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THE SKAGAFJÖRÐUR UNCONFORMITY NORTH ICELAND AND ITS GEOLOGICAL HISTORY

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CONTENT

Acknowledgements

The Geology of the Skagafjörður Valleys

A Summary

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Abstract

This first paper on the subject matter of my PhD thesis "The Skagafjörður Unconformity, North Iceland, and its Geological History" is a summary of work consisting of 7 papers, an appendix and a full colour bedrock map in the scale 1:50.000. Its purpose is to summarize the contents of the papers, explain their interrelations, describe the role of the appendix and highlight the main conclusions of the thesis. The Skagafjörður strata pile is divided into three main successions; the Quaternary succession, the sedimentary succession of the unconformity and the Neogene succession. The lowest one is separated from the others by the major Skagafjörður Unconformity. The lower part of the Neogene succession originated in the Snæfellsnes-Húnaflói Rift Zone that became extinct 6-7 Ma. Its upper part originated in the still active N. Iceland Rift Zone. In between is an unconformity that has not been recognised in the field but appears in the radiometric dates. Here it is called the Hidden Unconformity and is assumed to represent a rift relocation from the Snæfellsnes-Húnaflói Rift Zone to the presently active North Iceland Rift Zone. The largest formation of the Neogene pile is the Tinná Central Volcano. It was active during the period 6-5 Ma. The Skati rhyolite dome is the most prominent member of the volcano. It was formed through a powerful explosive eruption and heavy tephra fall, along with an extrusion of a lava dome, 5.5 Ma. It is the most voluminous monogenetic acidic formation known in Iceland both with respect to lava and tephra, corresponding to 18 km^3 of dense rock. The tephra layer is correlated to an acid ash layer found at ODP-site 907, 500 km NNE of Iceland. A collapse caldera was formed during the final stage of the volcano.

Unique geological circumstances make it possible to recreate the evolution of the topography of the Skagafjörður Valleys through time. This history spans 9 million years from the origin of the area near the diverging boundaries of the N. American and Eurasian crustal plates, drifting out from the accretion zone, subsequent erosion and

finally the sculpturing of the recent landscape. It is concluded that already at the end of Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had evolved. In the principal valley of Skagafjörður the erosion level was lower than at present and the mountains on each side were as high or even higher than they are today. The main tributary valleys were deeply eroded and many of the smaller tributaries also existed. The glacial erosion of the Pleistocene is of less importance than the Neogene erosion. By analogy it is assumed that all the main fjords and valleys in the Neogene regions of Iceland are old and already existed in the Pliocene and some even in Miocene times.

Volcanism started again in the Skagafjörður Valleys at the beginning of the Pleistocene. Lava flows covered the eroded Neogene landscape and the Skagafjörður Unconformity was formed. It is suggested that all the Pleistocene volcanic rocks above it belong to a temporary axial rift zone, the Skagafjörður Zone. Eruptive sites belonging to it are dispersed in the area between Hofsjökull and the mouth of Skagafjörður. Volcanism and rifting in the area started around 1.7 Ma. The activity culminated in the early Pleistocene but declined in the late Pleistocene and in the Holocene the activity seems to be restricted to the Hofsjökull Central Volcano. It is suggested that a decline in the activity of the Iceland Mantle Plume 2-3 million years ago caused a rift jump away from the hot spot below Vatnajökull and the formation of the temporary Skagafjörður Rift Zone. It bridged for a while the shortest way between the Reykjanes-Langjökull Zone and the Kolbeinsey Ridge.

The Skagafjörður Unconformity marks an abrupt change in the strike and dip of the strata sequence and in the composition of the sediments. The eruptives also change, compound lavas and lava shields, rare in the Neogene strata pile of Skagafjörður, become abundant in the Pleistocene pile above the unconformity. By comparison with postglacial volcanic history and lava production in Iceland it is concluded that this difference above and below the unconformity is due to early Pleistocene glaciations and isostasy. Data on the age and volumes of the postglacial lava shields reveal a production pulse culminating in the early Holocene 9,000-10,000 years ago.

Investigations on rift jumps, new Ar/Ar dates and discussions on the origin of the Neogene strata pile in Skagafjörður led to a re-evaluation of the spreading rate in Iceland. The Tinná Volcano seems to have drifted surprisingly far away from the volcanic zone and many other formations examined do not fit to the generally accepted plate velocities. Iceland is spreading towards east and west away from a spreading axis inside the country. Consequently the oldest rocks are found at the extreme east and west shores480 km apart. There the age is 13 and 15 Ma respectively. This gives a spreading rate equal to 3.4 cm/yr, compared with the conventional 1.8 cm/yr spreading of the North Atlantic. The spreading can be calculated in several other ways. They all give similar results, i.e. about 3.4 cm/yr. This seems valid for N. Iceland but the spreading is uneven from north to south, reaching a maximum over the mantle plume, where it might be > 6 cm/yr. The excess spreading diminishes along the rift axis and becomes normal, 1.8 cm/yr, 300-400 km from the centre of the hot spot.

Introduction

The Skagafjörður Unconformity, North Iceland, and its Geological History is the title of my PhD-thesis, written at the Geological Museum in Copenhagen. It is divided into 9 main sections:

Article 1: The Geology of the Skagafjörður Valleys (a summary article)

Article 2: The Late Miocene Tinná Central Volcano, North Iceland

Article 3: The Pliocene Valleys of the Skagafjörður District, North Iceland. (With J. Bonow)

Article 4: The Skagafjörður Zone – An ephemeral Rift Zone

Article 5: $^{40}Ar/^{39}Ar$–Dates from the Skagafjörður Valleys, North Iceland– Implications for Rift Relocations and the Deep-Sea Sedimentary Record. (With Harðarson and Pringle)

Article 6: Postglacial Lava Production in Iceland

Article 7: Crustal Spreading in Iceland

Geologic map: The Skagafjörður Valleys – Bedrock Map

Appendix: The Skagafjörður Valleys – Geological Description

This first section of the work is intended to serve as a summary for the whole thesis, summarizing the articles, explaining the role of the appendix and highlighting the main conclusions.

The geological investigations presented here along with the enclosed bedrock map have been carried out during the few last years mainly in association with hydropower planning, and some preliminary results have appeared in reports published by the National Energy Authority in Iceland (Orkustofnun) (Hjartarson et al. 1997, Hjartarson and Hafstað 1999). There the main emphasis was on geotechnical factors such as the stability and hardness of the rocks, their leakage and their quality for tunnelling. In this work, on the other hand, the emphasis is more on the stratigraphy, geomorphology, the rock age and the general geological history of the area.

The appendix contains the background data and all the details of the investigation, such as an historical overview, stratigraphic descriptions, definitions, methodology, subject analysis and various associated ideas and hypothesis. The articles deal with the main scientific yields of the investigations. The articles overlap each other in various ways because each of them was written as an individual paper. The articles also overlap the appendix, which nevertheless contains alternative additional information that is presented either very briefly or not at all in the articles, such as the research history of the area, stratigraphic definitions, descriptions of field methods, discussions about paleomagnetism, tectonics, geochemistry and a more detailed stratigraphy.

The Skagafjörður Valleys

The main emphasis in the project is on the geology of the research area (Figs. 1 and 2), i.e. the area covered by the enclosed map sheet, The Skagafjörður Valleys – Bedrock Map, hereafter simply referred to as the Bedrock Map. But in many cases the descriptions and discussions go beyond those limits, and thus the whole district from the Hofsjökull glacier to the tip of the Skagi Peninsula is in focus (Article 4 in this thesis) and in some cases the whole country (Articles 6 and 7). The research area, framed by the Bedrock Map, covers an area of about 820 km². The region is a mountainous terrain with the highest peaks reaching over 1,000 m. Exposures are in general good and in most cases the lithologies can be studied in canyons and gullies. The main farmlands of the Skagafjörður valleys lie north of the map borders, while towards south is the glacier Hofsjökull, an active central volcano with an ice-filled caldera and a summit reaching over 1,800 m a.s.l. (Björnsson 1988). The climate is fairly mild in the Icelandic context; the area is located in the precipitation shadow of the glacier and the highest growth limit of birch wood in Iceland is to be found there. The North Iceland Volcanic Belt is 100 km to the east of the study area and the centre of the Iceland Hot Spot is somewhat farther to the southeast (Sæmundsson 1979). The age of the volcanic pile described here spans about 9 million years from the late Miocene up through the Pliocene and Pleistocene.
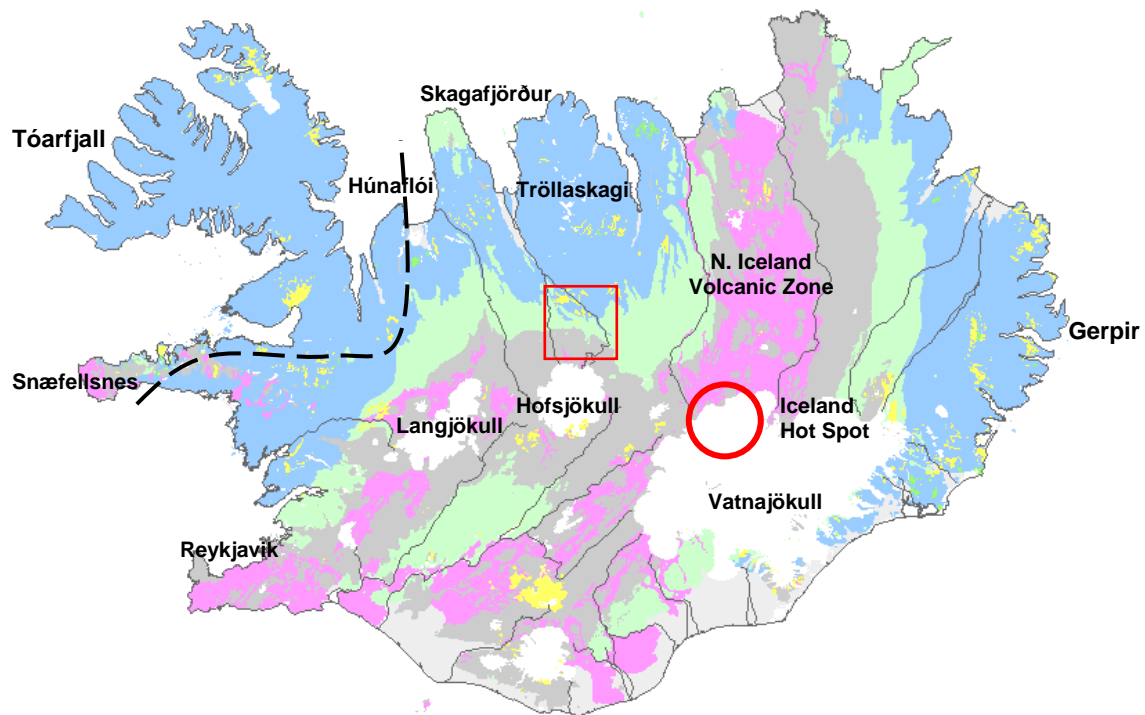
Fig. 1. Geological map of Iceland. Blue = The Neogene regions. Green = The Plio-Pleistocene areas (3.3-0.8 Ma). Grey = Late Pleistocene. Violet = Holocene lavas. Yellow = rhyolite. Red circle = The Iceland Hot Spot. Broken line = The Snæfellsnes – Húnaflói rift axis. The research area is indicated by a red frame. (Base map modified from Jóhannesson and Sæmundsson 1998a).
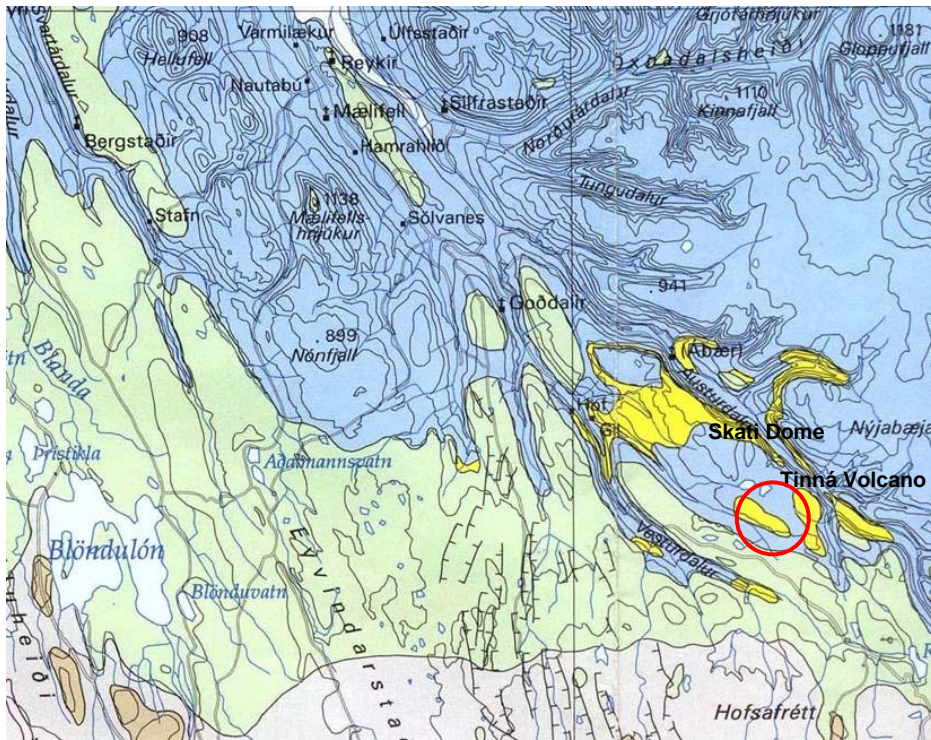
Fig. 2   The Skagafjörður valleys, showing the main geological features. Blue = Neogene succession. Green = Plio-Pleistocene (3.3-0.8 Ma). Grey = Late Pleistocene. Yellow = rhyolites of the Tinná Central Volcano. The red circle marks the approximate alignment of the caldera. (The base map is from Jóhannesson and Sæmundsson 1998a).

The main Successions of the Pile and the Unconformity

The Skagafjörður strata pile is divided into three successions, the lowest of which is separated from the others by a major unconformity:

1. The Quaternary succession
2. The sedimentary succession of the unconformity
3. The Skagafjörður Unconformity
4. The Neogene succession

Briefly, the geological history is as follows: The Neogene volcanic succession below the unconformity originates in an axial rift zone at the boundary between the North American and Eurasian crustal plates. According to Pálmason's (1980, 1981) model of crustal generation in Iceland, the volcanic belt at the plate margins was most probably similar in size to what it is today: 40-50 km wide. When the area drifted out of the volcanic zone, erosion took over, weathering the surface and carving out new landscape. The late Neogene-early Quaternary topography and the initial form of the Skagafjörður valley system can still be seen below more recent formations in the district. At a later stage a

sedimentary succession was deposited in the valleys, tens or even hundreds of meters thick in some places. After that temporary volcanic activity started again in the district and lavas of the Quaternary succession were erupted. They flowed along the old valleys, filled some of them up and covered extensive areas. In the late Pleistocene the volcanism died down and glacial erosion became the leading factor creating the current landscape of the district.

Articles 2-5, listed above, deal with different parts of this history. Articles 6 and 7, on the postglacial lava production and spreading rates, discuss geological problems and questions that arose during work on the thesis.

The main topics and conclusions of the articles are described below.

The focal point of the thesis is an unconformity in the strata pile, as is indicated in its title, *The Skagafjörður Unconformity, North Iceland, and its Geological History*. According to textbooks, an unconformity is a manifestation of a break in the sequence of strata in a certain area that represents a period of time during which no deposition took place. It may be a result of uplift and erosion or an interruption in the sedimentation. The absence of rocks normally present in a sequence, indicates a hiatus in the geological record (Dunbar and Rodgers1957). As an unconformity refers to the structural relations above and below a break in the geological record, a hiatus represents the break, or the erosion surface itself. To describe an unconformity and a hiatus it is necessary to study the successions above and below it and find out their age and evolutionary history.

The Skagafjörður Unconformity is angular in most places. In most places the underlying Neogene pile dips 4°-10° towards the south while the Pleistocene pile dips 0°-4° in a similar direction. The angle between the successions is widest in the northernmost part of the area but approaches a parallel alignment in the south (see Appendix and Chapter 6; see also the Bedrock Map). These geological conditions result in a widening of the time gap within the unconformity, from 1-2 million years in the south to up to 7 million years in the north (see Article 5 in this thesis).

The Tinná Central Volcano

The total thickness of the Neogene succession inside the research area is about 1,600 m (see the Bedrock Map). The proportional thickness of each rock type is given in Table 1. The percentage of acid and intermediate rocks is much higher than in adjacent areas such as Tröllaskagi (Sæmundsson et al. 1980) and Langidalur (Kristjánsson and Jóhannesson 1992). The reason is that here the section cuts the massive formations of evolved rocks belonging to the Tinná Central Volcano whereas the sections in Tröllaskagi and Langidalur avoid such volcanoes.

The Tinná Central Volcano is one of about 50 known central volcanoes in the Neogene regions of Iceland (Jóhannesson and Sæmundsson 1998b). It became active in the North Iceland Rift Zone about 6 million years ago and remained active for about 1 million years. The oldest acidic formation of the volcano is a rhyolite lava dome, the Ágúll dome, associated with a tephra layer.

Table 1. Thickness of different rock types in the Neogene succession of the Sakagafjörður Valleys

Type	Thickness (m)	%
Tholeiite	850	53
Olivine tholeiite	100	6
Porphyritic basalt	190	12
Intermediate lavas	90	6
Acidic lavas	310	19
Sediments	70	4

About a half a million years later or so, the largest formation of the volcano, the Skati dome, was formed in an violent eruption during extrusion of viscous lava dome accompanied by heavy tephra fall. It is the most voluminous monogenetic rhyolite formation known in Iceland both with respect to lava and tephra (see Article 2 in this thesis). The area of the acid lava dome is about 80 km² and its volume is 8 km^3. In addition, the tephra layer might correspond to 10 km^3 of dense rock, making a total volume of 18 km^3. The tephra layer is correlated to an acid ash layer found at ODP-site 907, 500 km NNE of Iceland as described by Lacasse and Garbe-Schönberg (2001). Two other rhyolite domes belonging to the Tinná Volcano have been mapped. A collapse caldera was formed in the final stage of the volcano. After its extinction the volcano was buried in a more recent lava pile as it drifted away from the active rift zone.

The Pliocene Valleys

Unusual geological circumstances make it possible to recreate the evolution of the topography in the Skagafjörður Valleys step by step (Article 3). This is possible because the Pleistocene volcanic succession covered, and thus preserved, the landscape that had been evolving for several million years during the late Neogene. The ancient landscape can still be observed in many places below the more recent formations. This history spans the 9 million years from the origin of the area near the diverging boundaries of the North American and Eurasian crustal plates, its drift out from the accretion zone, its erosion and finally the sculpturing of the recent landscape. Observations on the unconformity and the hiatus along with Ar/Ar dates (see Articles 4 and 5 in the thesis) reveal that as early as the end of the Pliocene, before the onset of the Pleistocene glaciations, a mature valley system seem to have been carved out in the primordial lava plain of the Skagafjörður District. In the principal valley the erosion level was lower than today and the mountains on each side as high or even higher than at present. The main tributary valleys, Norðurárdalur, Austurdalur, Vesturdalur and Savartárdalur, were deeply eroded and many of the small tributaries of the second order also existed. The glacial erosion of the Pleistocene seems to be of less importance than the Neogene erosion.

A. The Tertiary plateau 6 Ma

B. The Pliocene valley 2 Ma
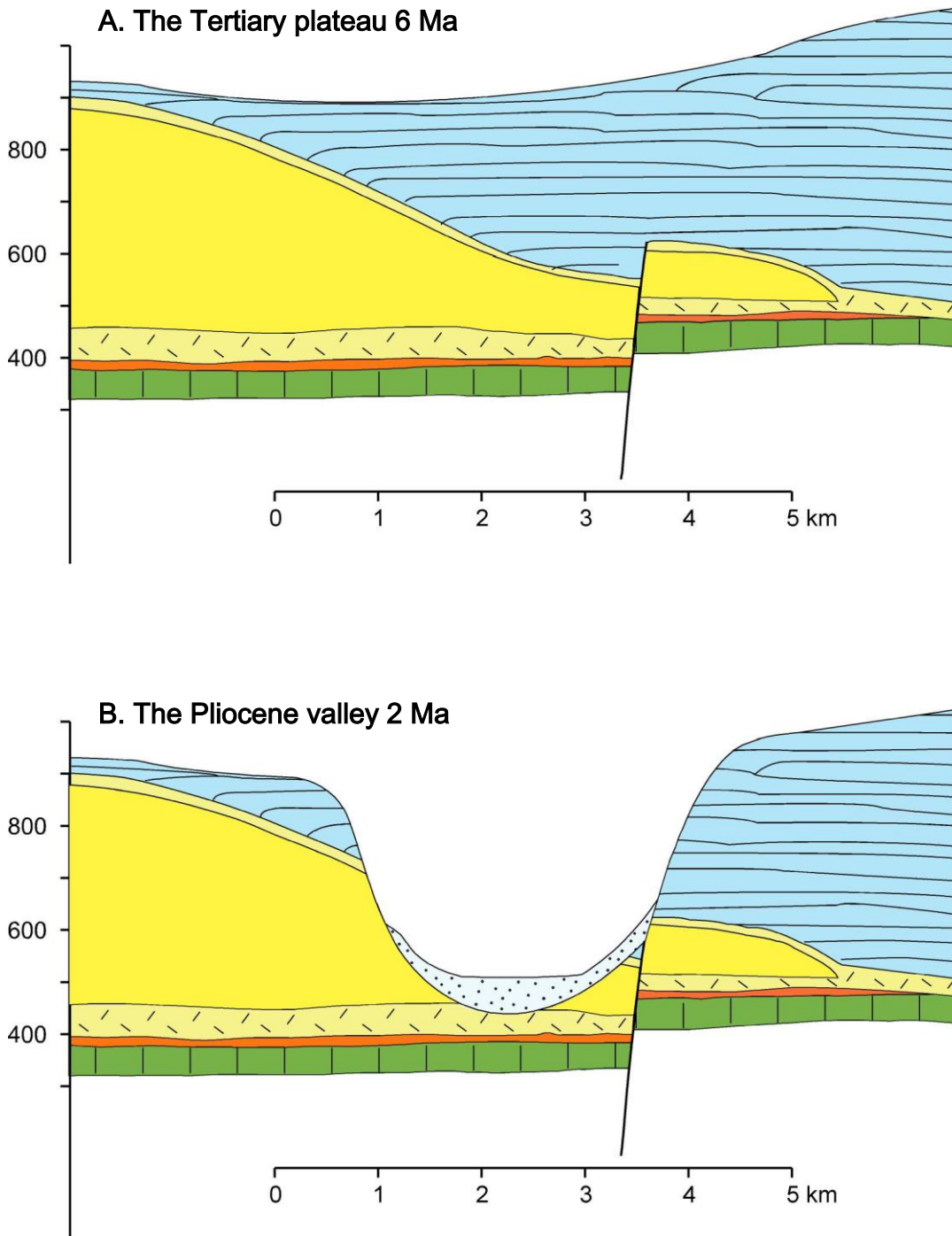
Fig. 3. *This sequence of figures, A-D, indicates the development of the Austurdalur valley from the time of the peneplanation of the Neogene lava pile up to the present date. Figure A shows the situation just after the area had drifted out of the volcanic zone. A Neogene lava pile (blue) banks up against the Skati rhyolite dome (yellow). Figure B, the Pliocene valley 2 Ma, shows the situation after three million years of erosion. The valley is 400-500 m deep with thick layers of fluvial sediments.*

