stresses caused by the magmatic overpressure associated with simultaneous, or near simultaneous, emplacement of dykes in nearby volcanic systems.

The numerical models presented in this paper indicate that, for central volcanoes and volcanic systems that are spaced at comparatively small distances, mechanical interaction may be common. In particular, volcanoes located within specific fields subject to loading such as plate pull are likely to interact mechanically in various ways (Figs. 11-14). Some may share seismogenic faults or dykes (Fig. 14) while others may deform in harmony. For example, deformation data indicate mechanical interaction between the closely spaced collapse calderas of the Galapagos Islands (Amelung et al., 2000). In view of the results presented in this paper, there is clearly a need for a closer look at mechanical interaction between volcanoes and volcanic systems worldwide.

8. Conclusions

- The Holocene rift-zone volcanic systems in Iceland are essentially gigantic cracks or elastic inclusions (Jaeger and Cook, 1979; Gudmundsson, 2006) that at the surface are composed of tectonic fractures, basalt volcanoes and, normally, a central volcano (composite volcano, caldera). The systems are mostly 5-20 km wide and 40-150 km long (Figs. 1, 2). At deeper levels, subvertical, regional dykes (Figs. 3-5) are the most important tectonic elements outside the central volcanoes, whereas inside the central volcanoes there are swarms of inclined sheets and dykes (Fig. 6). The inclined sheets are mostly 0.5 m thick (Fig. 6), whereas the regional dykes are commonly 3-6 m thick and occasionally as thick as 50-60 m (Fig. 4).

- When a dyke becomes emplaced its magma carries with it overpressure, that is, pressure in excess of the stress normal to the dyke (Eqs. 1-4). Because most dykes are extension fractures, the stress normal to the dyke is usually the minimum principal compressive stress, σ_3. A regional dyke is mostly injected from a deep-seated reservoir at the depth of 10-20 km, located in the lower crust or at the crust-mantle boundary (Fig. 2). Normally, the dyke magma has a density that is considerably less than the average crustal density, so that buoyancy contributes to the dyke overpressure which, thereby, can reach tens of mega-pascals (Eqs. 2,3).

- There are two principal ways by which rift-zone volcanic systems become loaded, namely through dyke-induced magmatic overpressure (Eqs. 2,3) and through plate pull (Fig. 1). In this paper we show that both loading conditions give rise to mechanical interaction between volcanic systems in general, and their central volcanoes in particular.

- Mechanical interaction caused by simultaneous emplacement of overpressured dykes is tested on the volcanic systems on the Reykjanes Peninsula (Fig. 7). The results show that zones of high shear stress develop between the nearby ends of the systems (Fig. 8A). While the detailed stress field is quite complex (Fig. 8B), it can trigger both dextral and sinistral faults, but in different regions between the volcanic systems. So far, only N-trending dextral faults have been observed on the Reykjanes Peninsula (Fig. 10), but both N-trending and ENEtrending sinistral faults are known from the nearby South Iceland Seismic Zone where, in fact, some individual faults show evidence of both dextral and sinistral slip.

- Mechanical interaction due to plate pull is tested on the volcanic systems of the Reykjanes Peninsula and a cluster of central volcanoes in Central Iceland. As for the Reykjanes Peninsula, the stress fields generated between the volcanic systems again encourage the formation of strike-slip faults (Fig. 9). As for Central Iceland, high-stress zones develop between many of the nearby central volcanoes, encouraging simultaneous seismogenic faulting and shared dykes. These model results are supported by various geophysical and geological data.

Acknowledgements.

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