C. The Early Pleistocene Valley 1.5 Ma

D. The Holocene valley
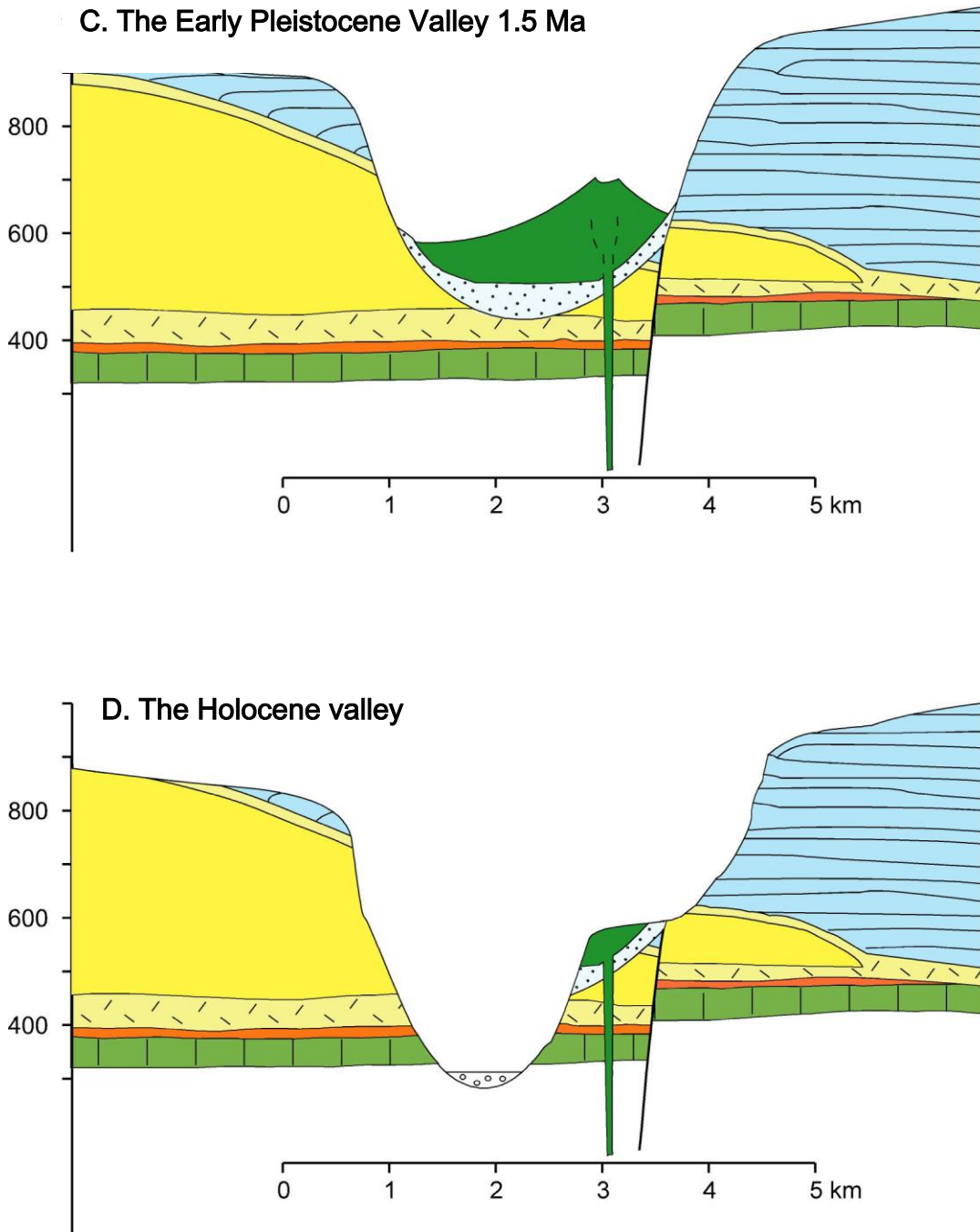
Fig. 3 (cont.) *Figure C. In the early Pleistocene an eruption in the valley built up the Austurdalur Pleistocene Volcano that covered the sedimentary layers and dammed up the valley for a while. Figure D shows a cross-section through the Austurdalur valley at present. It is 600 m deep and the traces of the volcano and its feeder dyke can be seen in its eastern slopes.*

It is possible to divide the history of the landscape in the Skagafjörður Valleys into six main stages spanning the interval from its genesis up to the present (Table 2, see also Fig. 3). The following age estimates are present mean values. In reality this development did not occur at exactly the same time all over the area. The valley formation started earlier in the principal valley than in the tributary valleys.

1. The formation of the Neogene volcanic succession took place at the Eurasian and North American plate boundaries 9-5 Ma (Sæmundsson et al. 1980, Article 5 in this thesis).

2. As the area drifted gradually out of the volcanic belt a peneplanation started and the erosion removed 100-200 m of the topmost layers of the lava plain forming the extensive lava plateau of central North Iceland (Hjartarson 1973).

Table 2: Main stages in the development of the Skagafjörður Valleys

No.	Stage	Time Ma
1	Formation of the Neogene volcanic succession	9-5
2	Peneplanation	~5
3	Formation of the Pliocene valleys and uplift of the Tröllaskagi plateau	5-3
4	Formation of a sedimentary succession and unconformity	3-2
5	Formation of the Pleistocene volcanic succession	2-1
6	Formation of the current landscape	1-0

3. More intensive weathering excavated the initial Skagafjörður valley system into the volcanic plateau 4-3 Ma. In the Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had been carved out. At the same time the plateau was raised by 200-300 m (Article 3).

4. In late Pliocene the deepening of the valleys halted and they started to fill up again, mostly with fluvial sediments. An angular unconformity was formed 3-2 Ma.

5. In the early Pleistocene a temporary episode of volcanism started up, continuing the filling of the valley system. The volcanic activity culminated in the early Pleistocene, but died out in the late Pleistocene, 0.5 Ma (Articles 4 and 5).

6. During the last million years or so, glacial and glaciofluvial erosion has cleaned out the ancient valleys and removed most of the Pleistocene volcanic succession, leaving only traces here and there. Although the valley system is more evolved than before the Pleistocene erosion is of less degree than the Neogene denudation (Article 3 in this thesis).

By analogy it can be predicted that all the main fjords and valleys in the Neogene regions are old in an Icelandic geological context and already existed in the Pliocene era and some even in the Miocene.

This scenario is in contradiction to a common interpretation stating that the Neogene topography was rather even, and that the landscape was dominated by an extensive volcanic plateau with low crater rows and occasional shield volcanoes, but without deep valleys, and the only high mountains were central volcanoes (Th. Einarsson 1968, 1991). According to this view, the present rugged topography of Iceland is mainly due to glacial erosion and subglacial volcanism during the Pleistocene. The conclusion of the present study is more in line with the interpretation that the Icelandic volcanic plateau was deeply eroded already in Neogene times and the main valleys and valley systems existed as early as in the Pliocene (T. Einarsson 1962, 1971).

The Skagafjörður ephemeral rifting and rift relocations

Volcanic activity started again in the area in the early Quaternary (see Article 4). New lavas covered the land and filled up the Pliocene valleys. The age of the oldest Quaternary volcanic formations in the Skagafjörður Valleys is around 1.7 Ma (see Article 5). The volcanism culminated in the early Pleistocene and died out in the late Pleistocene. In the Holocene this activity was entirely restricted to the Hofsjökull Central Volcano. The geological setting, with recent volcanic formations, tectonics and geothermal activity on a distinct belt between Hofsjökull and the Skagi peninsula, has lead to the conclusion that this was a short-lived and immature rift zone (Article 4). For a while it bridged the shortest way between the Reykjanes and the Kolbeinsey ridges. The zone was left incomplete in the sense that the activity was short-lived and did not form a tectonic syncline or a flexure zone or central volcanoes, with the exception of Hofsjökull.

A dominant process in the evolution of Iceland is the repeated eastward relocation of the spreading axis as it responds to the westward migration of the Eurasian-American plate system away from the Iceland Mantle Plume. During rift relocations the active spreading axis at the plate boundary is displaced and shunted for tens or hundreds of km in a short period of time.

Rift relocations (often called rift jumps) in Iceland, or a large eastward displacements of the spreading axis, with resulting unconformity, were first postulated by Ward (1971). The idea was supported by the observations of Sæmundsson (1974), Pálmason and Sæmundsson (1974) and Wilson and McElhiny (1974), but disputed by Mussett et al. (1980). Today it is generally accepted that large rift relocations have occurred inside the country although their tectonic process and history is not known in detail.

Several rift relocations are known in the geological record of the country. The Northwest Rift Zone is thought to have formed some 24 million years ago west of the Northwest Peninsula. (Hardarson et al. 1997). It was active for 8-10 million years but then the focus of spreading was shunted to the east, forming the Snæfellsnes-Húnaflói Rift Zone, which was active for another 8-10 million years. Then, 6 million years ago, the active spreading axis was shunted to its present location forming the Reykjanes-Langjökull Zone and the North Iceland Zone (Sæmundsson 1974,1979, Johannesson, 1980, Kristjánsson and Jónsson, 1998). The most recent rift relocation in North Iceland occurred when the Skagafjörður Rift Zone was generated at 1.7 million years ago as a temporary rift axis and stayed active for about a million years (Article 5). In South Iceland the picture is more complicated as the Eastern Volcanic Zone seems to be an evolving spreading axis

(Sæmundsson 1979). It was initiated 2-3 million years ago and is slowly propagating to the southwest, taking over the spreading of the Reykjanes-Langjökull segment (see Table 4 in Article 5). Figure 4 shows how the basalt regions of Iceland can be divided between the different spreading zones.
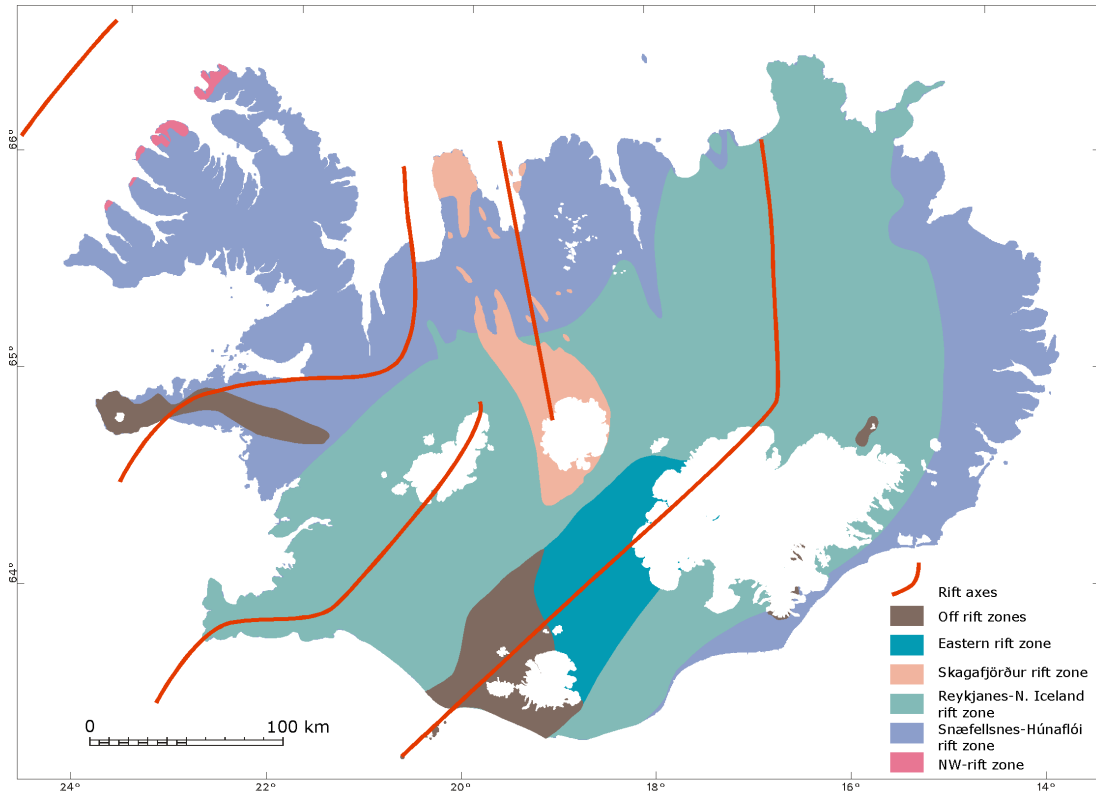
Fig. 4. The volcanic rocks of Iceland originate in several spreading zones. The picture indicates old and new spreading axes and the division of the country between the spreading zones.

The Skagaförður rifting was a remarkable event. A rift jump seems to have occurred from the North Iceland Rift Zone with a 100 km displacement towards the west. This jump was in the opposite direction to previous recorded ridge relocations in Iceland, away from the hot spot. The ephemeral existence of the Skagi Volcanic Zone might have been caused by pulses in the mantle plume. At certain times the activity of the plume culminates, but at other times it declines (Vogt 1971). It has been argued that rift jumps occur during the culminations (Abelson and Agnon 2001, Jones 2003). Here it is suggested that the reverse can also happen (Article 4) and in a phase of a low activity the spreading zone can slip away from the plume. The physical background for the geodynamic process resulting this is poorly understood and no attempt is made here to explain it, but a possible scenario will be described. The mantle plume beneath Iceland might have been in a phase of low activity at 1-2 Ma. It lacked energy to control the North Iceland Rift Zone and keep it in place and it failed to maintain the transform faults and their associated earthquakes. A new rift zone started to form without any transform faults and connected the Reykjanes and Kolbeinsey ridges directly. This became the Skagafjörður zone that temporarily took over a part of the spreading in North Iceland. Later on the

power of the mantle plume increased again. Volcanic activity increased in the North Iceland Volcanic Zone but faded out in the Skagafjörður area. The zone there was left incomplete.

Compound Lavas and Lava Shields

The mapping work in the Skagafjörður Valleys revealed that compound lavas and lava shields are rare in the Neogene strata pile of Skagafjörður, but become abundant in the Pleistocene pile above the unconformity. The reason for this is somewhat unclear. It has been noted that many of the postglacial lava shields of Iceland are of early Holocene or late glacial age (Thorarinsson et al. 1959, Sigvaldason and Steinþórsson 1974, Jakobsson et al. 1978) and some interaction with the glacial isostasy, along with rapid ice unloading, has been proposed (Guðmundsson 1986).

Table 3. Lava shields (volumetric order)

Name	Age [1]	Estim. age [2]	Area km²	Volume km³	Ref.
Stóravítishraun	11090-11980T	11300	470	18.4	7, 26
Skjaldbreiður	9600 S	9600	200	17	3, 25
Trölladyngja	> 7000 S	7250		15	4
Kollóttadyngja	> 4500 S	8000	69.1	14.5	6
Skildingahraun	>11980T	12200	250	10	8, 30
Heiðin há	7500 S	7500	150	6	8
Kerlingardyngja	> 4000 ?	6000		6	10
Kjalhraun	7800 T	7800	150	6	7, 25
Lambahraun	~3700 [14]C	4100	160	6	7, 25
Þráinsskjöldur	~13000 S	13000	130	5.2	11,28
Sandfellshæð	~12500 S	12500	120	4.8	11,28
Þingvallahraun	9130 [14]C	10200	200	4	3
Ketildyngja – Laxárhr. e.	4300 T	4300	300	4	12, 19
Hallmundarhraun	1100 T, [14]C	1100	225	3.4	7, 27
Hrútagjárdyngja	4000-5000 T	4500	80	3.2	11
Leitahraun	4560 [14]C	5200	100	3	11
Flatadyngja	3500-4500 T	4400	110.5	2.6	6
Selvogsheiði	8200 S	8200	50	2.2	11
Gjástykkisbunga	11080-11980T	11500	50	2	30
Litladyngja	2900-3500 T	3200	85.1	1.7	6
Útbrunahraun	> 10300T	10500	80	1	7
Skuggadyngja	>4500T	5900		1	30
Svartadyngja	> 4500 S	6800	38.7	1	6
Surtsey	AD 1964 H	30		1	15

1) Measured ages, [14]C = radiocarbon, T = tephrrachronology, S = stratigraphy, I = ice core, H = historical, 2) Corrected and estimated ages are in calendar years before AD 2000. **References**: 3 Sæmundsson 1992, 4 Kjartansson 1966, 6 Sigvaldason et al. 1992, 7 Hjartarson unpubl., 8 Hjartarson 1999, 10 Guðmundsson 2001, 11 Jónsson 1978, 12 Thorarinsson 1951, 15 Thorarinsson 1968, 16 Thordarson et al. 2001, 19 Thorarinsson 1979, 25 Sinton et al. 2001, 26 Sæmundsson 1973, 27 Jóhannesson 1989, 28 Sæmundsson 1995, 30 Sæmundsson pers.

The study has been confined to postglacial times. In the literature it has never been discussed whether unloading, as a result of earlier glaciations and ice melt, have caused increased production of compound lavas and shields. Therefore it seems to be of interest to pose some questions on the matter, such as: Is the activity of lava shields in general, and in Skagafjörður especially, stimulated by mass unloading and isostasy?

Shield volcanoes are generally thought to be monogenetic (Walker 1971, Jakobsson et al. 1978) and emitted from a single circular crater or a short fissure. The eruptions are thought to have been long-lasting but of low effusion rate with very little tephra production (Rossi 1996). The lavas are of olivine tholeiite or picrite composition and occur mainly at the margins of or outside the fissure swarms associated with the central volcanoes. No picrite is found in the Skagafjörður field.

At first it is necessary to examine the suggestion that the postglacial lava shields are related to deglaciation and isostasy. Guðmundsson (1986) and Jakobsson et al. (1978) found this valid for the Reykjanes Peninsula. Other workers state that this is valid for all postglacial volcanism in Iceland, the lava shields, the fissure lavas and the central volcanoes (Sigvaldason et al. 1992, Sigvaldason 2002, Jull and McKenzie 1996). If all the postglacial lava shields in Iceland are viewed, the picture should become more general. The problem is that information on ages is rather uncertain, especially in the North Iceland Volcanic Zone. Most of the large lava shields are, however, sufficiently well dated. In this investigation the large shields will be observed. A large shield is defined as ≥ 1 km^3 in lava volume. Twenty-four such volcanoes are known (Table 3).
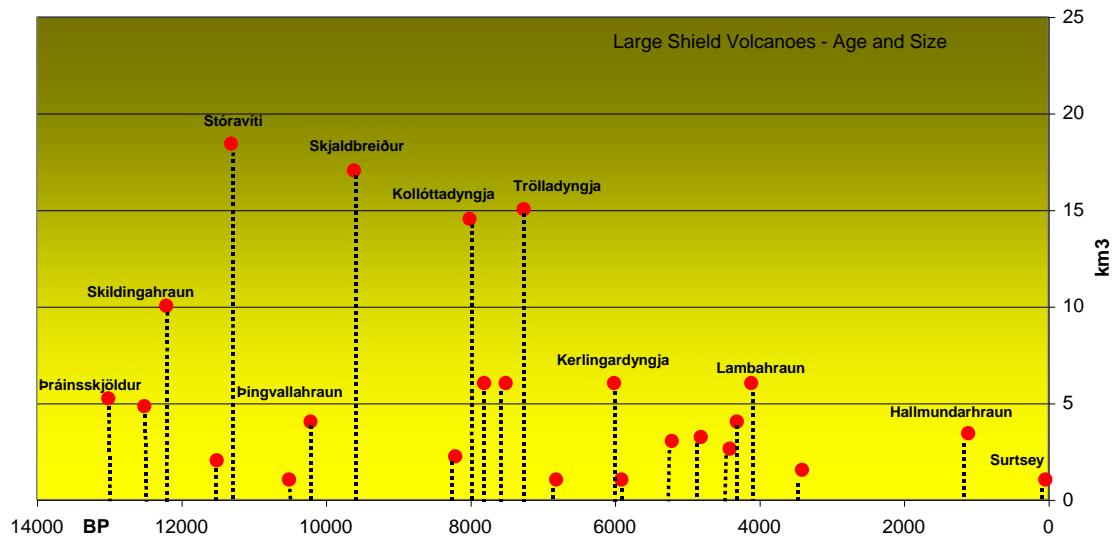
Fig. 5. Size and dating of 24 large postglacial shields. The lava productivity culminates in the early Holocene, with an ongoing decline during the Holocene.

Fig. 5 shows several large postglacial eruptions where the volumes of the large shields are plotted against their ages. The lava production of the shields started as soon as the glaciers retreated inland from the shore, and the largest shield, Stóravítishraun, in Þeistareykjarbunga, N. Iceland, erupted in late glacial times (Sæmundsson 1973). The lava production of the shields decreased in the mid-Holocene, but not the frequency of

the eruptions. Around 4,000 years ago a sudden decrease occurred and since then only three are known to have erupted (Litladyngja, Hallmundarhraun and Surtsey); this might be similar to the Neogene frequency of such volcanism. No small shields erupted in this period. Therefore it will be stated that the period of the shields lasted from 13,000 to 4,000 BP, i.e. for 9,000 years. Fig. 6, illustrating the lava production of the large shields per millennia, demonstrates this even better. There, the early Holocene production pulse is striking. According to this, some relationship between lava shields, deglaciations and isostasy might be suggested. But the occurrence of the shield volcanism during and after the deglaciation could also as well be a coincidence and caused by some other reasons, such as fluctuations in the Iceland Mantle Plume.
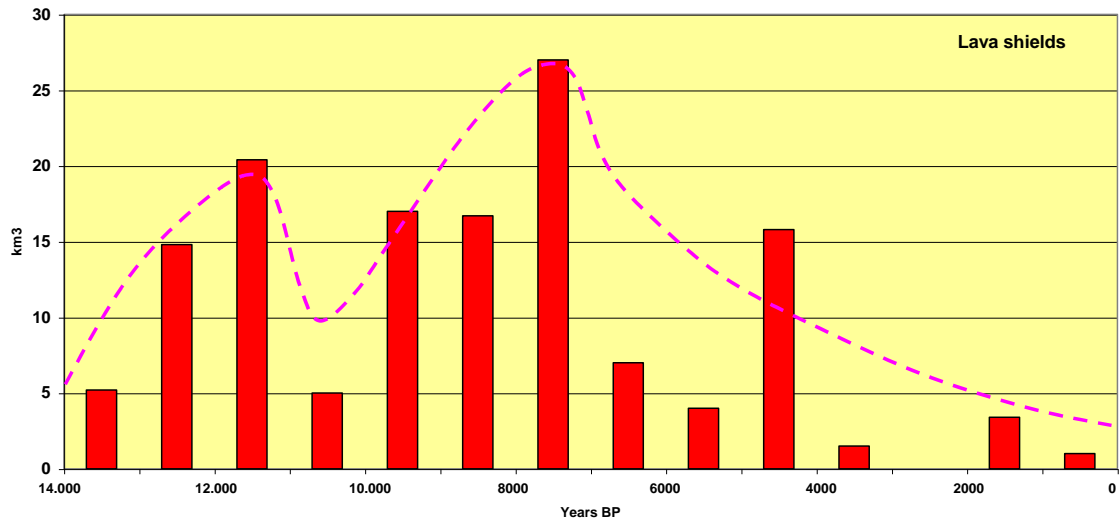
Fig. 6. Lava production of the large shields per millennia. The picture indicates early Holocene production pulses.

Observations showing that lava shields suddenly become abundant in the Pleistocene pile in Skagafjörður after the beginning of the glaciations (Appendix, Chapter 5), might support the former suggestion. This difference has also been observed in Fnjóskadalur, North Iceland, where compound lavas are rare in the lower part of the pile (Miocene), but become abundant in the Plio-Pleistocene pile above it (Jancin et al. 1985). In the Reykjavík region, an increasing abundance of shields during the late Pleistocene also seems evident (Torfason et al. 1997, 1999, 2000). An increasing number of lava shields in the volcanic pile of these regions, after the onset of the ice age, endorse the relations between shields and deglaciations. The conclusion is therefore that the abundance of lava shields in the Pleistocene pile of the Skagafjörður Valleys, compared to how rare they are in the Neogene pile, demonstrates their relations with glaciations and crustal rebound.

Whether this is also the case with the fissure lavas and central volcanoes remains unanswered. Article 6, "Postglacial Lava Production in Iceland", deals with the matter.

Origin of the Skagafjörður Neogene succession and the spreading rate in Iceland

It has not been known with any certainty where the Neogene volcanic succession of the Skagafjörður Valleys originated, i.e. whether it was in the Snæfellenes-Húnaflói Zone or in the North Iceland Zone. According to the new radiometric dates (Article 5), the lower half of the Neogene succession is supposed to originate in the Snæfellsnes-Húnaflói Zone, while the upper half originated in the North Iceland Zone.

The Ar/Ar-age of the Tinná Volcano indicates its origin in the North Iceland Volcanic Zone. Since then it has drifted out of the zone towards west and belongs to the North American crustal plate, located some 100 km from the axis. It is, however, remarkable how far west it reached. Its initial location inside the volcanic zone, where it was formed, is not known with any accuracy but some estimates can be suggested. It must have originated west of the rift axis; otherwise it would have drifted towards the east, into the Eurasian crustal plate. It was later covered by younger lavas but secondary minerals and its stage of hydrothermal alteration indicate that it never went deep down into the crust (Article 2). It can therefore be suggested that its original locus was somewhere midway between the axis (where the Askja caldera is now) and the border of the zone, about 10 km from the axis. If this is so, then it has drifted 90 km away from its original site or birthplace. The Ar/Ar-age of the main formation of the volcano is 5.2 Ma. Now its velocity can be calculated:

90 km/5.2 Ma = 17.3 km/Ma

This is half the spreading rate, hence the full spreading would be 34.6 km/Ma. The spreading rate in Iceland has been estimated at 18 km/Ma (1.8 cm/year) in the direction N105E (DeMets et al. 1990, Guðmundsson 2000). The Tinná Volcano seems to be drifting faster than the crustal plate to which it belongs. This appears paradoxical. According to the present accepted spreading rate, the volcano should have drifted

9 km/Ma x 5.2 Ma = 47 km

If this is correct it would have originated outside the rift zone and belonged to the off-rift volcanoes, as is the case with Snæfellsjökull and Öræfajökull among the active volcanoes of Iceland today. The petrology and stratigraphy of the Tinná Volcano does not favour this interpretation (Article 2).

But what happens if some other formation is examined? A well-dated paleomagnetic transition zone in the mountain Sólheimafjall, central Skagafjörður District, seems to be appropriate (Sæmundsson et al. 1980, see also Article 5). The lavas in the transition zone, hereinafter called the reference lavas, are about 9.3 Ma in age. They belong to the lower part of the Neogene pile and originate in the extinct Snæfellsnes-Húnaflói Volcanic Zone.
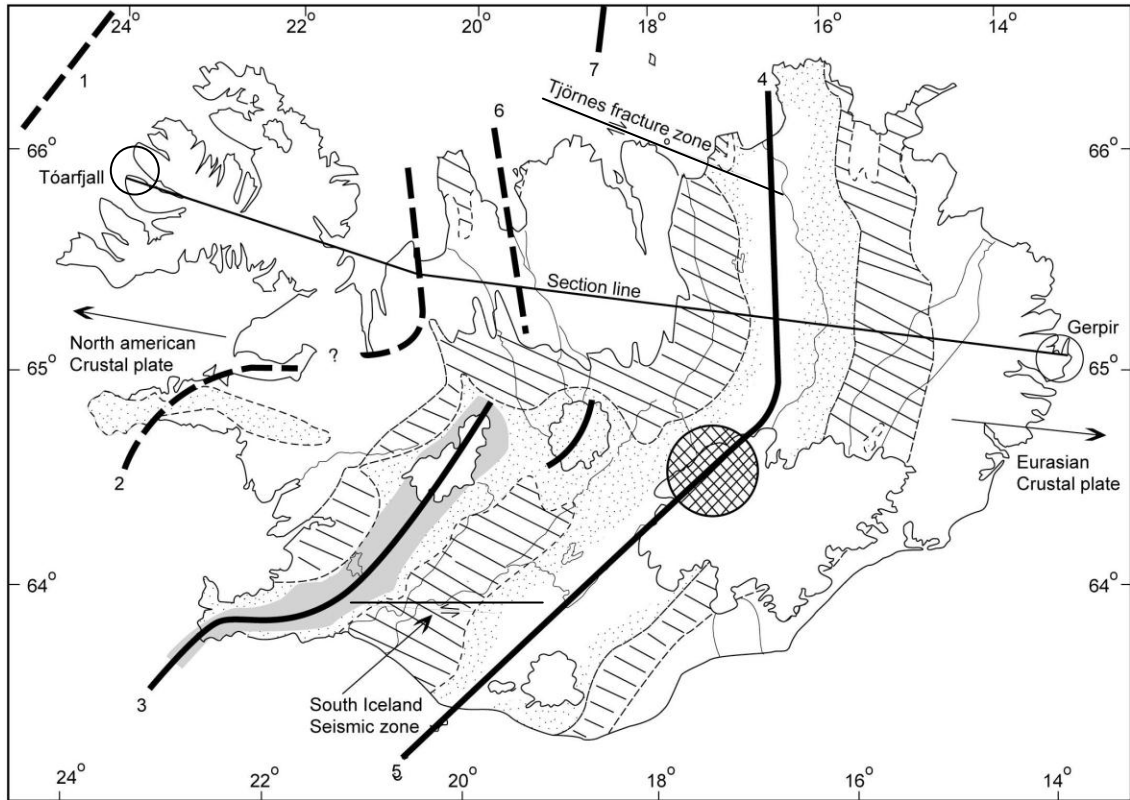
Fig. 7. Map of Iceland displaying the extinct rift axis (broken lines), the active rift axis (solid lines), the Iceland hot spot, the transform faults, the oldest rocks at Tóarfjall and Gerpir and the section line between them. The neovolcanic zones (0-0.8Ma) and the Plio-Pleistocene areas (0.8-3.3 Ma) are also indicated. (1 = Northwest Rift Axis; 2 = Snæfellsnes-Húnaflói Rift Axis; 3 = Reykjanes-Langjökull Rift Axis; 4 = North Iceland Rift Axis; 5 = The propagating Eastern Rift Axis; 6 = Skagafjörður Rift Axis).

The old rift axis is located roughly 80 km west of the Skagafjörður Valleys (Jóhannesson and Sæmundsson 1998b). Since the reference lavas were erupted they have drifted towards the east so they are now 68 km away from the old axis. Their initial location inside the volcanic belt is not known, but here it is assumed to have been midway between the axis and the eastern border of the volcanic zone, about 10 km from the axis. Then the reference lavas must have drifted 58 km towards the east, away from their volcanic vents. The spreading activity in the Snæfellsnes-Húnaflói Zone ceased some time about 6 Ma. A half-spreading rate therefore seems to be:

58 km/(9.3 - 6) Ma = 17.6 km /Ma

This example gives a full spreading equal to 35.2 km/Ma (3.5 cm/yr), once again far higher than the plate velocity.

Further examples are tested in order to find out whether something is missing from the argument. The geological formations discussed here are the oldest dated formations in the extreme east and west of Iceland, the Gerpir central volcano and the Tóarfjall series (Fig. 7). Some more examples are taken in Article 5. Four main assumptions are made in the following discussion:

1. The spreading rate in Iceland is 18 km/Ma (1.8 cm/yr).

2. The oldest dated rocks in Iceland are 15 Ma in the west and 13 Ma in the east.

3. The spreading activity in the Snæfellsnes-Húnaflói Zone stopped between 7-5 million years ago and the North Iceland Zone took it over.

4. The formations in question originated in the Snæfellsnes-Húnaflói Zone.

The trajectories of Gerpir and Tóarfjall are described, from their origin in the volcanic belt to their current locations, in Article 5. This is also shown in the time-distance diagram in Fig. 8 below, where the drifts of Tóarfjall and Gerpir are reconstructed both according to the present accepted spreading rate and a higher rate of spreading. On the basis of the four assumptions stated above, it is hard to explain how these rocks managed to move so far away from their site of origin inside the Snæfellsnes Rift Zone. For the 15 million years old Tóarfjall series to cover the entire distance in the time available, a half-spreading rate of 17 km/Ma would have been required. For the 13 million years old Gerpir to reach its present location, the half-spreading rate must have been 17.5 km/Ma.
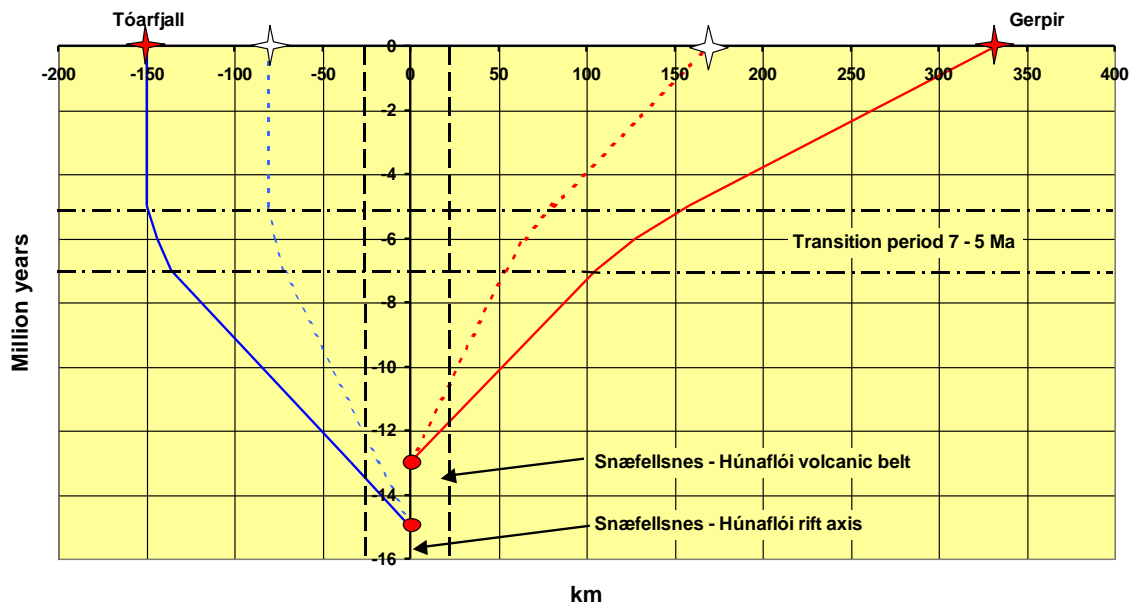
Fig. 8. Spreading in North Iceland (time-distance diagram). Solid lines indicate the trajectories of the Tóarfjall series and the Gerpir rocks from their origins to their recent locations with respect to the Snæfellsnes rift axis. Dashed lines indicate the trajectories for the 18 km/Ma spreading rate. (Further explanation in Article 5).

According to the scenarios above the apparent spreading rate in Iceland seems to be 34-35 km/Ma, i.e. 90-100% higher than generally accepted. This proves to be valid for West Iceland and for East Iceland as well as for central North Iceland. This conclusion can also be reached in a simple way for the whole country by dividing the width of the country by the average rock age in the west and east:

480 km/14 Ma = 34 km/Ma = 3.4 cm/yr

In this case the dating of the rift jump is no longer relevant, and assumption No. 3 can therefore be eliminated. In fact assumption No. 4 also can be omitted because the location of the spreading axis should not affect the outcome of the calculations, it only needs to be

inside the country. Only if substantial areas of older crust (> 15 Ma) are trapped in between the spreading axes would it affect the calculations, but then assumption No. 2 would be wrong. Now the options have been reduced: only assumptions Nos. 1 and 2, regarding ages and rates, still remain valid, and only one of them can be correct.

Table 4. Spreading Rates

	Type of formation	*Time span (Ma)*	*Spreading (km/Ma)*
Tinná Volcano	Central volcano	5.2	35
Mt. Sólheimafjall reference lavas	Basalt lava series	3	35
NW-Iceland (Tóarfjall)	Basalt lava series	9	34
E-Iceland (Gerpir)	Central volcano	13	35
All of Iceland (Tóarfjall–Gerpir axis)		14	34

The idea of a higher rate of spreading in Iceland than on the sea floor all around it has appeared earlier, though in a different context. Walker (1975) showed that the aggregate width of the active volcanic zones is greatest in South Iceland and, by postulating that this width is proportional to the spreading rate, inferred that the spreading rate in South Iceland might be several times greater than in North Iceland as well as in the adjacent mid-ocean ridges. Serle (1976) pointed out a serious objection to this idea. Such excess spreading would result in a major compressive deformation of at least one of the relevant lithospheric plates, but no such deformation was known. With respect to this it has been generally accepted that the spreading rate in Iceland is the same as at the submerged ridges in the north and south, i.e. 18 km/Ma.

The statement that the spreading rate in Iceland is much higher than on the Eurasian and North American crustal plates in general might be hard to accept, as Serle (1976) points out. But then something must be wrong in the dating instead. The ages should be higher. One possible explanation is that the dated rocks in Tóarfjall and Gerpir are from lavas that flowed tens of kilometres away from the eruptive sites in the direction of the spreading. If that is the case, then 20-30 million-year-old local rocks should be found in Northwest and East Iceland, for example in dykes or inside the central volcanoes. And that would be in contradiction with paleomagnetic data, polarity time scales and all the geochronology for Iceland (Kristjánsson et al. 1995). Moreover this argument does not match the Gerpir dates. These have been very carefully estabilshed and double checked because of their key stratigraphic position, first by Moorbath et al. (1968) and then by McDougall et al. (1976). The stratigraphy is also closely investigated by Walker (1959) and described by Watkins and Walker (1977). The magneto-stratigraphy is also carefully examined by laboratory measurements (McDougall et al. 1976, Kristjánsson et al. 1995). The Gerpir dates therefore seem to be very reliable and few places, if any, inside the Neogene regions have been more thoroughly researched.

After this discussion it seems evident that the strange drift of the Tinná Central Volcano, which in the beginning was suggested might be based on some erroneous assumptions, is

most likely real. In contrary to the generally accepted spreading rate in Iceland, 18 km/Ma, observations in the Skagafjörður District which also seem valid for the whole country, indicate a much greater spreading rate. The apparent spreading rate in between Tóarfjall in the west and Gerpir in east is about 34 km/Ma. The Icelandic land mass is spreading faster than the ocean floor all around it.

Conclusions

The main conclusions of the thesis will not be described here, only briefly summarized. All of them are given in the relevant articles. These conclusions are:

The Skati Dome of the Tinná Central Volcano is the largest monogenetic rhyolite formation found in Iceland, equivalent to18 km^3 of dense rock. It is 5.21 Ma old according to a new Ar/Ar date. It is correlated to a tephra layer at the ODP site 907, 500 km north of Iceland.

At the end of the Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had evolved in Skagafjörður. The principal valley and the main tributary valleys were deeply eroded and many smaller tributaries also existed. The glacial erosion of the Pleistocene seems to be of less importance than the Neogene erosion. By analogy it is assumed that all the main fjords and valleys in the Neogene regions of Iceland already existed in Pliocene and some even in Miocene times.

Around 1.7 Ma ago, volcanism and rifting started again in the Skagafjörður Valleys after being passive for a few million years. Lava flows covered the eroded Neogene landscape and the unconformity was formed. The activity culminated in the early Pleistocene but decreased in the late Pleistocene, and in the Holocene it is restricted to the Hofsjökull central volcano. The time gap of the unconformity increases from south to north, being 1-2 million years in the highlands north of Hofsjökull but reaching 7 million years in the central Skagafjörður district.

The Pleistocene volcanic rocks above the unconformity all are assumed to belong to a temporary axial rift zone, the Skagafjörður Zone. It is suggested that a decline in the activity of the Iceland Mantle Plume, 2-3 million years ago, caused a rift jump away from the hot spot below Vatnajökull and the formation of the zone.

Compound lavas and lava shields, rare in the Neogene strata pile of Skagafjörður, become common in the Pleistocene pile above the unconformity. By comparison with the postglacial volcanic history and lava production in Iceland it is concluded that this difference is due to early Pleistocene glaciations and isostasy. Data on ages and volumes of the postglacial lava shields in Iceland reveal a production pulse culminating in the early Holocene 9,000-10,000 years ago.

Investigations in Skagafjörður led to a re-evaluation of the spreading rate in Iceland. The spreading can be calculated on the basis of several different approaches. They all give similar results, i.e. about 3.4 cm/yr. This seems to be valid for Northwest and East Iceland as well as for central North Iceland.

References

Abelson, M. and Agnon, A. 2001: Hotspot activity and plume pulses recorded by geometry of spreading axes. Earth and Planetary Science Letters 189, 31-47.

Björnsson, H.1988: Hydrology of Ice caps in Volcanic Regions. Societas Scientiarum Islandica 45, Reykjavik, 139 pp + maps.

DeMets, C., Gordon, R.G., Argus, D.F. and Stein, S., 1990: Current Plate motions. Geophys. J. Int. 101, 425-478.

Dunbar, C. O. and Rodgers, J., 1957: Principles of Stratigraphy. John Wiley & sons, New York, 365 pp.

Einarsson, T., 1962: Upper Tertiary and Pleistocene rocks in Iceland. Societas Scientiarum Islandica XXXVI. Reykjavik, 197 pp + maps.

Einarsson, T., 1971: Eðlisþættir jarðarinnar og jarðsaga Íslands. Almenna Bókafélagið, Reykjavík, 267 pp.

Einarsson, Th., 1968: Jarðfræði. [Geology]. Mál og Menning, Reykjavík

Einarsson, Th., 1991: Myndun og mótun lands. [Formation and figuration of the landscape]. Mál og Menning, Reykjavík.

Guðmundsson, A. T., 2001: Íslandseldar. Mál og menning, Reykjavík.

Guðmundsson, Á., 1986: Mechanical Aspects of Postglacial Volcanism and Tectonics of the Reykjanes Peninsula, Southwest Iceland. Journal of Geophysical Research, 91, 12711-12721.

Guðmundsson, Á., 2000: Dynamics of Volcanic Systems in Iceland: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge System. Annual Review of Earth and Planetary Sciences 28, 107-140.

Hardarson, B.S., Fitton, J.G., Ellam, R.M. and Pringle, M.S., 1997: Rift relocation – a geochemical and geochronological investigation of a paleo-rift in NW Iceland. Earth and Planetary Science Letters 153,181-196.

Hjartarson, Á., 1973: Rof jarðlagastaflans milli Eyjafjarðar og Skagafjarðar og ísaldarmenjar við utanverðan Eyjafjörð. B.Sc. – thesis (unpublished). University of Iceland 37 pp. + map

Hjartarson, Á., 1999: Yfirlit um jarðfræði í Ölfusi og grend. ÁH 99/06, Orkustofnun, Reykjavík, 11 pp. + map.

Hjartarson, Á., 2002: The Skagafjord Ephemeral Volcanic Zone. In (ed. S. S. Jónsson). The 25th Nordic Geological Winter Meeting. January 6th – 9th, 2002 – Reykjavík, Iceland. Abstract Volume.

Hjartarson, Á., Friðleifsson, G. Ó. and Hafstað, Þ. H., 1997: Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020 Orkustofnun, Reykjavík, 55 pp. + maps.

Hjartarson, Á. and Hafstað, Þ. H., 1999: Jarðfræði á Hofsafrétt í Skagafirði. OS-99031, Orkustofnun, Reykjavík, 13 pp. + maps.

Jakobsson, S. P., Jónsson, J. and Shido F., 1978: Petrology of the Western Reykjanes Peninsula, Iceland. Journal of Petrology 19, 669-705.

Jancin, M., Young, K. D., Voight, B., Aronson, J. L. and Sæmundsson, K., 1985: Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. Journal of Geophysical Research 90, 9961-9985.

Jones, S.M. 2003. Test of ridge-plume interaction model using oceanic crustal structure around Iceland. Earth and Planetary Science Letters 208, 205-218.

Jóhannesson, H., 1980: Jarðlagaskipan og þróun rekbelta á Vesturlandi. Náttárufræðingurinn 50, 13-31.

Jóhannesson, H., 1989: Aldur Hallmundarhrauns í Borgarfirði. Fjölrit Náttúrufræðistofnunar Íslands nr. 9. 12 p.

Jóhannesson, H and Sæmundsson K., 1998a: Geological map of Iceland 1:500.000. Bedrock Geology. Icelandic Institute of Natural History, Reykjavik (2nd edition)

Jóhannesson, H and Sæmundsson K., 1998b: Geological map of Iceland 1:500.000. Tectonics. Icelandic Institute of Natural History, Reykjavik.

Jónsson, J., 1978: Jarðfræðikort af Reykjanesskaga. OS-JHD 7831, Orkustofnun, Reykjavík, 333 pp. + maps.

Jull, M. and McKenzie,D., 1966: The effort of deglaciation on mantel melting beneath Iceland. Journal of Geophysical Research 101, 21815-21828.

Kjartansson, G., 1966: Rúmmál hraundyngna. Náttúrufræðingurinn 36, 125

Kristjánsson, L., Jóhannesson, H. and McDougall, I., 1992: Stratigraphy, age and paleo-magnetism of Langidalur, Northern Iceland. Jökull, 42, 31-44.

Kristjansson, L., Guðmundsson, Á. and Haraldsson, H.,1995: Stratigraphy and paleomagnetism of a 3-km-thick Miocene lava pile in the Mjoifjördur region, eastern Iceland. Geol. Rundsch. 84, 813-830.

Kristjánsson L. and Jónsson, G., 1998: Aeromagnetic results and the presence of an extinct rift zone in Western Iceland. J. Geodyn. 25, 99-108.

Lacasse, C. and Grabe-Schönberg, C.-D., 2001: Explosive silicic volcanism in Iceland and Jan Mayen area during the last 6 Ma: sources and timing of major eruptions. Journal of Volcanology and Geothermal Research 107, 113-147.

Larsen, G. and Thorarinsson, S., 1977: H4 and other acid Hekla tephra layers. Jökull 27, 28-45.

McDougall, I., Watkins, N.D., Walker, G.P.L. and Kristjansson, L., 1976: Potassium-Argon and Paleomagnetic Analyses of Icelandic Lava Flows: Limits on the Age of Anomaly 5. Journal of Geophysical Research 81, 1505-1512.

Moorbath, S., Sigurdsson, H. and Goodwin, R., 1968: K:Ar ages of the oldest exposed rocks in Iceland. Earth and Planetary Science Letters 4, 197-205.

Pálmason, G. 1980. A continuum model of crustal generation in Iceland: Kinematic aspects. Journal of Geophysics 47, 7-18.

Pálmason, G. 1981. Crustal rifting and related thermomechanical processes in the lithosphere beneath Iceland. Geol. Rundsch. 70, 244-260.

Pálmason, G. and Sæmundsson, K., 1974: Iceland in Relation to the Mid-Atlantic Ridge. Annual Review of Earth and Planetary Sciences 2, 25-50.

Rossi, M.J. 1996: Morphology and mechanism of eruption of postglacial shield volcanoes in Iceland. Bulletin of Volcanology 57, 530-540

Serle, R., 1976: Spreading rate in Iceland. Nature 261, 75.

Sigurðsson, H., 1970: Structural origin and plate tectonics of the Snaefellsnes volcanic zone, western Iceland. Earth and Planetary Science Letters 10, 129-135.

Sigurðsson, H., Schilling, J.G. and Meyer, P.S., 1978. Skagi and Langjökull Volcanic Zones in Iceland., 1. Petrology and Structure. Journal of Geophysical Research 83, 3971-3982

Sigvaldason, G. E., 2002: Volcanic and tectonic processes coinciding with glaciation and crustal rebound: an early Holocene rhyolitic eruption in the Dyngjufjöll volcanic centre and the formation of the Askja caldera, north Iceland. Bulletin of Volcanology 64, 192-205.

Sigvaldason, G. E., Annertz, K. and Nilsson, M., 1992: Effect of glacier loding/deloding on volcanism: postglacial volcanic production rate of the Dyngjufjöll area, central Iceland. Bulletin of Volcanology 54, 385-392

Sigvaldason, G. E. and Steinþórsson, S. 1974: Chemistry of tholeiitic basalts from Iceland and their relationship to the Kverkfjöll hot spot. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, p. 155-164.

Sinton, J. 2001: Postglacial volcanics of the Western Volcanic Zone. Geologic map. University of Hanaii, Honolulu, HI, USA.

Sæmundsson, K., 1973 Straumrákaðar klappir í kringum Ásbyrgi. Náttúrufræðingurinn 43, 52-60.

Sæmundsson, K., 1974: Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. Geological Society of America Bulletin 84, 495-504.

Sæmundsson, K., 1979: Outline of the geology of Iceland. Jökull 29, 7-28.

Sæmundsson, K., 1992: Geology of the Thingvallavatn area. In: ed. Jónasson 1992, Ecology of oligotrophic, subarctic Thingvallavatn, pp. 434

Sæmundsson, K., 1995: Um aldur stóru dyngnanna á utanverðum Reykjanesskaga. In: Björn Hróarsson, Dagur Jónsson og Sigurður Sveinn Jónsson (eds): Eyjar í eldhafi. Gott mál, Reykjavík, pp.165-172.

Sæmundsson, K., Kristjánsson, L., McDougall, I. and Watkins, N. D.1980: K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. Journal of Geophysical Research 85, 3628-3646.

Thorarinsson S., 1951: Laxárgljúfur and Laxárhraun a tephrochronological study. Geografiska annaler 1951, 1-89.

Thorarinsson, S., 1968: Síðustu þættir Eyjaelda. Náttúrufræðingurinn 57, 113-135.

Thorarinsson, S., 1979: The postglacial history of the Mývatn area. Oikos 32, 17-28.

Thorarinsson, S., Einarsson, T. and Kjartansson, G. 1959: On the geology and geomorphology of Iceland. Geografiska Annaler 41, 143-153.

Thordarson, Th., Miller, D.J., Larsen, G., Self, S. and Sigurdsson H. 2001: New estimates of sulfur degassing and atmospheric mass-loading by the 934 AD Eldgja eruption, Iceland. J Volcanol Geotherm Res 108:33-54

Torfason, H., Hjartarson, Á., Jóhannesson, H., Jónsson, J. and Sæmundsson, K. 1997: Bedrock map, Viðey 1613 III NV-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason, H., Hjartarson, Á., Jóhannesson, H., Jónsson, J. and Sæmundsson, K. 1999: Bedrock map, Vífilsfell 1613 III SA-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason, H., Hjartarson, Á., Jóhannesson, H. and Friðleifsson, I. B. 2000: Bedrock map, Mosfellssveit 1613 III NA, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Walker, G.P.L., 1959: Geology of the Reyðarfjörður area, eastern Iceland. Geological Society of London Quaternary Journal 114, 367-393.

Walker, G.P.L., 1971: Compound and Simple Lava Flows and Flood Basalts. Bulletin of Volcanology 35, 579-590.

Walker, G.P.L., 1975: Excess spreading axes and spreading rate in Iceland. Nature 255, 468-470.

Ward, P. L. 1971: New Interpretation of the Geology of Iceland. Geological Society of America Bulletin 82, 2991-3012.

Watkins, N.D. and Walker, G.P.L., 1977: Magnetostratigraphy of Eastern Iceland. American Journal of Science 277, 513-584.

Wilson, R. L. and McElinny, M. W. 1974: Investigation on the large scale paleomagnetic field over the last 25 million years. Eastward shift of the Icelandic spreading ridge. Geophysical Journal of the Royal astronomical Society 39, 570-586.

The Late Miocene Tinná Central Volcano, North Iceland

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Abstract. The Tinná Central Volcano in the Skagafjörður Valleys belongs to the Neogene succession of N-Iceland. It was active during the period 6 – 5 Ma. The total volume of the volcano is 210 km^3. Four rhyolite lava domes are described. The largest one, the Skati Dome, was formed during an immense explosive eruption accompanied by a heavy tephra fall 5.5 Ma. It is the most voluminous monogenetic rhyolite formation known in Iceland both with respect to lava and tephra. The area of the dome is about 80 km² and its volume is 8 km^3. The tephra layer might additionally account for 10 km^3 of dense rock, or 18 km^3 altogether. The tephra layer is correlated to an acid ash layer found at ODP-site 907, 500 km NNE off Iceland's coast and can serve as an important marker horizon for the late Miocene in the deep-sea sediments. A collapse caldera was formed during the final stage of the volcano. After its extinction it was buried in younger lava pile while it drifted away from the active rift zone.

Introduction

The Tinná Central Volcano is one of 40-50 known central volcanoes in the Neogene regions of Iceland (Jóhannesson and Sæmundsson 1998). The name was introduced by Hjartarson et al. (1997) because the Tinná river and the Tinná valley are situated near the centre of the volcano. In Icelandic the word tinna (or hrafntinna) means obsidian and the name Tinná means "obsidian river". Obsidian is the black and glossy form of rhyolitic glass that is often found at the top or the bottom of acid lavas and thus indicates a central volcano. The first aim of this paper is to describe the volcano and put it in context with other Icelandic central volcanoes. The second aim is to introduce the Skati rhyolite member and emphasise the unique volume of this monogenetic eruptive formation. The third aim is to correlate the Skati tephra layer with an ash layer in the deep-sea sediment.

Regional and geological setting

The Tinná Volcano is located in the Skagafjörður Valleys, central North Iceland (N65°15', W18°50'. Fig.1). The district is a mountainous terrain with the highest peaks reaching over 1,000 m. Towards south the glacier Hofsjökull covers an active

but dormant central volcano with an ice-filled caldera and a summit reaching over 1,800 m a.s.l. (Björnsson 1988) The North Icelandic Volcanic Zone is 100 km to the east and the centre of the Iceland Hot Spot is somewhat farther to the southeast (Sæmundsson 1979). The Tinná Central Volcano belongs to the Neogene succession of North Iceland and is assumed to originate in the North Iceland Volcanic Zone. Exposures are in general good in the deeply eroded landscape. Excellent natural cross sections are cut down through the lava pile exposing the stratigraphy and tectonics of the volcano's surface formations but its roots and magma chamber are still covered and hidden deeper in the crust.

In the Eyjafjörður District, 25 km east of the Skagafjörður Valleys, there is another central volcano named Torfufell. It seems to be separated from the Tinná volcano by the high mountains of Nýjabæjarfjall but in fact this might only be an apparent separation because the young strata pile in the uppermost part of the mountains hides their interconnection. Torfufell has never been mapped or investigated with any accuracy. It is in the strike east of the Tinná Central Volcano and of similar age. Hjartarson et al. (1997) suggested it belonged to the Tinná Central Volcano. Jóhannesson and Sæmundsson (1998), on the other hand, interpret them as two individual central volcanoes on their geological map of Iceland. Here this problem will be left for later investigations.

Research history

Geological research in the Skagafjörður Valleys began in the 18th century (Ólafsson 1943, Pálsson 1983). The first observations concentrated on the lignite seams that were known in several places in the area; later investigations have revealed that these all belong to a sedimentary layer associated with the central volcano. In those days a great deal of attention was given to the lignite, in spite of its poor quality, because many people believed it might indicate valuable hidden coal deposits deep inside the mountains. Hallgrímsson [1989] visited Skagafjörður in 1839 to study and take samples from the supposed "coal seams" for the Danish government. In his report they are stated to be thin layers of lignite of poor quality and too small in quantity to be of any practical value. After this no geological investigations were made in the area for more than a hundred years.

The acid rocks in the Skagafjörður Valleys were first shown on Kjartansson's (1965) Geological Map of Iceland but there they cover far too limited an area. The areas were enlarged on later maps, but still the volcano is shown as far too small.

Various studies of the stratigraphy, local tectonics and geothermal activity of the Skagafjörður Valleys were carried out between 1970 and 2000, mainly in association with geothermal research and utilization and hydropower planning (Jónsson et al. 1977, Kaldal and Víkingsson 1978, 1979, Harðarson and Guðmundsson 1986, Guðmundsson 1991, Jóhannesson 1991, Hjartarson et al. 1997).

The Stratigraphy

The strata pile belonging to the Tinná Central Volcano has been defined as one stratigraphic group, the Tinná Group. It is divided further into 11 formations. These will be described in ascending order. Table 1 gives an overview of the stratigraphic names, the thickness of the units and their polarity. The classification has not been given a formal stratigraphical status.
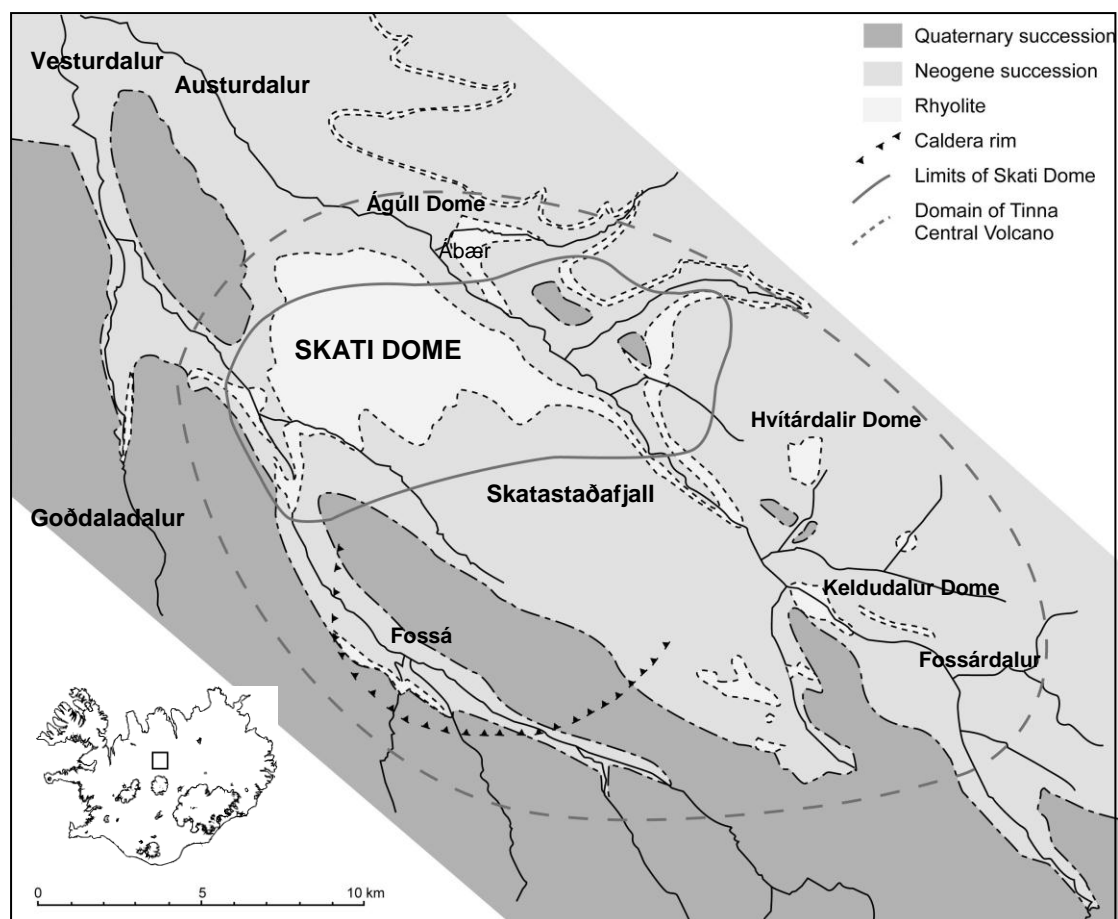
Fig. 1. A simple geological map of the Tinná Central Volcano. The domain of the volcano includes all the acid and intermediate lavas related to it. The approximate limits of the Skati Dome and the proposed alignment of the caldera are indicated.

Table 1. The stratigraphic units of the Tinná Group

Group	Formation	Average thickness m	Polarity*
Tinná	Nýibær rhyolite	80	N
	Nýjibær andesite	100	N
	Nýjibær tholeiite	150	N
	Tinná tholeiite	150	R
	Tinná andesite	50	R
	Skati rhyolite	100	R
	Tinná olivine tholeiite	25	R
	Tinná lignite sediments	18	
	Ábær tholeiite	280	N
	Ágúll rhyolite dome	100	N
	Fjóslækur olivine tholeiite	300	N/R/N/R/N

*N = normal polarity, R = reverse polarity
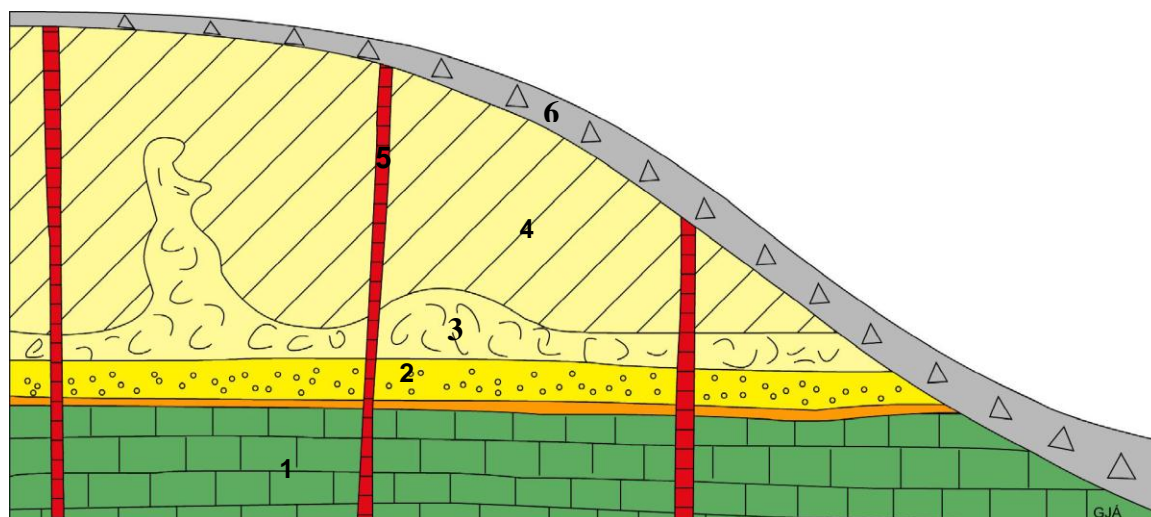
Fig. 2. The Ágúll rhyolite dome, a cross-section along the strike in Ábær river gorge. 1) Olivine tholeiite lavas. 2) Acid tephra. 3) Basal breccia. 4) Rhyolite dome. 5) Dyke. 6) Till cover.

The *Fjóslækur formation* is the lowest formation of the Tinná Group. In Austurdalur it is 350 m thick. In Vesturdalur it is 125 m thick. The Fjóslækur formation is dominated by normal paleomagnetic polarity (N) with two short reverse subchrons (R). The lower one, which contains only two lavas, wedges into the pile in the outer Goðdaladalur canyon. The other one is found near the top of the formation. No acid or intermediate layers have been recognized inside this formation.

The *Ágúll rhyolite dome.* The first clear signs of the onset of acid magmatism and the evolution of a central volcano in the Skagafjörður Valleys are found near the abandoned vicarage Ábær. A tributary river flows in rapids and cascades in a deep and rocky gorge from the hanging Ábær valley to form a broad alluvial fan joining the Jökulsá river in Austurdalur. On the north side the bedrock is covered by a rugged scree, while in the south wall of the gorge the river has excavated the lowest part of a large rhyolite extrusion that can be classified as a typical lava dome (Fink and Anderson 2000). This rhyolite is the lowest of several acidic layers in the Skagafjörður Valleys, and it indicates the early formation of a magma chamber underneath the area (Fig. 2).

The acid layers of the dome in Ábær gorge can be divided into three units. The lowest is a light-coloured, fine-bedded, sandy tephra, 15-20 m thick with coarser pyroclastic lenses containing angular fragments; most often these are 2-4 cm in diameter but can reach up to 50 cm. The upper part of the tephra layer is coarser than the lower one.

Brecciated rhyolite and acid scoria are found above the tephra, at 280 m a.s.l.. Angular rhyolite blocks, up to 1 m in diameter, are found in more finely-grained material. This unit attains a thickness of 10 m and forms the basal breccia of a massive lava dome. In some places this breccia forms huge pillows rising many meters up into the lava itself (Fig. 2).

The lowest 60 m of the lava are excavated in the gorge. The upper part is more or less capped by screes and till but reaches 500 m a.s.l. in the mountain slopes east of the gorge. The texture of the lava is fine-grained with small plagioclase phenochrysts that make up 30% of the rock volume. The rock is dense, consisting of irregular flow

layering and flow foliations and forming coarse polygonal columns. Cracks and fissures are most often perpendicular to the layering. The acid lava can be traced for 3 km upstream, along the Ábæjará river, where it disappears because of the local tectonic tilt (10° SE). Along the Jökulsá river it also can be traced 3 km upstream where it disappears below the bottom of the valley. A landslide covers the western part of the lava dome but it most probably terminates below the western part of the slide. The dome has not been found on the south slopes of the Austurdalur valley. Instead of it there is a thick sedimentary layer, consisting perhaps of reworked material from the dome and is its lateral equivalent. The surface breccia of the dome, or the contact between its top and the country rock, is not exposed. The acid dome is cut by numerous basaltic dykes related to the overlaying basaltic lava pile.

   The dimensions of the lava dome are not clear because only the northern part of it is exposed. The southern part is hidden below the bottom of the valley due to the regional dip. At the surface it can be traced for 5 km from east to west and 3 km from north to south. Its thickness reaches at least 250 m. The base of the dome might either be semi-circular or elongated (which is more likely), with the long axis running north-south according to the dominating tectonics. If, however, it is assumed to be circular, 5 km in diameter and with an average thickness around 100 m, its volume would be 0.6 km^3.

The Ábær tholeiites make up a thick pile of thin-layered lavas that bank up against, and cover, the Ábær dome. The thickness of the lava pile is 250 – 300 m. All interbeds are thin and seem to indicate intensive and continuous volcanism. The lavas most probably originate in flank eruptions inside the juvenile central volcano.

The Tinná lignite sediment is a fine-grained, light-coloured sand-siltstone, rather soft and with layers rich in acid volcanic tephra indicating contemporary acidic eruptions in distant central volcanoes. In several localities the layer contains lignite seams (Magnússon 1980, 1981). It extends through much of the area with a maximal thickness, 30 m, in the Goðdaladalur canyon. Its thickness varies from 3 to 30 m from place to place. The thickness of the lignite, on the other hand, is only 1-30 cm. The thickest lignite layers occur at 300 m a.s.l. in the Goðdaladalur canyon. Five seams can be seen, consisting of black, compressed wooden trunks and branches, inside a 4.4 m thick light-coloured layer of siltstone. Imprints of leaves or seeds are rarely seen. No investigation has been made of the flora or possible fauna in the layers in Skagafjörður. The Tinná lignite sediment indicates a time of quiescence before increased volcanic activity in the following ages. A reversal of the magnetic field occurred during the accumulation of the sediment.

Tinná olivine tholeiite. The deposition of the widespread Tinná sediment was stopped by an eruption of thick and extensive lava, the Tinná olivine tholeiite. This is the most voluminous basalt lava in the whole area. It varies in thickness from place to place between 20 and 60 m, and in some localities it is absent. In most places the layer is formed of one thick lava unit, unlike the more common composite olivine tholeiites that are made of several monogenetic lava bands. Its extent is at least 25 km x 10 km, i.e. 250 km². Its average thickness has been estimated at 25 m, thus its volume is 6 km³. This is not very large compared to the great Holocene lavas, but it is still the most voluminous reported Neogene basalt lava of Iceland.
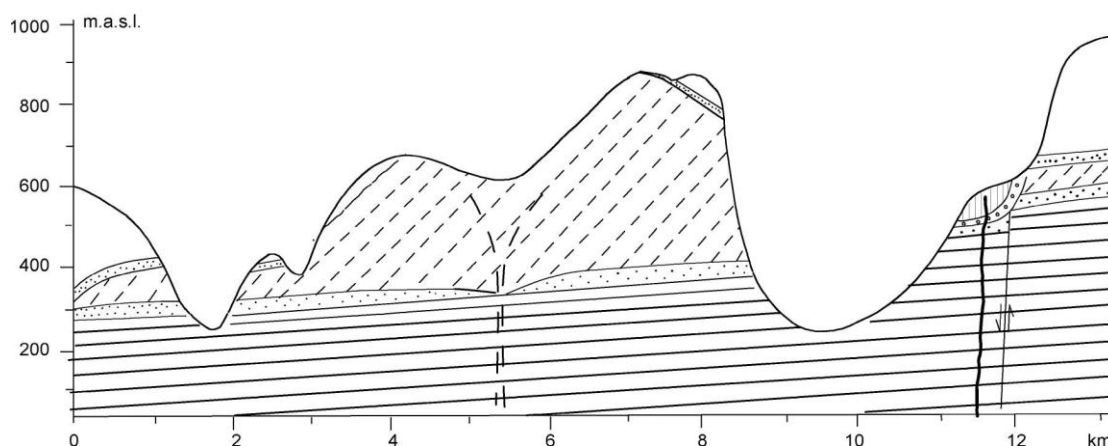
Fig. 3. The Skati Rhyolite Dome. Hatched =The Skati Dome (rhyolite lava). Dotted = Skati rhyolite tephra above and below the dome. Solid lines = Gently sloping pile below the dome. White = Tholeiite lava pile above the dome. The Austurdalur Pleistocene Volcano and its feeder dyke are shown on the right. The estimated location of the feeder dyke of the Skati Dome also shown. (Vertical/horizontal = 1/10)

The Skati rhyolite formation consists of two members, Skati lava and Skati tephra. This is an exceptionally thick and extensive monogenetic eruptive unit that has built up a mountain, the Skati Dome, rising at least 500 m above its base with the highest peaks in the present Skatastaðafjall area (Fig. 3). The eruption seems to have started with a powerful plinian phase. Viscous rhyolite lava was extruded along with the tephra emission. It covered the pyroclasts from the earliest phase of the eruption but the pyroclastic fall from the later phases continued on top of it. The crater site is not known but the conduit is most likely buried below the main body of Skatastaðafjall mountain. From there the lava flowed out in all directions from the crater on a sole of black basal breccia of obsidian aggregate. Towards the east it can be traced for over 8 km into the central Tinnárdalur valley; on the west side it disappears into the mountain Hofsfjall 7 km from the assumed crater. The south and north extensions are

Table 2. Thickness of the Skati tephra

Place	N°	W°	Thick-ness m	Comment
Merkidalur, north	65°21.0'	18°52.0'	9	
Merkidalur, south	65°20.5'	18°54.1'	12	
Brennigil	65°19.4'	18°52.1'	20	
Göngufjall	65°18.3'	18°47.7'	58	
Tinnárdalur, mouth	65°17.1'	18°44.6'	94	lava 78 m
West of Illagil	65°17.7'	18°41.0'	41	
Illagil	65°17.6'	18°39.6'	51	
Tinnárdalur, head	65°17.0'	18°36.6'	50	
Tinnárdalur, SE side	65°17.05'	18°42.7'	82	lava 103 m
Geldingaskarð	65°17.0'	18°43.0'	60	lava 55 m
Sandafjall	65°14.7'	18°42.5'	120	lava 60 m
Skatastaðafjall	65°17.4'	18°51.5'	100	lava 370 m
Skriðugil, at Hof	65°16.5'	19°01.8'	44	
Vesturdalur	65°16.4'	18°57.9'	10	lava 110 m
Goðdaladalur canyon	65°15.5'	19°04.4'	>20	
ODP-907A	69°15.0'	12°41.9'	0.018	Deep sea floor

not known but the width of the lava is at least 6 km. The maximal thickness is found to be 460 m but it was originally greater as the highest peaks have been eroded away.

The tephra layer is extremely voluminous and indicates a large and explosive plinian eruption. The thickness of the airborne tephra has been measured in 15 sections in the area (Table 2) and the isolines in Fig. 7 are based on these measurements. It must, however, be kept in mind that a map based on this material is not as reliable as maps of Holocene tephra sectors that are founded on tens or hundreds of soil sections, such as those of the Icelandic Hekla layers (Larsen and Thorarinsson 1977). The direction of the main axis of the tephra sector is towards the east, indicating a westerly wind during the main phase of the eruption. Farther east (and higher up) the wind direction might have been more southerly and southwesterly, curving the tephra sector towards the north.

The total volume of the Skati rhyolite can be estimated using the figures given above. According to geological mapping, the area of the dome is over 80 km^2 (Fig. 1) and if the mean thickness is 100 m then the lava volume is 8 km^3. Furthermore, gross estimates indicate that the volume of the Skati tephra corresponds to at least 10 km^3 of dense rock making the total volume equivalent to 18 km^3 of acid magma. This is far the largest monogenetic rhyolite formation yet reported in Iceland and an order of magnitude larger than any Holocene plinian eruption in the country (Simkin 1994). Compared with well-known plinian eruptions of historical times, it is larger than Krakatau 1883 (1.2 km^3) but smaller than Tambora 1815 (80 km^3) (Pyle 2000).
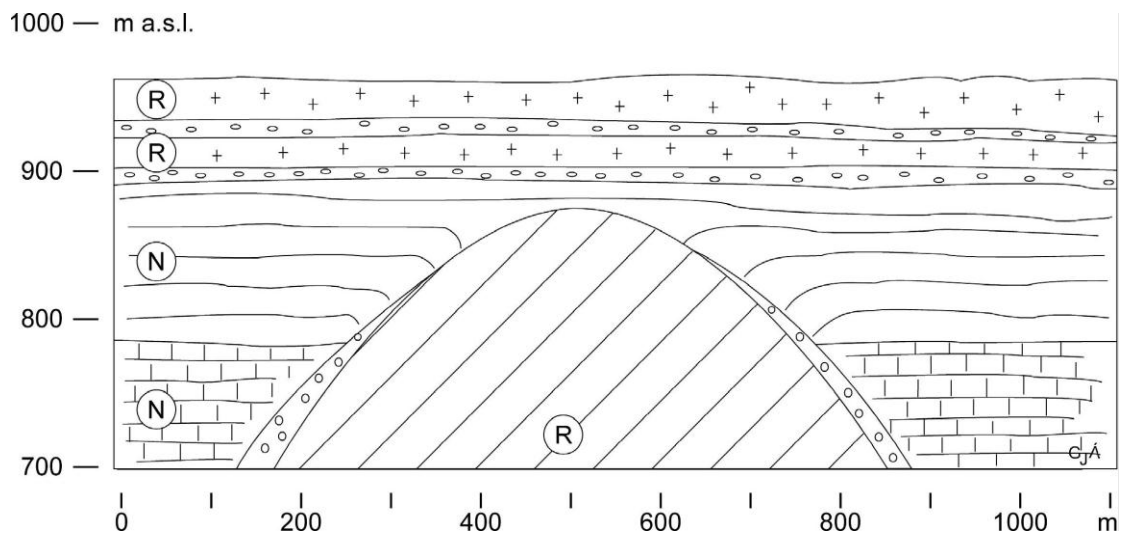
Fig 4. The Hvítárdalir dome with fossil screes at its sides buried in younger lavas. Fossá porphyritic basalt with intercalated sediments is at the top. Nýibær andesite in the middle. Nýibær tholeiites at the base. N and R show the polarity.

The Hvítárdalir Dome. In the narrow tributary valley Hvítárdalir a rhyolite dome, or ridge, is buried in the lava pile (Fig. 4). An excellent cross section can be seen in the steep slopes on both sides of the valley. It is at least 100 m thick, but the base of the dome is unexposed below the bottom of the valley. The diameter of the dome along the river is 750 m. Chemical analyses show close petrological relations with the Tinná rhyolite (Table 3). A pronounced sedimentary layer, a fossil scree, is situated between the dome and the surrounding lavas banking up against it. The dome has reverse

polarity and might be of similar age to the Skati Dome and a member of that formation.

The *Tinná andesite formation.* After the great eruption of the Tinná rhyolite, volcanism continued with production of intermediate lavas. Several andesite lavas, 20-70 m thick, and a few acid lavas that pile up against the southern slopes of the Skati Dome have been mapped. Their origins are somewhere south and southeast of the dome; thus, the centre of volcanic activity seems to have migrated in that direction immediately after the Skati eruption

The *Tinná tholeiite formation* is a pile of thin-layered lavas with thin interbeds covering the andesite lavas, banking up against the Skati rhyolite dome and encircling it. The maximal thickness of the pile is 160 m. The upper limit of these layers is marked by a shift of the magnetic field from reverse (R) to normal (N). The production rate of the tholeiites seems to have been high in the light of the thin interbeds and the magnetic reversal field measurements show how the strength of the field declines and increases again after the shift recording a polarity transition zone. Magnetic transition zones are believed to have a duration of only 1,000-8,000 years, so if they are recorded in a lava pile, the eruptions must have been frequent (Merrill and McFadden 1999, Coe et al. 2000, Pedersen et al. 2002).

The *Keldudalur dacite formation.* Keldudalur is a tributary valley in upper Austurdalur. Acid rocks are found in several places near its mouth and in the valley itself, all of which are thought to be exposures of a single formation. The rock type is dacite (68% SiO_2, see Table 19 in the Appendix). It forms a ridge or a row of two or more volcanic domes trending north-south. This is partly covered by loose scree and till so its base and contacts with the adjacent rocks have not been observed. Its highest part reaches 600 m a.s.l. The visible size is 3 x 2 km and its maximal thickness is at least 200 m. Assuming an elliptical shape and a mean thickness of 100 m, the volume of the ridge is 0.5 km^3.

The location of this unit inside the geological column is not clear and the Keldudalur dacite formation might even represent an intrusion. Further investigations are needed to determine this.

The *Nýibær tholeiite formation* is very similar to the Tinná tholeiite in all respects except polarity. From a petrological point of view, they would be regarded as the same member. These lavas finally immersed the Skati Dome totally. Remnants of the Nýjibær tholeiite lava cap can still be seen near the summit of the former acid dome in the mountain Skatastaðafjall.

The *Nýibær andesite* and the *Nýibær rhyolite formations*. At the end of the Nýibær period, a intermediate phase of acidic volcanic activity started up in the area, forming the Nýibær andesite and the Nýibær rhyolite formations. In the Vesturdalur valley there is a hill of andesitic hyaloclastite and fragmental basalt lavas with remnants of pseudo-craters. This might indicate an eruption inside a lake and high groundwater level and wetlands. The acidic rock consists mostly of lavas; tephra layers have not been reported. Chemical analysis shows that at least some of the lavas are in fact dacite, with 65% SiO_2 (see the Appendix). Tholeiite lavas are also found in this pile, the thickness of which reaches 180 m. Several cone sheets (cone dykes) indicate caldera formation during the final stage of the volcano, although no distinct caldera

fault has been found. The location and form of the caldera shown on the map (Fig. 1) is manly based on the existence of cone sheets near the Fossá river.

The upper limit of the Nýibær formations is in many places marked by the major Skagafjörður unconformity. The Nýibær acid and intermediate lavas are the youngest known formations of the Tinná Central Volcano but it cannot be stated that they indicate the extinction of the volcano. Still younger formations have been eroded away as seems to be indicated by cone sheets and acid dykes cutting the topmost layers of the Nýibær formations below the unconformity. Major caldera faults along with possible surface formations, such as screes and lacustrine sediments or other remains of a caldera lake, have not been located.

The centre of the activity seems to have migrated towards the southwest in the course of time. It was initiated near Ábær in Austurdalur (fig 1), after that it migrated to Skatastaðafjall, then farther south where the Tinná andesite were erupted and finally ended near Fossá in Vesturdalur, 15 km southwest of Ábær.

The *Fossárdalur group*. This group overlies the Tinná Group. It is dominated by porphyritic basalt lavas and prominent sandstone layers. It is about 100 m thick and forms a good marker horizon different from the adjacent formations, both in rock type and polarity. The group is found in the mountains in the upper part of the Austurdalur valley. It is assumed to have been erupted from a remote volcanic system after the extinction of the Tinná Central Volcano. It encircled the volcano but its highest peaks were perhaps never totally immersed. At any rate, the volcano seems to have inflienced the erosion of the Austurdalur and Vesturdalur valleys and their location at its sides and thus laid down the dominant features of the recent landscape (see Article 3 in this thesis).

Intrusions

Intrusions and dykes related to the Tinná Central Volcano are not conspicuous. The most common dykes in the area are basaltic ones, ranging between 1 and 5 m in thickness, but they can reach up to 40 m in thickness. The average thickness is around 4 m. These dykes penetrate the strata pile below the volcano and many of them cut the formations of the volcano itself, though none are seen to be connected to a distinct volcanic formation. Acid dykes are seen in Vesturdalur only near the river Fossá. As a general rule the dyke intensity becomes higher with increasing depth; they are rare at the highest exposed levels but can occupy 10-20% of the rock volume in dyke swarms at low levels. In the Skagafjörður Valleys the dyke intensity is much lower, i.e. about 7% at the maximum. Dyke swarms, which are often associated with magma chambers and central volcanoes in Iceland (Guðmundsson 2000), have not been detected in the Skagafjörður Valleys.

Intrusions other than dykes are small and rare. A few cone sheets from the last episode of the volcano have been found near river Fossá in Vesturdalur, as has already been mentioned. The mutual reason for the lack of intrusions and dyke swarms is that the volcano is so moderately eroded, and its roots are located deep below the valleys. This is also indicated by a low degree of alteration and zeolitisation. The Tinná Group is mostly situated inside the uppermost alteration zones i.e. the zeolite free zone and the chabasite-thomsonite zone, showing that the group was never buried below a thick pile of more recent lavas (Walker 1960). No signs of geothermal fields or high-temperature areas have been found in the vicinity of the volcano. Such areas are associated with most central volcanoes in Iceland

(Sæmundsson 1979). In this case there might be an unexposed high-temperature area below the Pleistocene formations in the southern part of the Tinná Volcano.
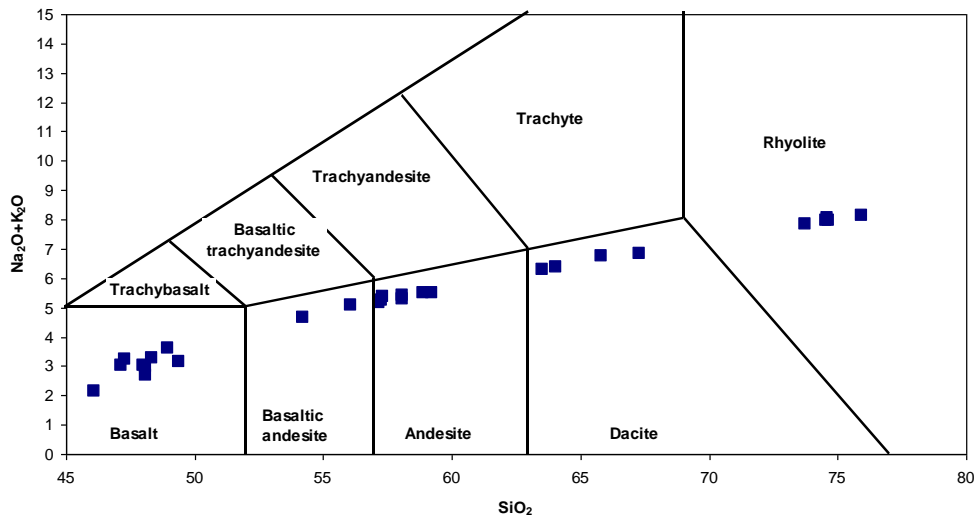
Fig. 5. Alkali-silica diagram (Le Maitre et al. 1989). The data is based on chemical analyses of rock samples from the stratigraphic units of the Tinná Group.

Geochemistry

About 30 analyses of rock samples from all the main stratigraphic units of the Tinná Central Volcano have been carried out (see Appendix, Table 19). Fig. 5 shows an alkali-silica diagram of the samples according to international standards (IUGS) (Le Maitre et al. 1989, Fig. B.13). The conclusion of the analysis is that basic, andesitic and acidic rocks from the volcano follow a tholeiite trend. The trend, appearing here, is close to the one that has been found for the Þingmúli Central Volcano in East Iceland and is often used for comparison (Carmichael 1964, Hjartarson et al. 1997, Hards et al. 2000). The main rhyolite formation of the Tinná Central Volcano is very high in silica, or up to 75% SiO_2. The geochemical data indicate a volcano from an axial rift zone and not an off-rift volcano (Lacasse and Grabe-Schönberg 2001).

Size and Age

The total volume of the Tinná Central Volcano can be roughly estimated. Its base at the bottom of the Ágúll Dome is about 30 km wide and the average thickness 300 m. If it is assumed to be semicircular, its area is 700 km^2. Hence the volume would be 0.3 km x 700 km^2 = 210 km^3. Here the 300 m thick Fjóslækur basaltic formation below the volcano is not taken into account. For comparison, the Breiðdalur Central Volcano, East Iceland, has been estimated to be 400 km^3 (Walker 1963). (If the rhyolites of Torfufell belong to it, it would be ~500 km^3 and among the greatest ones known in Iceland).

The age of the Tinná Central Volcano is based on an Ar/Ar-date from the Skati rhyolite dome (see Article 5 in this thesis). This gives 5.212 +/- 0.016 Ma. There seems to be some disharmony between this age and the magnetic polarity. The polarity of the rock is reverse but at 5.212 Ma the Earth's magnetic field was normal, representing the Thverá subchron (C3n4n), the lowest normal interval of the Gilbert polarity chron. Before the Thverá subchron a rather long reverse subchron occurred, representing the lowest part of the Gilbert polarity chron, C3r, 5.89 – 5.23 Ma (Cande and Kent 1995). The Skati Dome is assumed to belong to the central or the lower part of this subchron (Fig. 8) and its Ar/Ar-age is therefore slightly too low. A more credible age would be 5.5 Ma.

The lifetime of an ordinary central volcano is 0.5 – 1 Ma (Guðmundsson 2000). After that it cools down and is buried by younger lavas while it drifts out of the active volcanic zone. The Tinná volcano seems to be typical in this respect, since it remained active for at least three polarity subchrons (C3An.1n – C3r – C3n.4n), from c. 6 Ma to 5 Ma. This was the end of the Miocene and beginning of the Pliocene (Harland 1989).

Distant correlation

The Skati tephra might be recognizable in ODP cores from the ocean floor around Iceland. Lacasse and Garbe-Schönberg (2001) studied the explosive volcanism in Iceland and the Jan Mayen area during the last 6 million years from tephra layers found in the deep-sea cores. They recreated a composite marine tephra record based on data from ODP sites 907, 985, 919, 983, 984 and from the sites SU9029 and SU9032 (Fig 6). They found out that 90% of all identified tephra layers were recognized in ODP sites 907 and 985, downwind from Iceland N and NE off the coast. Almost a 100% recovery of well-preserved sediments was achieved there. The lowest part of their record is entirely based on these two sites. It is highly likely that the Skati tephra will be found there. The main attributes of this ash layer should be the following:

> Discrete layer
> Age around 5.2 Ma
> Colourless glass shards
> Silica content around 75% SiO_2 (on a water-free basis)
> Reverse polarity (in the centre or lower part of a polarity zone)

Four tephra layers from ODP-core 907A fulfil these requirements (indicated as Z, AA, AB and AC in Lacasse and Garbe-Schönberg 2001, Table 2). Chemical analysis indicates that one of them is alkaline and originates in an off-rift volcanic zone in Iceland. It can therefore be eliminated. Two others are near to the upper boundary of the Thverá subchron, which makes them hard to correlate with the Skati tephra. The only layer left is the oldest one in the composite core, indicated as AC. If the Skati tephra is to be found in the data collection at all, this is the best candidate. Lacasse and Garbe-Schönberg (2001) give it a special attention saying that its dispersal, thickness and grain size clearly indicate that it was derived from one of the largest explosive eruptions that ever occurred in the Neogene rift zones of Iceland.

Comparing the chemistry (Table 3) it must be kept in mind that the ODP-samples are made of a few glass fragments from a tephra layer that were analysed using a microprobe but the Skagafjörður samples are of lava that was analysed by a conventional XRF technique (see Appendix). Nevertheless, comparison reveals similar composition in most of the major elements as well as in the trace elements and

does not indicate any difference that cannot be explained by the different analytical methods and sedimentary environment.

Table 3. Chemical composition of the Skati Dome and the tephra layer in ODP-907A. Major elements are given in weight %, trace elements in ppm

Element	ODP-907* AC	Skati Dome S-23	Skati Dome S-31	Skati Dome S-16	Skati Dome S-17	Hvítárdalir Dome S-11
SiO_2	71.49	74.60	75.95	73.75	74.65	74.56
TiO_2	0.14	0.142	0.100	0.166	0.122	0.146
Al_2O_3	12.18	13.17	12.87	12.86	13.12	13.27
Fe_2O_3	2.48	2.13	1.8	2.21	2.11	2.28
MnO	0.09	0.06	0.05	0.05	0.02	0.03
MgO	0.01	0.03	0	0.05	0.02	0.05
CaO	1.11	1.01	0.94	1,20	0.95	1.07
Na_2O	5.36	4.64	4.69	4.54	4.69	4.71
K_2O	2.38	3.40	3.42	3.30	3.26	3.24
P_2O_5		0.031	0.010	0.018	0.017	0.021
LOI		0.84	0.09	1.9	0.74	0.84
Tot.	94.99	100.15	99.98	100.15	99.8	100.32
BaO	508 (Ba)	713	712	671	685	675
Ce	162	138	103	149	141	130
Co	0.77	26	22	10	20	21
Cr_2O_3	12 (Cr)	< d/l	< d/l	18	< d/l	< d/l
Cu	23.5	19	19	32	11	52
Ni	6.90	< d/l	< d/l	3	< d/l	< d/l
Sc	1.55	< d/l	< d/l	< d/l	< d/l	< d/l
V		< d/l	< d/l	< d/l	< d/l	12
Zn	236	128	131	132	119	121
Ga	28.7	26	26	25	25	26
Nb	81.4	73.3	71.8	70.7	71.5	69.8
Pb	9.32	3.2	3.1	3.0	3.2	3.2
Rb	63.3	81	83	80	79	77
Sr	76.3	62	55	66	70	78
Th	8.15	5.5	4.9	5.9	5.6	5.4
U	2.47	< d/l	< d/l	< d/l	< d/l	< d/l
Y	126	113	133	129	131	123
Zr	403	253	233	268	289	324

*From Lacasse and Garbe-Schönberg 2001. d/l = detection limit.

Lacasse et al. (1996) have described this layer and given it the label AC. It is bimodal and the top differs from the bottom. The bottom is crystal-poor tephra of 100% colourless glass shards and can be considered as exclusively silicic. The top contains 9% feldspar, 3% clinopyroxene and up to 2% olivine. This sorting might have taken place as the ash particles settled through the 1,800 m deep water column. Normal grading is also observed between the bottom and top of the layer. It can also be interpreted as the result of extensive size fractionation in the water column. This difference has not been recognized inside in the Skati tephra in the Skagafjörður Valleys. ODP-site 907 is in the Artic Ocean NNE of Iceland, 550 km away from the Tinná Volcano. This distance has not changed much since the eruption. The holes (A, B and C) are 1,800 m below sea level and the tephra layer is at a depth of 85 m in the sediments. The age, magnetostratigraphic alignment within subchron C3r, its thickness (18 cm) and chemical composition, all favour the distant correlation between the deep-sea ash layer and the Skati tephra.
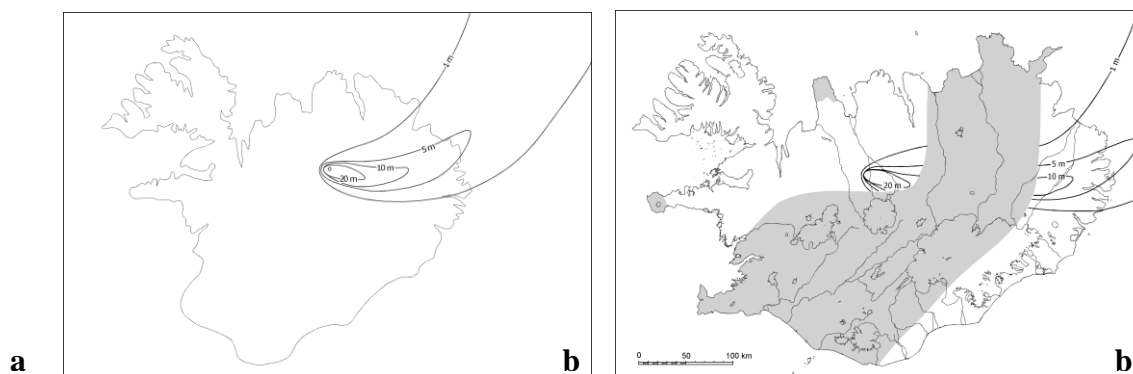
Fig. 6. The tephra sector close to the eruptive site. The innermost ellipse is the estimated isoline for the 20 m tephra thickness based on measurements. The outer isolines are hypothetical. a) The tephra sector is projected on the country after removing rocks younger than 5.5 Ma. b) The tephra sector is projected on the country as it looks like today. The shaded areas indicate rocks younger than 5.5 Ma. The regional distribution is illustrated in Fig. 7.
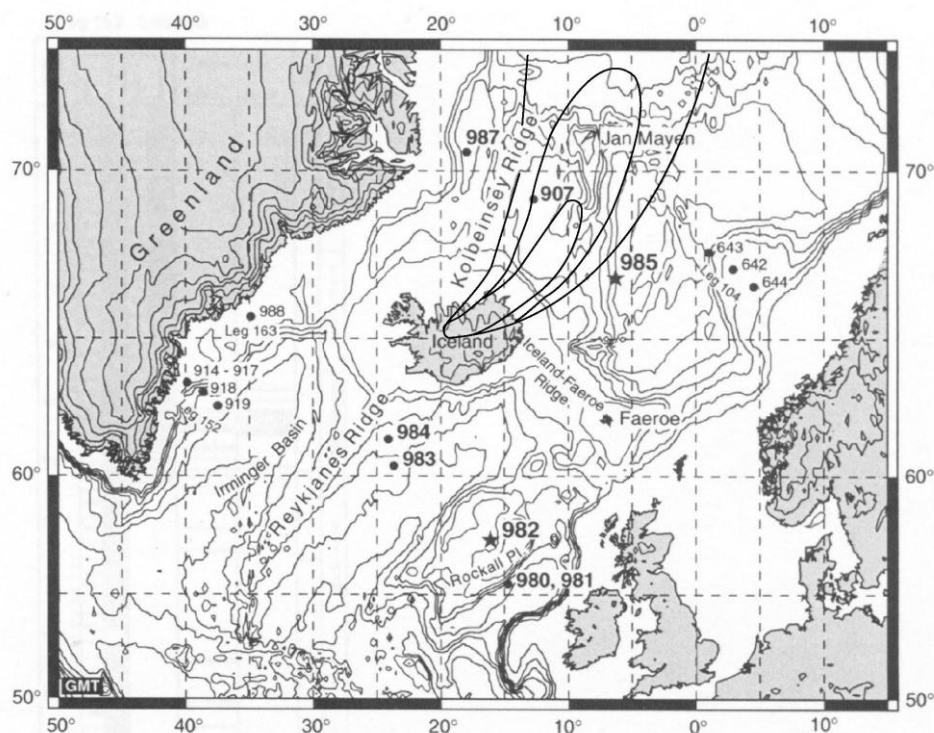
Fig. 7. Map of the central North Atlantic showing locations of ODP-sites and suggested distribution of the Tinná tephra. The alignment of the tephra sector is in fact unknown; here the most probable alignment is shown.

Volcanic history

The development of the Tinná Central Volcano and its eruptive history can be divided into three main phases, all experiencing rhyolitic volcanism at or shortly after its beginning (Fig. 8).

Phase1: The initial stage took place ca. 6 Ma. An acidic eruption started with the extrusion of the Ágúll rhyolite dome. This was at least 250 m high and 0.6 km^3 in volume. The stratigraphy of the Ábær gorge indicates nearby eruptive vents. The volcanism began with an explosive plinian phase and the accumulation of a thick and coarse tephra layer. During the plinian phase an acid lava was extruded. This was very viscous and piled up around the conduit and flowed slowly out over the light coloured tephra carpet, forming a layer of basal breccia and a massive rhyolit dome (or a ridge). The acidic eruption was followed by intense basaltic volcanism producing the Ábær tholeiites that finally covered the dome. The phase ended in a long period of quiescence during which the Tinná lignite sediment was formed. The lignite seams contain trunks and branches, indicating Tertiary woodlands.

Phase 2: The central phase started with the eruption of the large Tinná olivine tholeiite lava. The relationship between the lava and the central volcano is unclear. Shortly after the emission of the lava, violent explosive volcanism took place with an immense pyroclastic fall and the extrusion of a large new rhyolite dome, building up at least 500 m high mountain, the Skati Dome. The tephra is found both below and on top of the dome, indicating a simultaneous emission of ash and lava. The total volume of eruptives emitted was equivalent to at least 18 km^3 of magma. A thick and wide spread tephra sector was accumulated. Westerly winds directed the tephra towards the east but farther away (and higher up) the winds seem to have been more southerly, turning the tephra sector towards the north (Figs 6, 7). The Skati eruption was followed up by intermediate volcanism and later on the accumulation of the thin-layered lava pile of the Tinná and Nýibær tholeiites which in the end immersed the Skati Dome.

Phase 3: The final stage appears as an alternating volcanic phase producing acid and intermediate lavas with basaltic eruptions in between. The centre of the activity was farther south than at the beginning, and a caldera was formed. This is assumed to have happened about 5 Ma. After that the volcano became extinct and drifted to the west, away from the active volcanic zone of North Iceland, while younger lavas piled up against it and covered it partly, though its highest summits always seem to have extended above the environment.

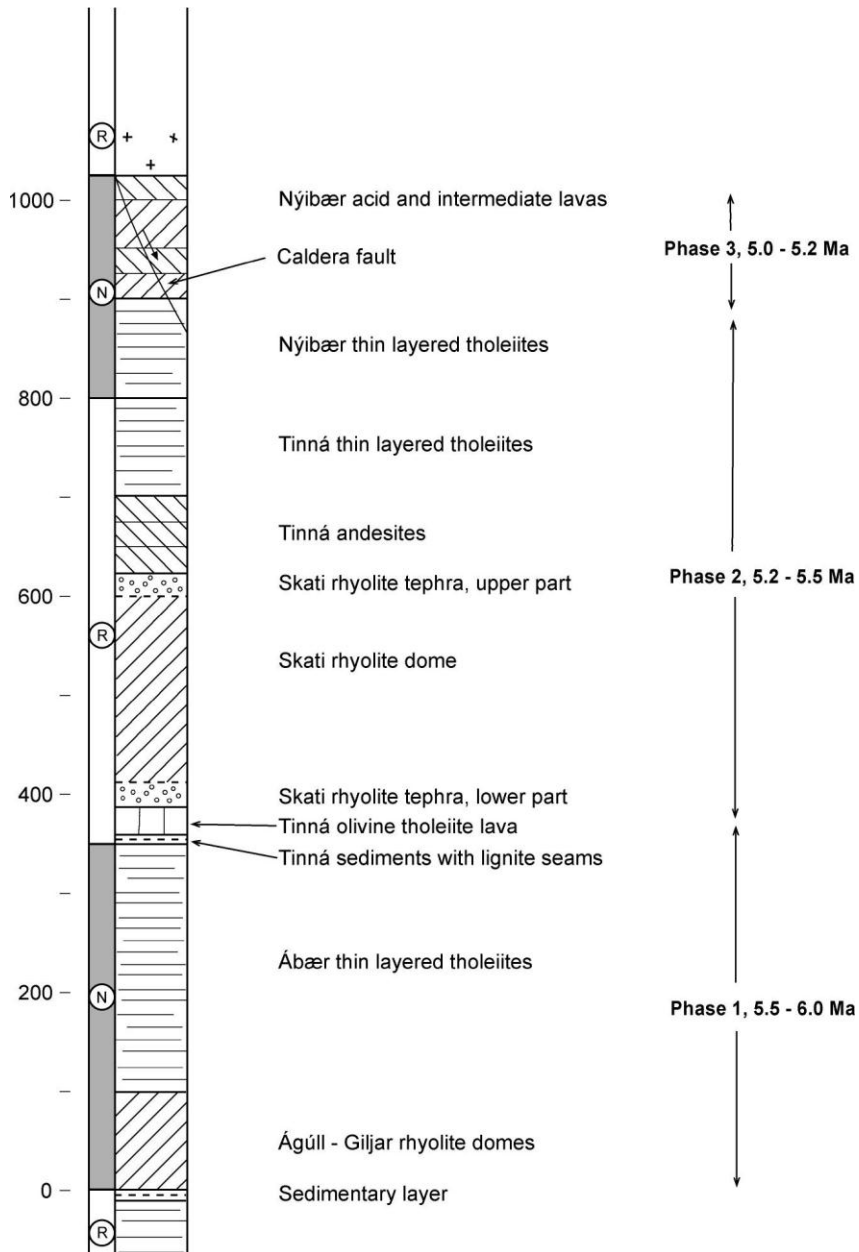*Fig. 8. Stratigraphy and history of the Tinná Central Volcano. A simplified geological column. The thickness of the formations is given in m on the left.*

Table 4. The acid domes of the Tinná Volcano

Name	Thick-ness m	Volume km^3	SiO$_2$ %
Ágúll dome	100	0.5	
Skati dome	100	8	75
Hvítárdalir dome	75	< 0.1	75
Keldudalur dome	100	0.5	68

Discussion

Among the 40-50 known Neogene central volcanoes in Iceland the Tinná Volcano is distinct in some respects. This is not because of its overall size, the duration of its activity or the great variety of rock types formed in periods of explosive acid eruptions in between more quite periods of basic and intermediate volcanism: all these topics seem fairly representative for central volcanoes in general (Walker 1963, Jóhannesson 1975, Franzson 1978, Friðleifsson 1983). It is the volume of the monogenetic Skati formation that makes it unique, especially the Skati lava dome. Rhyolite domes such as those described here (Table 4) are rare in Iceland. Only a few domes are known in the Neogene areas. G.P.L. Walker (1963) i.e. described small domes in the Breiðdalur Central Volcano. In the Quaternary areas the Hágöngur mountains in Central Iceland are the most prominent examples, but they might be subglacial formations. No such dome is known to have been formed in the Holocene time. The Skati lava dome is of an order of magnitude larger than all other rhyolite domes in the country. The reason for this is unknown. The magma chamber might have been located at an unusually shallow depth in the crust, allowing acid lavas an easy access to the surface. The low degree of alteration and lack of a geothermal aureole around the volcano support this suggestion. But the absence of a caldera formation along with the eruption and the absence of a high-temperature area seem to argue against it. A shallow magma chamber, however, with a thin roof might be harder to identify than a deep-seated one. It might have collapsed contemporarily with the extrusion of the dome, leaving the caldera unexposed below it. (The caldera near the Fossá river (Fig. 1) was formed later, during the final stages of the volcano).

  Geothermal activity and high-temperature areas are commonly associated with most of the rift-related central volcanoes of Iceland (Sæmundsson 1979, Guðmundsson 2000). No indications of such a high-temperature area are found near the Tinná Volcano. Possibly it is still buried below the Pleistocene volcanic formation at the southwest border of the volcano.

Conclusions

The Tinná Central Volcano in the Skagafjörður Valleys belongs to the Neogene succession of North Iceland. It originated inside the North Iceland Volcanic Zone and was active during the period 6 − 5 Ma. The total volume of the volcano is at least 210 km^3, comprising four rhyolite lava domes (Table 4). Its volcanic products belong to the tholeiitic rock series. The major eruption of the Tinná volcano, the Skati eruption 5.5 Ma producing 18 km^3 DRE, with a high fraction of lava, is outstanding among Icelandic acid eruptions. The tephra layer is correlated to an acid ash layer found at ODP site 907, 500 km NNE off Iceland's coast and can serve as an important marker horizon for the late Miocene in the deep-sea sediments. A collapse caldera was formed during the final stage of the volcano. The Tinná Volcano was not a stratovolcano but rather an irregular massif of heaps and domes without any major summit crater. At times it rose high above the environment but while it was being formed, flood-basalts were issuing from fissures throughout its surroundings. Finally it was buried, or nearly buried, by them as the volcano drifted westwards, away from the active rift zone of North Iceland.

Acknowledgements

References

Cande SC, Kent DV (1995) Revised calibration of the geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. J Geophys Res 100:6093-6095

Carmichael ISE 1964: The petrology of Thingmúli, a Tertiary volcano in E Iceland. Journal of Petrology 5, 435-460

Coe RS, Hongre L, Glatzmaier GA (2000) An examination of simulated geomagnetic reversals from a paleomagnetic perspective. Philos. Trans. Royal Soc. London, A 358, 1141-1170

Fink JH, Anderson SW (2000) Lava Domes and Coulees. In: Sigurdsson H (2000) Encyclopedia of Volcanoes. Academic Press, London, pp 307-319

Franzson H (1978) Structure and petrochemistry of the Hafnarfjall – Skarðsheiði central volcano and the surrounding basalt succession, W-Iceland. PhD-thesis, University of Edinburgh, pp 275

Friðleifsson GÓ (1983) The geology and alteration history of the Geitafell central volcano, southeast Iceland. PhD-thesis, University of Edinburgh, pp 371

Guðmundsson Á (2000) Dynamics of Volcanic Systems in Iceland: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge System. Annual Review of Earth and Planetary Sciences 28:107-140

Guðmundsson Á (1991) Jökulsár í Skagafirði. Geological map, 1:50.000. Unpublished manuscript. Orkustofnun, Reykjavík

Hallgrímsson J (1989) Ritverk Jónasar Hallgrímssonar I-IV. Svart á hvítu, Reykjavík

Hards VL, Kempton PD, Thompson RN, Greenwood PB 2000. The magmatic evolution of the Snæfell volcanic centre; an example of volcanism during incipient rifting in Iceland. J Volcanol Geotherm Res 99, 97-121

Harðarson BA, Guðmundsson Á (1986) Stafnsvatnavirkjun. Mannvirkjajarðfræði. OS-86039/VOD-14 B, pp 1-63 + maps

Harland WB, Armstrong RL, Cox AV, Craig LE, Smith AG, Smith DG (1990) Geologic Time Scale, 1989 edition, Cambridge University Press, Cambridge, pp 1-263

Hjartarson Á, Friðleifsson GÓ, Hafstað ÞH (1997) Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020 Orkustofnun, Reykjavík, pp 1-55 + maps

Jóhannesson H (1975). Structure and petrochemistry of the Reykjadalur central volcano and the surrounding areas, Midwest Iceland. PhD-thesis, University of Durham, pp 273

Jóhannesson H (1991) Yfirlit um jarðfræði Tröllaskaga (Miðskaga). Árbók FÍ 1991, pp 39-56

Jóhannesson H, Sæmundsson K (1998) Geological map of Iceland 1:500.000. Tectonics. Icelandic Institute of Natural History, Reykjavik

Jónsson B, Egilson D, Zóphóníasson S (1977) Villinganes. Mannvirkjajarðfræði. OS-ROD 7709, Orkustofnun, Reykjavík, pp 1-19

Kaldal I, Víkingsson S (1978) Jökulsár í Skagafirði I. Jarðfræði. OS-ROD 7805, Orkustofnun, Reykjavík, pp 1-33 + maps

Kaldal I, Víkingsson S (1979) Jökulsár í Skagafirði II. Jarðgrunnskort. OS-79044/ROD-16. (Two maps in an album), Orkustofnun, Reykjavík

Kjartansson G (1965) Geological map of Central Iceland [Jarðfræðikort af Íslandi. Mið-Ísland]. Náttúrugripasafn Íslands, Reykjavík

Le Bas et al. (1986)

Lacasse C, Paterne M, Werner R, Wallabre-Adams, H-J, Sigurðsson H, Carey S,, Pinte G (1996) Geochemistry and origin of Pliocene and Pleistocene ash layers from the Icelandic plateau, site 907. pp 309-331. In: Thiede et al. 1996: Proceedings of the Ocean Drilling Program, Scientific Results, Vol. 151, 309-331

Lacasse C, Grabe-Schönberg C-D (2001) Explosive silicic volcanism in Iceland and Jan Mayen area during the last 6 Ma: sources and timing of major eruptions. J Volcanol Geotherm Res 107:113-147

Le Maitre RW, (Editor), Bateman P, Dudek A, Keller J, Lameyre J, Le Bas MJ, Sabine A., Schmid R, Sörensen H, Streckeisein A, Wolley R, W., Zanettin B (1989) A Classification of Igneous Rocks and Glossary of Terms, Recommendations of the International Union of Geological Sciences, Subcommission on the Systematics of Igneous Rocks, pp 193

Magnússon G (1980) Um nokkra surtarbrandsstaði í Vesturdal í Skagafirði. Týli 10:7-12

Magnússon G (1981) Um nokkra surtarbrandsstaði í Austurdal. Týli 11:53-58

Merrill RT, McFadden LP (1999) Geomagnetic polarity transitions. Reviews of Geophysics 37, 201-226

Ólafsson E (1943) Ferðabók Eggerts Ólafssonar og Bjarna Pálssonar um ferðir þeirra á Íslandi 1752-1757, vol. II. Reykjavík, pp 1-317

Pálsson S (1983) Ferðabók Sveins Pálssonar. Dagbækur og ritgerðir 1791-1797. Örn og Örlygur, Reykjavík, pp 1-813

Pedersen A K, Larsen L M, Riisager P, Dueholm K S (2002) Rates of Volcanic deposition, facies changes and movements in a dynamic basin: the Nuussuaq Basin, West Greenland, around the C27n-C26r transition. In: Jolley, D. W. and Bell, B. R. (eds.) The North Atlantic Igneous Province: Stratigraphy, Tectonic, Volcanic and Magmatic Processes. Geological Society of London, Special publications, 197:1-25

Pyle DM (2000) Sizes of Volcanic Eruptions. In: Sigurdsson H (2000) Encyclopedia of Volcanoes. Academic Press, London, pp 263-269

Simkin T, Siebert L (1994) Volcanoes of the World, 2nd edition: Geoscience Press in association with the Smithsonian Institution Global Volcanism Program, Tucson AZ, pp 1-368

Sæmundsson K, Kristjánsson L, McDougall I, Watkins ND (1980) K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. J Geophys Res 85:3628-3646

Walker, GPL (1960) Zeolite zones and dike distribution in relation to the structure of the basalts of eastern Iceland. Journal of Geology 68, 515-528.

Walker GPL (1963) Breiddalur Central Volcano, eastern Iceland. Quarterly J Geol Soc London, 119:29-63

THE PLIOCENE VALLEYS OF THE SKAGAFJÖRÐUR DISTRICT, NORTH ICELAND

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Abstract

Unique geological circumstances in the Skagafjörður district, North Iceland, make it possible to reconstruct the evolution of the topography step by step. This history spans 9 million years, from the origin of the area near the diverging boundaries of the North American and Eurasian crustal plates, through its drift out of the accretion zone and the erosional stages leading to the present landscape. It is concluded that already at the end of the Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had been carved out in the primordial lava plain of the Skagafjörður district. The erosion level in the principal valley was lower than today and the mountains on each side were as high or even higher than they are now. The main tributary valleys were deeply eroded and many of the small tributaries of the second order also existed. The glacial erosion of the Pleistocene is of less importance than the Neogene erosion. By analogy it can be assumed that all the main fjords and valleys in the Neogene regions of Iceland are old and already existed in the Pliocene and some even in Miocene times.

Introduction

The aim of this article is to describe the development of the topography of central North Iceland, establish the main stages in the evolutionary history, reconstruct the valley generations in the research area and evaluate the contribution of different erosion agents to this development. The research area is a volcanic region, less than 10 million years old, and its history can be outlined from its formation in the volcanic belt at the plate boundaries between the Eurasian and North American crustal plates and up to the most recent erosional stages where the current landscape, with wide fjords and deep valleys, was carved out. Special geological circumstances make it possible to reveal this history step by step. Landscape development in Iceland has been investigated to a very limited extent, but two different opinions have appeared. The first suggests that mature fjords and deep valleys already existed in Iceland before the onset of the Pleistocene glaciation (T. Einarsson 1958, 1959, 1959b, 1961, 1972). The second presumes that the Neogene landscape of Iceland was characterized by an immature volcanic lava plain with even topography and shallow valleys, and that the recent landscape was created mostly by Pleistocene glacial erosion. (Th. Einarsson 1968, 1991).
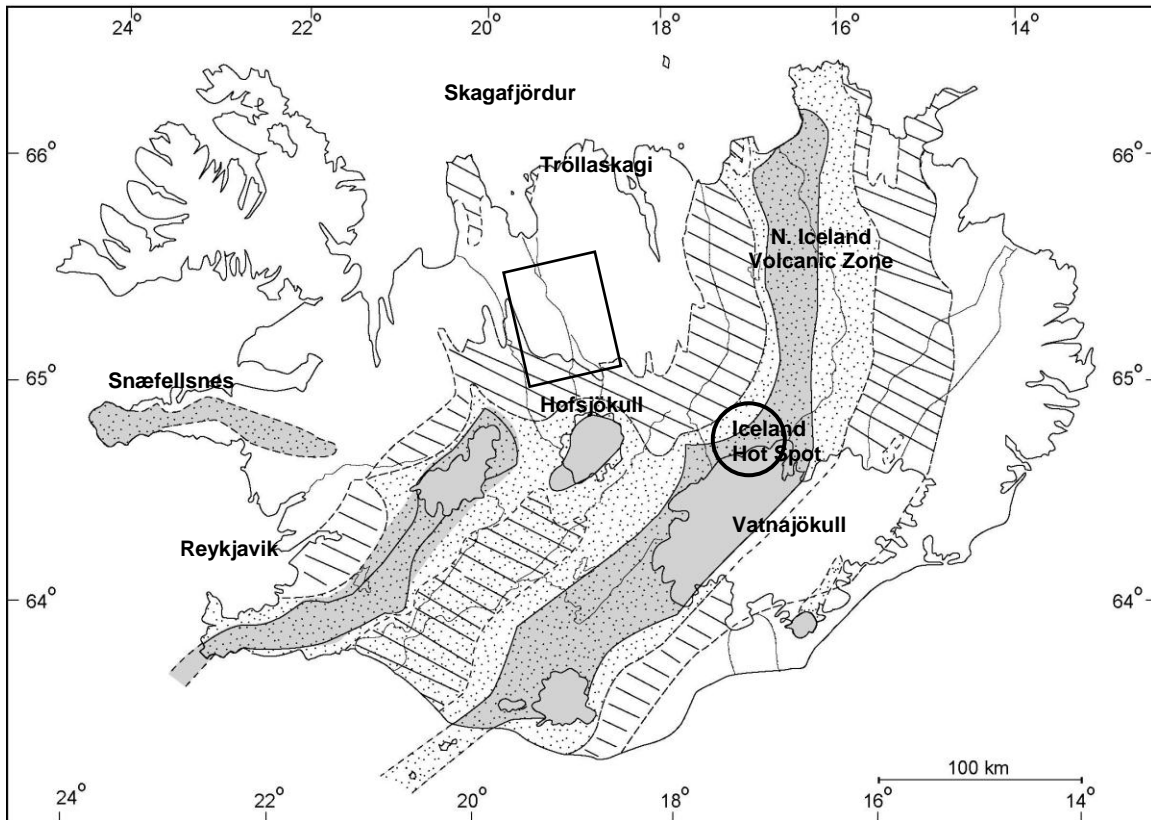
Fig. 1. Geological map of Iceland. The research area is indicated by a frame. Shaded = The Neovolcanic zone. Dotted =Early Pleistocene rocks. Oblique lines = Plio – Pleistocene rocks (0.8-3.2 Ma)

The first full-scale glaciations in Iceland with glaciers occupying the main valleys and fjords are assumed to have occurred after 2.5 million years ago (Geirsdóttir 1990). Prior to that time smaller, local glaciations had overridden the central highlands of Iceland several times.

The research area

The Skagafjörður region is a mountainous terrain with the highest peaks reaching over 1,000 m. The research area extends from the central Skagafjörður district and up the deep valleys further upstream where the large glacial rivers Austari-Jökulsá and Vestari-Jökulsá have eroded canyons into the bedrock (Figs. 1, 2). Hofsjökull, one of the largest ice caps in the country, forms the southern border of the area. This covers an active central volcano with an ice-filled caldera and its summit reaches over 1,800 m a.s.l.(Björnsson 1998). The North Icelandic Volcanic Belt is some 100 km to the east and the center of the Iceland Hot Spot is located somewhat farther to the southeast (Sæmundsson 1979) (Fig. 1). The stratigraphy can be divided into three main groups: (1) Neogene volcanic succession, 9-5 Ma. (2) Unconformity with overlying sedimentary succession, 5-2 Ma. (3) Pleistocene volcanic succession, 2-0.5 Ma (Hjartarson et al. 1997). The reason why the Skagafjörður Valleys are favorable for investigating the history of the landscape is based on the fact that after the primordial formation of the area during the Neogene period and considerable weathering and erosion of an initial valley

system, temporary volcanism started and an early Pliocene succession of lavas and sediments filled up the valleys, covering and thus preserving the ancient topography (Einarsson 1959, 1962, and Article 4 in this thesis). After the late Pleistocene erosion the paleo-landscape has been exposed and can be observed and dated in several places.

Paleo-landscape preserved below widespread more recent formations, as in Skagafjörður, is known in a few places within the Icelandic lava pile. The best-known localities are on the lignite-bearing sedimentary horizons in W and NW Iceland (Sigurðsson and Sæmundsson 1984); below the unconformities bordering the North Iceland spreading zone on each side (Sæmundsson 1979, Jancin 1984), and at the unconformity below the Snæfellsnes Pleistocene/Holocene volcanic formations (Sigurðsson 1970). A mature paleo-landscape is only seen at two of these horizons, i.e. in Skagafjörður and Snæfellsnes (Fig. 1).
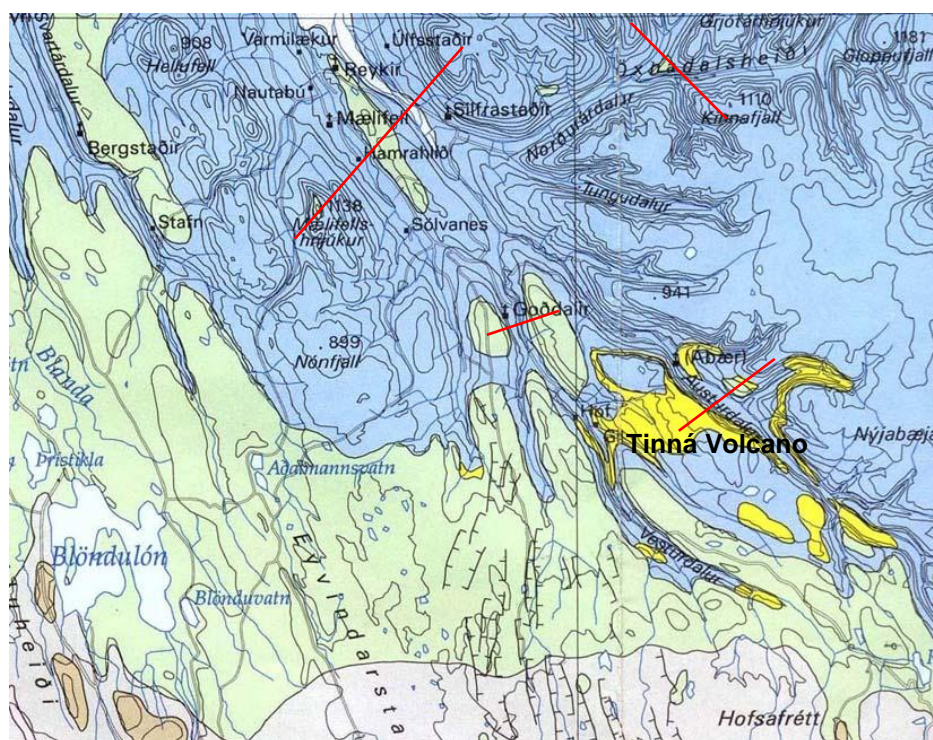
Fig. 2. Main features of the geology of the Skagafjörður Valleys. Red lines locate the sections. Blue = Neogene succession. Green = Plio-Pleistocene. Grey = Late Pleisto-cene. Yellow = rhyolites of the Tinná Central Volcano. (From Jóhannesson and Sæmundsson 1998).

Landscape on the Boundary of a Crustal Plate

The Skagafjörður Valley system is carved out of the western flank of the Tröllaskagi volcanic plateau, which is now one of the most mountainous regions in Iceland. The presently fairly flat and smooth mountain tops indicate the former existence of a continuous, extensive lava plain that has been deeply eroded and weathered (Fig. 3).

It is necessary to bear in mind that the area is located near the divergent boundaries of the Eurasian and North American crustal plates and has drifted out of the active volcanic belt at the plate boundaries according to Pálmason's (1980, 1981) model of crustal generation in Iceland. The evolution that will be discussed here is therefore the sculpturing of landscape just after crustal accretion at active plate boundaries. The accumulation of new eruptive material was a dominating factor while the area was inside the volcanic belt, but when it drifted out of it this accretion stopped and erosion, together with ablation of the young strata pile, took over. Similar conditions are rare on a world scale but though they may exist in the East African rift valley, between the African plate and the Somali sub-plate. There an early stage of plate separation and spreading on dry land can be observed, along with associated landscape evolution (Schlüter 1997). The erosion of the basaltic regions of Greenland might also in a certain way correspond to the Icelandic conditions (Japsen et al. 2002).

The first stage, the formation of the Neogene succession, lasted from 9 to 5 Ma according to radiometric dates of the bedrock (Sæmundsson et al. 1980 and Article 5 in this thesis). During the accumulation of the volcanic pile the Tinná Central Volcano was the major edifice of the landscape. Its activity started with a great rhyolite eruption at around 6 Ma, followed by intensive basaltic volcanism. Half a million years later a powerful plinian eruption took place along with extrusion of the large Skati rhyolite lava dome. This event has been described as one of the largest explosive eruptions in Iceland during the late Neogene (Article 2 in this thesis). It was followed by prolonged volcanism that ended in the formation of a caldera about 5 million years ago. Then the volcano rose to a height of perhaps 500 m above the surrounding lava plain all around. Later on, regional lavas surrounded the mountain and partly covered it but never buried it completely, and it seems to have been a controlling factor in the erosion and the formation of the initial valleys.

The peneplanation

The primordial lava plain drifted slowly out of the active volcanic zone. Accumulation of volcanic extrusives slowed down and stopped, and erosion took over (Pálmason 1980,1981), weathering and carving out a new landscape. This, the episode of peneplanation, might have occurred near the end of the Miocene and in the beginning of the Pliocene, around 5 million years ago. Originally, when the area was still inside the volcanic zone, and initially after it had drifted out of it, the landscape is believed to have been relatively flat with extensive lavafields and with crater rows and low ridges, as in the present-day volcanic zone. The major structures of the Tinná Volcano were mostly buried under younger formations (Article 2 in this thesis). The dominating tectonics in the area had a NNW-SSE orientation and the main crater rows, dykes, faults and fissures also lay in that direction (Jóhannesson and Sæmundsson 1998, see also the Appendix). Thus the present large-scale topography, the main fjords and valleys of central North Iceland, tend to be NNW-SSE oriented (Fig. 1).

The hydrology was probably similar to what it is in the present volcanic belts. Most of the precipitation penetrated into the ground forming groundwater streams, issuing in great spring areas on the borders of the lava plains (Hjartarson et al. 1980). Glaciers did not play a major role at this time (~5 Ma), if they existed at all. Erosion seems to have been

Fig. 3. Map of Tröllaskagi peninsula showing the remains of the Neogene plateau (black) and possible contour lines indicating the landscape of the peneplain. The plateau reaches 1,500 m sloping gently towards Bárðardalur and Skagafjörður. The thick dashed line borders the area where the ancient plateau is completely eroded away. The line A − B shows the location of the 80 km long cross-section in Fig. 4. (From Hjartarson 1973).
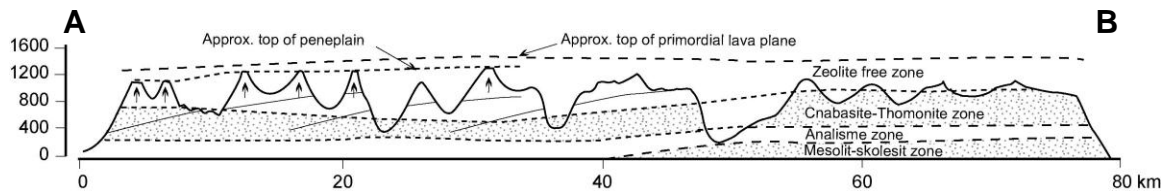
Fig. 4. A section across the Tröllaskagi peninsula. The alignment is shown in Fig. 3. The dip is towards the SW so the stratigraphy becomes older towards the right. The zeolite zones, the approximate top of the primordial lava plain and the hypothetical surface after the peneplanation are shown. The flat mountain-tops indicating the ancient plateau are marked with arrows. (Based on a drawing in Sæmundsson et al. 1980).

of low degree, perhaps mainly in the form of wind and chemical weathering. The remnants of this gently undulating and uniform landscape can still be seen in the fairly even top level of the mountains in the Tröllaskagi massif (Fig. 5).

It seems possible to map and reconstruct a continuous plateau across the southern part of the Tröllaskagi massif, from Eyjafjörður to Skagafjörður (Figs. 3 and 5). Both fjords appear in the paleolandscape, though in different ways. The peneplain slopes gently towards Skagafjörður, while Eyjafjörður is more like a rift zone extending into it. In the northern part of Tröllaskagi, especially at the mouth of Eyjafjörður fjord, the plateau is totally eroded away and the mountain-tops are all lower than the peneplain.

At their highest, the remnants of the plateau lie between 1,000 and 1,500 m a.s.l. There are two summits, reaching 1,450 m near the Skíðadalur valley and 1500 m near the town Akureyri. From this highest section the plateau slopes gently in all directions, towards the sea in the north and the Icelandic central highlands in the south, towards the Skagafjörður district in the west and the north Iceland Volcanic Belt in the east. The lowest-lying remains of the plateau are found near the northernmost shores of Skagafjörður at 300-400 m a.s.l. (Fig. 3).
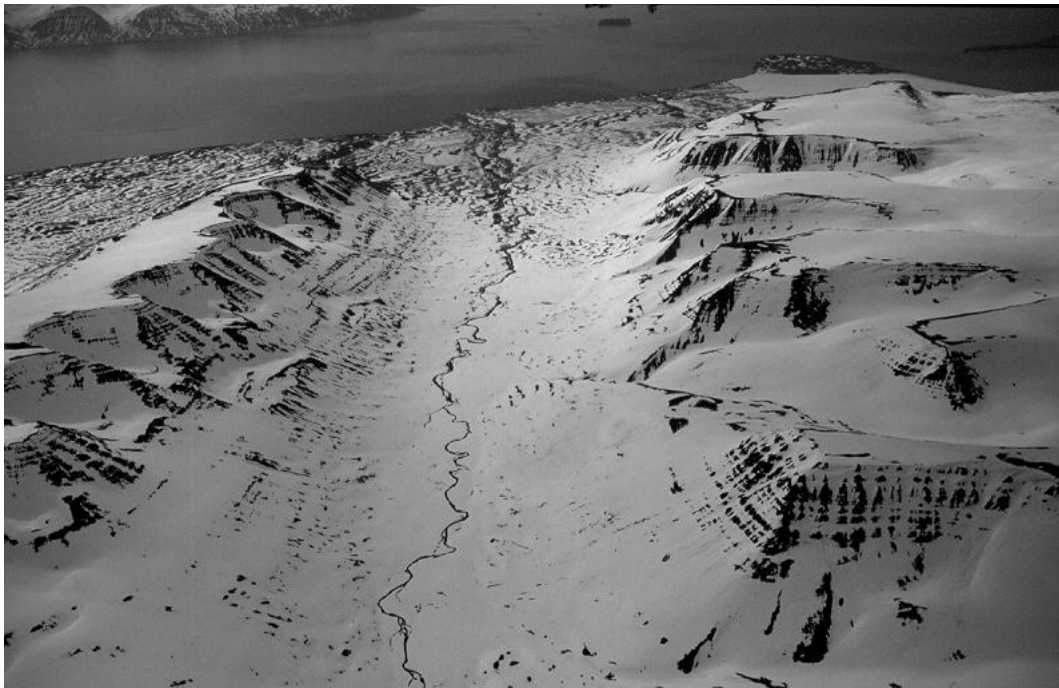
Fig. 5. The remains of the Tröllaskagi volcanic plateau are seen in the flat and even mountain-tops.

Originally the elevation of the plateau must have been lower, but later on, along with the erosion of the fjords and valleys, an appreciable uplift is believed to have taken place. The form of the plateau might indicate an uplift of 200-300 m due to isostasy. Sigmundsson has surmised that the uplift in Tröllaskagi resulted from erosion (Sigmundsson pers. com. See also Sigmundsson 1991). It has removed approximately 30% of the rock volume that is now above sea level. Accordingly the isostatic uplift might have reached 200 m near the center of the area, and 125 m at the mouth of Skagafjörður. These two estimates fit each other fairly closely.

The extent of the ablation of the primordial lava plain during peneplanation is hard to estimate. Alteration and zonal distribution of amygdale minerals (especially zeolites) in the lava pile have often been used to estimate the geothermal heating and pressure they were subjected to. The heat and pressure are proportional to the depth the layers reached and the geothermal gradient of the area. By this method it is possible to estimate how much of the original surface layers have been removed by the erosion (Walker 1960). In Tröllaskagi the ablation seems only to have been in the order of 100-200 m during the peneplanation (Fig. 4) (Sæmundsson et al. 1980). In the northernmost and oldest part of the section the remnants of the plateau are weathered away and the mountain-tops are believed to lie about 400 m below the level of the primordial lava plain.

The Valley System

With time the lava plain became less permeable because of the alteration of the rock. Permeability decreased from 10^{-2} m/s in recent lavas down to 10^{-6} m/s in Neogene basalts (Sigurðsson and Ingimarsson 1990). Accordingly, water erosion on the surface became more effective and the weathering rate increased. The period of peneplanation ceased when a distinct valley system began to erode into it. Before going further into the evolutionary history some sections across the main valleys of the district will be described, i.e. the principal valley of the district and the large tributary valleys Vesturdalur, Austurdalur and Norðurárdalur (Fig. 2). These cross-sections reflect the development of the valleys and their age as will be described in the following pages.
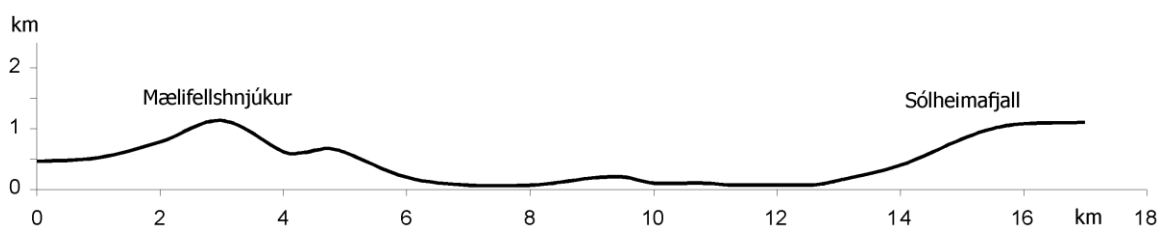
Fig. 6. Simple section across the principal valley of Skagafjörður, height/length = 1/1. Located at the same site as the cross-section in Fig. 7.

The Principal Skagafjörður Valley

The parish of Tungusveit, in the central part of the main Skagafjörður Valley, is a hillocky spur between two of the main rivers of the district, Héraðsvötn and Svartá. A layer of consolidated sediment about 200 m thick covers the floor of the main valley, forming a rugged landscape. This layer belongs to the sedimentary succession of the unconformity and is composed of sandstone, conglomerate and diamictate. This locality, named Eggjar (Fig. 7), lies about 40 km inland from the present shore. The elevation of the valley floor is at about 70 m a.s.l. and below the northernmost part of the sedimentary pile the ancient floor disappears down into the current glaciofluvial flood plain of the river Héraðsvötn. Thus, the erosion level had already become lower than it is today before the accumulation of the sedimentary layer.
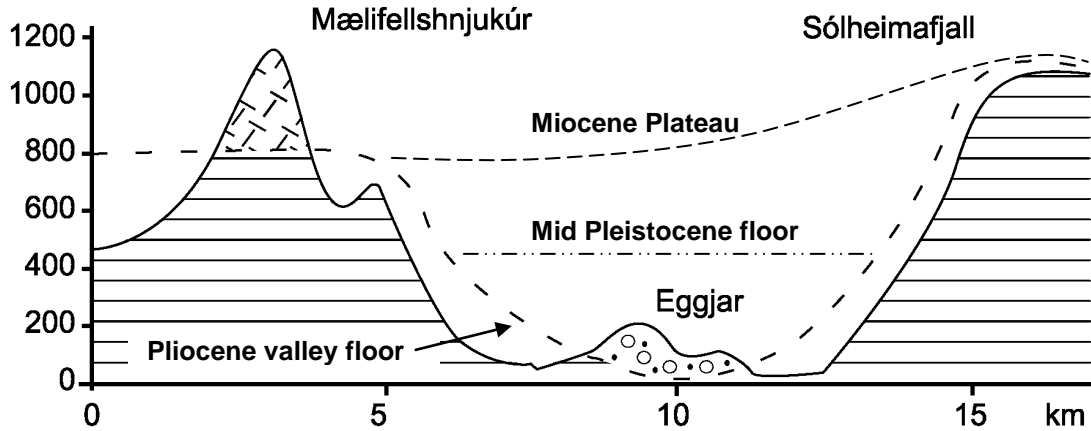
Fig. 7. Skagafjörður principal valley, cross-section. Horizontal lines = the Neogene lava pile (The lavas dip gently towards the south; the section lies along the strike). Dots and circles = the sedimentary succession of the unconformity. Hatched = Pleistocene hyaloclastite in the Mælifellshnjúkur peak. Dashed lines illustrate the elevation of the Pliocene valley at different times; Miocene Plateau 6 Ma, Pliocene valley 3 Ma, mid-Pleistocene valley floor 1 Ma (height/length = 1/5, see Fig. 6).

The size and shape of this primeval valley cannot be reconstructed in any detail, but the mountains on each side were about 1,000 m high. Mælifellshnjúkur, which now reaches 1,138 m, did not exist at this time. It is a hyaloclastite peak, belonging to the Pleistocene volcanic succession, and was formed in a subglacial eruption (Líndal 1940, see also the Appendix 5.1.2). It rests unconformably upon the old Neogene plateau at around 900 m a.s.l. on the west side of the valley.

On the east side, in the Tröllaskagi highlands, the mountains were even higher and remains of the old plateau are seen at 1,100 m. The ancient valley seems to have been a U-shaped valley, as deep or even deeper than the present valley, but narrower (Fig. 7). The development of V- and U-shaped valleys will be discussed later. The sedimentary layer at Eggjar is a remnant of a once more extensive deposit in the valley that was most probably paved by lavas that have since been worn away.

According to Ka/Ar dates and paleomagnetic studies, the Neogene pile is 8-9 million years old (Sæmundsson et al. 1980) but the Mælifellshnjúkur peak is estimated to be about 1.5 Ma (Article 5 in this thesis). In between lies the unconformity, spanning 7 million years.

Vesturdalur

The Neogene stratigraphy of the lower Vesturdalur and Austurdalur valleys is dominated by rhyolite formations from the large, extinct Tinná Central Volcano (Fig. 2). This was active during the period 5-6 Ma and must have been the primary structure of the landscape during most of Pliocene, as has been mentioned earlier. Finally it was covered, or partly covered, by more recent lava flows. The valleys were later carved out on each side of the main body of the volcano.
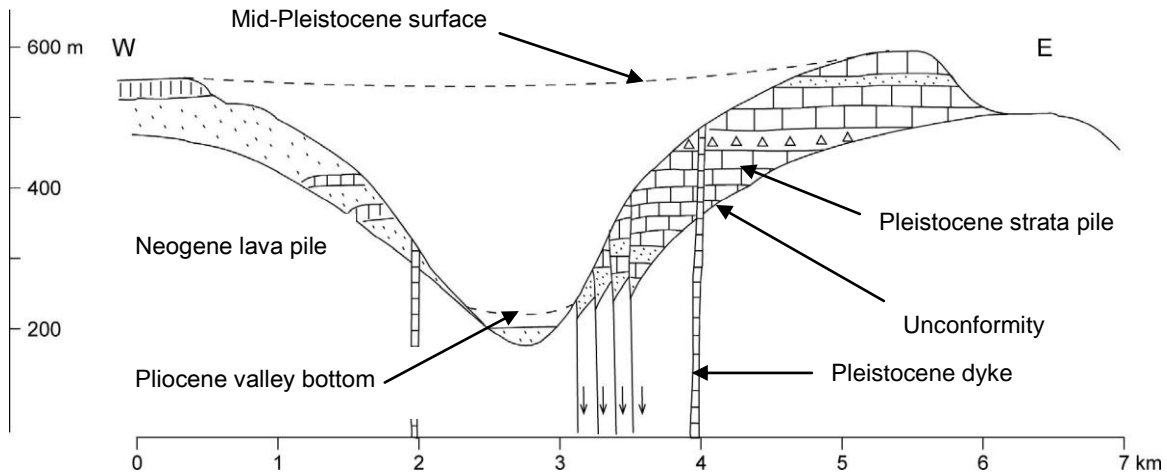
Fig. 8. Cross-section through Vesturdalur, showing the Pliocene valley, the unconformity, the Pleistocene lavas and sedimentary layers that filled up the valley, Pleistocene dykes and faults and the shape of the valley today. The Neogene lava pile dips 5°-10°. (Height/length = 1/5).

The Vesturdalur valley seems, in earlier days, to have been more or less a direct continuation of the principal valley and not a tributary valley as it is today. The ancient form of the outer Vesturdalur valley is preserved below the Pleistocene pile. It was a U-shaped valley, eroded into the old Neogene plateau along the western side of the Skati Dome, the largest rhyolite member of the Tinná Volcano (Article 2 in this thesis). The ancient valley floor can be observed at around 230 m a.s.l., i.e. about 30 m higher than today (Fig. 8). The width of the flat floor seems to have been 500-1,000 m, i.e. similar to what it is now, but the slopes may have been rather more gentle, with the mountains on each side lower than they are today. Later on, sometime in the Mid-Pleistocene, this deep primeval valley was completely filled up by sediments and lavas as is shown in Fig. 8. The unconformity is found on both sides of the valley and can be followed from its mouth and 30 km inland.

The Neogene pile in Vesturdalur is 5-6 Ma. Ma but the earliest lavas of the Pleistocene succession are 1.7 Ma according to Ar/Ar dates (Paper 5 in this thesis) and paleomagnetic studies (Appendix 9.2). Here the unconformity spans 4-5 million years, a considerably narrower time gap than in the principal valley.

Austurdalur

The Austurdalur valley is eroded into the old Neogene plateau along the eastern side of the Skati Dome. The ancient erosion surface of the unconformity is not as continuous as in Vesturdalur but it can be observed here and there below the early Pleistocene succession. The most instructive place is below the Austurdalur Pleistocene Volcano (see Appendix 5.1.1). An eruption in the valley built up the volcano. It rests on thick layers of fluvial sediments that had accumulated in the valley itself before the eruption. Fig. 9

shows a cross-section through Austurdalur and the form of the ancient valley. Here it is suggested that it was U-shaped, like Vesturdalur and the principal valley, but this can not be seen from the cross-section itself. It might have been a little narrower than it is today, and 100-200 m shallower than the present valley, with its floor at 450 m a.s.l. Further inland, the erosion surface appears again below more recent sediments and lavas, reflecting a valley floor at 600-650 m a.s.l. The floor seems to have been more steeply sloping upstream, along the river, than today. According to this, Austurdalur was not as mature in the early Pleistocene time as Vesturdalur; now the valleys are both in a similar state of erosion. The reason is that Vesturdalur was completely filled up by lavas in the early Pleistocene while Austurdalur was only partly filled.
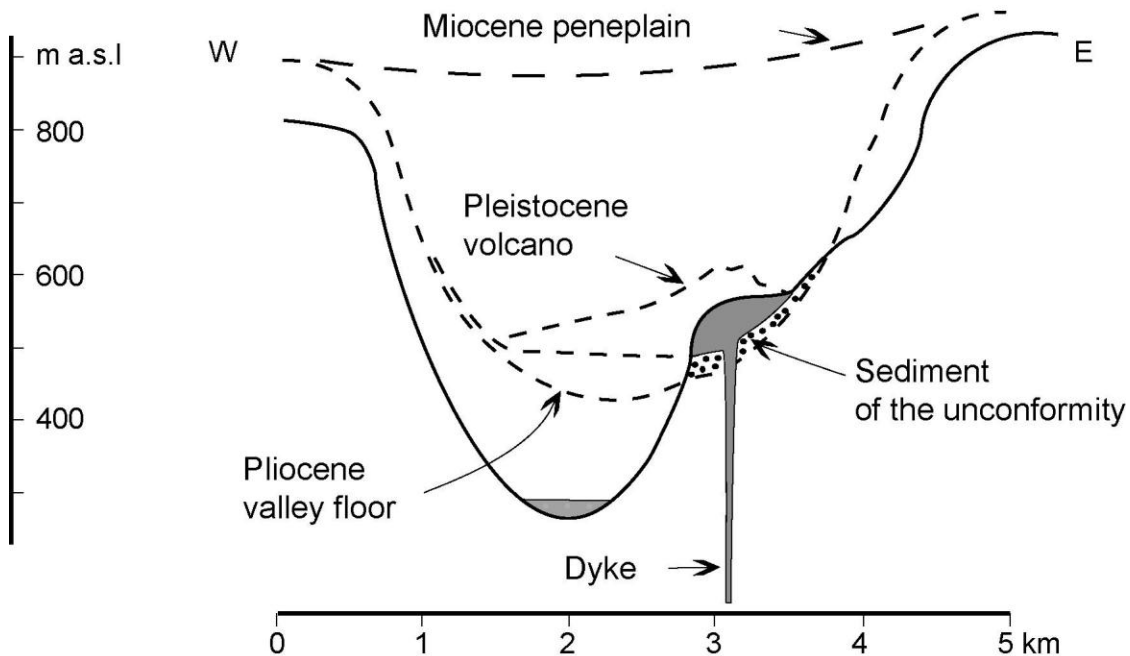
Fig. 9. Cross-section through Austurdalur. Traces of the Pleistocene volcano can be seen in its eastern slopes along with the feeder dyke resting on the remains of old sediments. The dashed lines indicate some stages in the development of the valley; the Miocene peneplain (~5 Ma), the Pliocene valley and its sedimentary layer (~2 Ma). The Pleistocene volcano (~1.5 Ma) is also outlined. Today the valley is 600 m deep; in the Pliocene it was 400-500 m deep.

Ancient sediments found in the slopes of some of the tributary valleys of the second order prove that these already existed in the early Pleistocene.

The age of the Neogene pile in Austurdalur is, as in Vesturdalur, 5-6 Ma. The Pleistocene Volcano has not been dated, but it is most probably of an age similar to that of the Pleistocene lavas of Vesturdalur, i.e. about 1.5 Ma.

Norðurárdalur

Indications of the early existence of the Norðurárdalur valley and its ancient shape are preserved below early Pleistocene volcanic formations at two sites in the valley. These formations are the Kotagil lava and Heiðarsporður hyaloclastite (see Appendix 5.1.3 and 5.1.4). Neither of them has been dated radiometrically but they have reverse polarity, indicating > 0.78 Ma. The Heiðarsporður hyaloclastite was most likely formed subglacially by an eruption in the upper part of the valley. Its base rests on a thin sedimentary layer at between 410 and 510 m a.s.l. (Fig. 10). Here the bottom of the valley is near 280 m a.s.l. On both sides the mountain summits reach 1,100 m. According to this, Norðurárdalur was at least 700 m deep in the early Pleistocene and was only deepened by 120 m during the last one million years of full-scale glaciation.
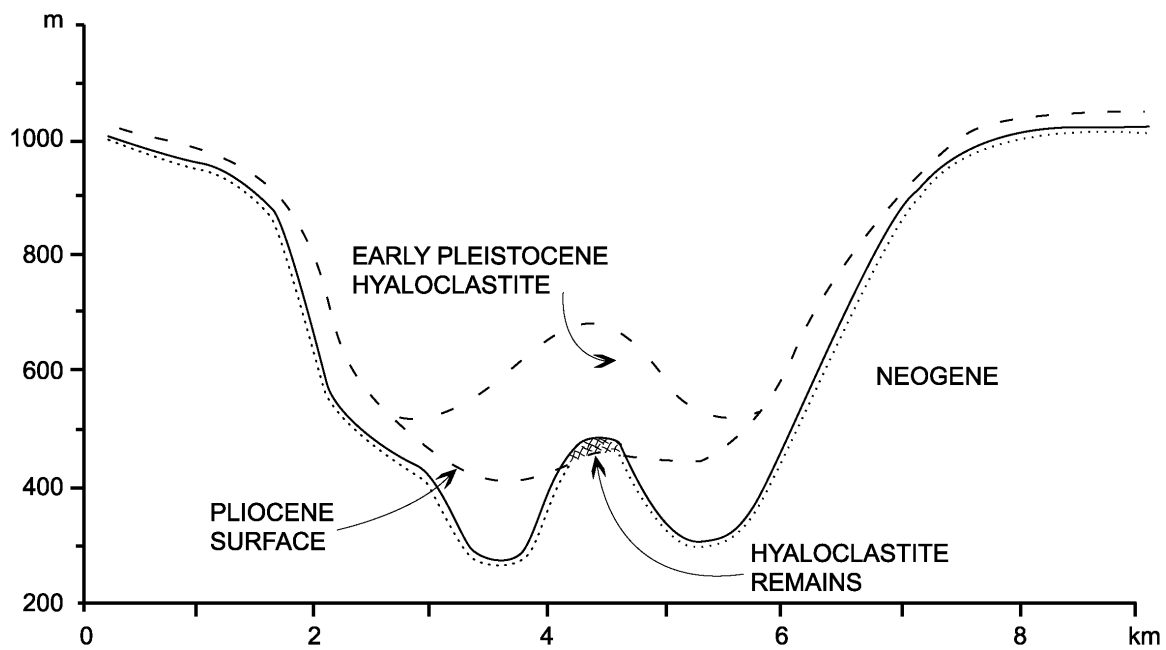
Fig. 10. Cross-section through Norðurárdalur. The hyaloclastite mountain Heiðar-sporður was erupted subglacially in the evolved Norðurárdalur valley in the early Pleistocene. Since then the valley has been deepened by 120 m.

The Kotagil lava is a patch of early Pleistocene interglacial lava erupted near 610 m a.s.l. in a small tributary valley in Norðurárdalur. It filled up a deep ravine that had been cut into the valley floor. Now erosion has reopened the ravine, known as Kotagil, which is a well-known and scenic geological site. The lava not only indicates the age of the main valley but also reflects the early development of its tributaries and their ravines.

The Evolutionary History of the Valleys – Discussion

An account of the evolution of the valley system may be proposed on the basis of the cross-sections described above. All the valleys seem to have undergone similar development during approximately the same span of time; nevertheless it is suggested that the principal valley is slightly older than the tributaries.

Erosion of the ancient Tröllaskagi plateau began after the period of peneplanation and a valley system started to form. At the beginning it might have been dominated by shallow V-shaped valleys excavated by fluvial erosion. As has already been mentioned, the first full-scale glaciations are believed to have occurred later than 2.5 million years ago (Geirsdóttir 1990). Limited glaciations are thought to have overridden the central highlands of Iceland earlier. As the Pleistocene succession in Skagafjörður seems to have been downloaded into mature U-shaped valleys, glacial erosion seems to have turned the initial V-shaped valleys into a system of U-shaped valleys prior to that time, i.e. as early as in the Pliocene.

During the period of peneplanation and valley formation the bedrock was gently tilted towards the volcanic belt where extension and downloading of new volcanic material caused subpression of the underlying strata (Sæmundsson 1979, Appendix 6.1). It is suggested that most of the 200-300-m uplift mentioned earlier also took place during this period as a result of the ablation and unloading of the local bedrock.

In the late Pliocene, 2 Ma, a mature landscape had evolved in Skagafjörður district, featuring the fjord itself, the principal valley, all the main tributary valleys, many of the smaller valleys and even ravines such as Kotagil.

The Sedimentary Succession

According to the cross-sections described above, the situation in the Skagafjörður Valleys changed for some unknown reason in the late Pliocene, when the deepening of the valley system ended and the valleys started to fill up again. An extensive sedimentary succession was deposited unconformably on the eroded bedrock and the Skagafjörður angular unconformity was formed. This change, from carving out to filling up, does not seem to have been due to a great shift in the erosive energy. The grain size in the sedimentary layers indicates large-scale fluvial activity with high transport ability. Instead, the erosion level seems to have changed drastically. Volcanic activity near the shore or in the fjord itself or even offshore on the Skagi Bank might be the reason (Everts 1975), but other changes on a wider scale might also have caused this, such as a major transgression of the sea on the north coast of Iceland, or all over Iceland, possibly as a result of the formation of large glaciers in the central highlands. On the other hand, the eustatic sea level generally fell after the mid-Pliocene warm period (Haq et al. 1987, Sen et al. 1999).

The age or the time-span of the sedimentary succession of the unconformity is somewhat unclear, but here it is suggested that it was formed between 3 and 2 million years ago. The period ended when the lavas of the Pleistocene volcanic succession started to flow along the valleys.

The Pleistocene Succession

Temporary volcanism and rifting resumed in the Skagafjörður Valleys in the early Pleistocene, 1.7 Ma, after millions of years of quiescence (Article 5 in this thesis). It never became intensive and the accumulation of lavas was slow. The lavas flowed along the proto-Austurdalur and Vesturdalur valleys, which at this time had been excavated on

the east and west sides of the Skati rhyolite dome. The western valley seems to have been the larger one and it was there that most of the lavas flowed. Thick sedimentary layers, including tillites, were accumulated between the lavas. Hyaloclastite formations from subglacial eruptions such as those in Mælifellshnjúkur and Heiðarsporður are also found (Figs. 6 and 9). The stratigraphy indicates the cooling climate of the early Pleistocene.

The succession is most extensive in the highlands north of Hofsjökull where it forms a continuous cover. Towards the north it becomes thinner and scarcer, disappearing in the central part of the Skagafjörður district. If it existed there at all, it has been worn away by erosion. Farther north, on the Skagi Peninsula, the Pleistocene succession appears again as a continuous covering of interglacial lavas (Everts 1975).

The Pleistocene eruptive sites and lavas form an elongated zone, the Skagafjörður volcanic belt. Its length, from Hofsjökull to the northern tip of the Skagi Peninsula, is 150 km, and its width is 50-60 km (Article 4 in this thesis). The thickness of the Pleistocene succession is very variable from place to place. It is thickest (up to 300 m) in the old valleys, but becomes thinner, or disappears completely, in the highlands flanking them. In the mid-Pleistocene the elevation level of fluvial erosion was raised by as much as several hundred meters from the former Pliocene elevation, i.e. about 300 m in Vesturdalur, 100-200 m in Austurdalur and over 100 m in the principal valley (Fig. 11).

The present landscape

The activity of the Skagafjörður volcanic belt culminated in the early Pleistocene (Everts 1975, Article 5 in this thesis). In the late Pleistocene it faded out and erosion became the dominating factor again. At the same time, the climate had cooled and full-scale glaciations were spreading across the country. Glaciers and glacial rivers started to clean out, widen and deepen the valleys of Skagafjörður. After a million years of glacial denudation, most of the Pleistocene succession was removed from the valleys, leaving only scattered remains. The valleys became slightly (though surprisingly little) larger than they were at the end of Pliocene. They were only deepened by a few tens of meters, or at the most by 100-200 meters. The glacial erosion of the late Pleistocene seems not to have been the leading factor creating the landscape; it only added the final touch to the topography, that was outlined already during the Pliocene.

Conclusions

It seems possible to divide the history of the landscape in the Skagafjörður Valleys into six main stages spanning the interval from its genesis at the plate boundaries up to the present time. These stages are summarized in Table 1. The following age estimates are mean ages. In reality this development did not occur at exactly the same time all over the area. For example, valley formation must have started earlier in the principal valley than in the tributary valleys.

Stage 1. The formation of the Neogene volcanic succession took place on the Eurasian and North American plate boundaries 9-5 Ma.

Stage 2. As the area drifted gradually out of the volcanic belt, peneplanation started and erosion removed 100-200 m of the topmost layers of the lava plain, forming the extensive lava plateau of Tröllaskagi.

Stage 3. More intensive erosional activity cut the initial Skagafjörður valley system into the volcanic plateau 4-3 Ma. In the Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had been carved out. At the same time the plateau was uplifted by 200-300 m.

Table 1: Main stages in the development of the Skagafjörður Valleys

No.	Stage	Time Ma	
1	Formation of the Neogene volcanic succession	9-5	Accumulation
2	Peneplanation	~5	Erosion
3	Formation of the Pliocene valleys and uplift of the Tröllaskagi plateau	5-3	Erosion
4	Formation of a sedimentary succession and unconformity	3-2	Accumulation
5	Formation of the Pleistocene volcanic succession	2-1	Accumulation
6	Formation of the current landscape	1-0	Erosion

Stage 4. In the late Pliocene the deepening of the valleys came to an end and they began to fill up again, mostly with fluvial sediments. An angular unconformity was formed 3-2 Ma.

Stage 5. In the early Pleistocene a temporary episode of volcanism started up, continuing the filling-in of the valley system. The volcanic activity culminated in the early Pleistocene and died out in the late Pleistocene, 0.5 Ma.

Stage 6. During the last one million years or so, glacial and glaciofluvial erosion cleaned out the ancient valleys and removed most of the Pleistocene volcanic succession, leaving only traces here and there. Although the valley system is more evolved than it was before, the Pleistocene weathering seems to be of less degree than the Neogene erosion.

The elevations of the valley floors have fluctuated up and down due to interaction between erosion and accumulation of sediments and lavas. Fig. 11 indicates the hypothetical elevation changes in the principal valley of Skagafjörður. It is suggested that the area was slowly elevated while it was inside the volcanic belt (Stage 1). During peneplanation (6-5 Ma) a lowering of 200 m occurred (Stage 2). The main lowering, however, took place during the following valley formation. The 1,000 m deep principal valley was excavated over c. 3 million years (Stage 3). After that the valley floor was elevated by 200 m under the accumulation of the sedimentary layer (Stage 4). During the Pleistocene volcanism it was elevated by a further 200 m (Stage 5). Finally it was lowered again by 400 m under the late Pleistocene glacial erosion (Stage 6).
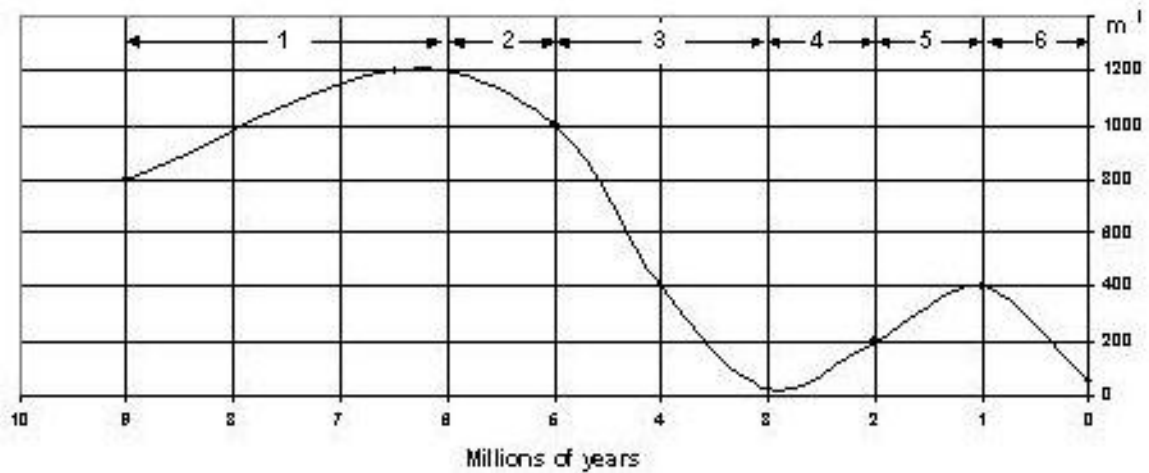
Fig. 11. Hypothetical elevation changes in the principal valley of Skagafjörður. Numbers refer to the stages in Table 1. See the text for explanation.

By analogy it can be predicted that all the main fjords and valleys in the Neogene regions are old in Icelandic geological context and already existed in the Pliocene and some of them even in the Miocene.

This scenario is in contradiction to a common interpretation which states that the Neogene topography was rather even, and that the landscape was dominated by an extensive volcanic plateau, without deep valleys and with central volcanoes as the only high mountains. According to this view, the present rugged topography of Iceland is mainly due to glacial erosion and subglacial volcanism during the Pleistocene (Th. Einarsson 1968, 1994).

The conclusion is more in line with the interpretation that the Icelandic volcanic plateau was already deeply eroded in Neogene times and the main valleys and valley systems existed as early as in the Pliocene (T. Einarsson 1961, 1971).

Acknowledgements

References

Björnsson, H.1988: Hydrology of Ice caps in Volcanic Regions. Societas Scientiarum Islandica 45, Reykjavik, 139 pp + maps.

Einarsson, T., 1958: Landslag á Skagafjallgarði, myndun þess og aldur. [Development and age of topography in the mountains of Skagi]. Náttúrufræðingurinn 28, 1-56.

Einarsson, T., 1959: Jarðeldasvæði um norðanverðan Skagafjörð. Aldursákvörðun á landslagi á Miðnorðurlandi. [The Pleistocene volcanism in the north of Skagafjörður and the paleomagnetic dating of topography in Middel Northern Iceland]. Náttúrufræðingurinn 29, 113-168.

Einarsson, T., 1959b: Studies of Pleistocene in Eyjafjörður, Middle Northern Iceland. Societas Scientiarum Islandica XXXII, 1-79.

Einarsson, T., 1962: Upper Tertiary and Pleistocene rocks in Iceland. Societas Scientiarum Islandica XXXVI. Reykjavik, 197 pp + maps.

Einarsson, T., 1962: Eðlisþættir jarðarinnar og jarðsaga Íslands. [Nature of the Earth and the geological history of Iceland]. Almenna Bókafélagið, Reykjavík, 267 pp.

Einarsson, Th. 1968: Jarðfræði. [Geology]. Mál og Menning, Reykjavík

Einarsson, Th. 1991: Myndun og mótun lands. [Formation and figuration of the landscape]. Mál og Menning, Reykjavík.

Everts, P., 1975: Die Geologie von Skagi und der Ost-Küste des Skagafjords (Nord-Island). Sonderveröffentlichung des Geologischen Instituts der Universität Köln.

Geirsdóttir, Á. 1990: Diamictites of late Pliocene age in western Iceland, Jökull 40, 3-25

Haq, B. U., Hardenbol, J. and Vail, P. R., 1987: Chronology of fluctuating sea levels since the Triassic /250 million years ago to present) Science 235, 1156-1167.

Hjartarson, Á., 1973: Rof jarðlagastaflans milli Eyjafjarðar og Skagafjarðar og ísaldarmenjar við utanverðan Eyjafjörð. [Erosion of the strata pile between Eyjafjord and Skagafjord and glacial formations at outer Eyjafjord].B.Sc. – thesis (unpublished). University of Iceland 37 pp. + map.

Hjartarson, Á., Andersen, L. J., Kelstrup, N. Rasmussen, J. and Struckmeier, W., 1980: International Hydrogeological Map of Europe, 1:1,500,000. Explanatory Notes. Sheet B2 Island. Bundesanstalt fur Geowissenschaft und Rohstoffe, Hannover, UNESCO, Paris. 55 pp + map.

Hjartarson, Á., Friðleifsson, G. Ó. and Hafstað, Þ. H., 1997: Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020 Orkustofnun, Reykjavík, 55 pp. + maps.

Jancin, M., Young, K. D., Voight, B., Aronson, J. L. and Sæmundsson, K., 1985: Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. Journal of Geophysical Research 90, 9961-9985.

Japsen, P., Bonow, J., Klint, K. E. and Jensen, K., 2002: Neogene uplift, erosion and resedimentation in West Greenland. Field report summer 2002. Geological survey of Denmark and Greenland, rapport 2002/71.

Jóhannesson, H and Sæmundsson K., 1998: Geological map of Iceland 1:500.000. Bedrock Geology. Icelandic Institute of Natural History, Reykjavik (2nd edition)

Lacasse, C. and Grabe-Schönberg, C.-D., 2001: Explosive silicic volcanism in Iceland and Jan Mayen area during the last 6 Ma: sources and timing of major eruptions. Journal of Volcanology and Geothermal Research 107, 113-147.

Líndal, J. H., 1940: Mælifellshnjúkur. Náttúrufræðingurinn 10, 51-67.

Pálmason, G. 1980. A continuum model of crustal generation in Iceland: Kinematic aspects. Journal of Geophysics 47, 7-18.

Pálmason, G. 1981. Crustal rifting and related thermomechanical processes in the lithosphere beneath Iceland. Geol. Rundsch. 70, 244-260.

Schlüter, T., 1997: Geology of East Africa. Gerbrüder Borntraeger, Berlin.

Sen, A., Kendall, C. G. St. G. and Levine, P., 1999: Combining a computer simulation and eustatic events to date seismic sequence boundaries: a case study of the Neogene of Bahamas. Sedimentary Geology 125, 47-59.

Sigmundsson, F. 1991: Post-glacial rebound and astenosphere viscosity in Iceland. Geophysical Research Letters 18, 1131-1134.

Sigurðsson, F. and Ingimarsson, J. 1990: Lekt Íslenzkra jarðefna. In ed. Sigbjarnarson 1990: Vatnið og landið. Orkustofnun, Reykjavík 121-28.

Sigurðsson, F., Sæmundsson, K. 1984: Surtarbrandur á Vestfjörðum: Orkustofnun Reykjavík, OS-84039/ODB-02, pp 43.

Sigurðsson, H., 1970: Structural origin and plate tectonics of the Snaefellsnes volcanic zone, western Iceland. Earth and Planetary Science Lettars 10, 129-135.

Sæmundsson, K., 1979: Outline of the geology of Iceland. Jökull 29, 7-28.

Sæmundsson, K., Kristjánsson, L., McDougall, I. and Watkins, N. D.1980: K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. Journal of Geophysical Research 85, 3628-3646.

Walker, G. P. L., 1960: Zeolite zones and dike distribution in relation to the structure of the basalts of eastern Iceland. Journal of Geology 68, 515-528.

The Skagafjörður Volcanic Zone – An ephemeral Rift Zone

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Abstract

In the Skagafjörður district, North Iceland, an unconformity with a sedimentary layer divides the strata pile into Neogene and Pleistocene volcanic successions. Neogene volcanism faded out 4-5 Ma when the area drifted out of the volcanic zone near the diverging boundaries of the North American and Eurasian crustal plates, after which erosion took over for about a million years. Volcanism started again at the beginning of the Pleistocene. Lava flows covered the eroded Neogene landscape and an unconformity was formed. It is proposed that the Pleistocene volcanic rocks above the unconformity all belong to a short-lived axial rift zone, the Skagafjörður zone. Eruptive sites belonging to this zone can be found dispersed in the area between the Hofsjökull central volcano and the mouth of Skagafjörður. The oldest volcanic formations of the Pleistocene rock series are from about 1.7 Ma, when rifting in the area started. The activity culminated in the early Pleistocene but declined in the late Pleistocene, and in the Holocene the activity seems to have been restricted entirely to the Hofsjökull central volcano. A prominent fault and fissure system belongs to the Skagafjörður zone. The tectonics indicate extension. It is suggested that a decline in the activity of the Iceland Mantle Plume, 2-3 million years ago, caused a rift jump away from the hot spot below Vatnajökull and formation of a temporary rift zone, the Skagafjörður zone, which for a while bridged the shortest way between the Reykjanes-Langjökull zone and Kolbeinsey ridge.

Keywords: Volcanic zone, spreading ridge, ridge jump, Iceland hot spot, mantle plume.

1. Introduction

The study area covers in a broad sense the Skagafjörður district, North Iceland (Fig. 1). The southern part of the area is a mountainous terrain with peaks reaching over 1,000 m, culminating in the glacier of Hofsjökull, an active central volcano with an ice-filled caldera and a summit reaching over 1,800 m a.s.l (Björnsson 1988). In the northern part lies the large Skagafjörður fjord, with its small islands, between the major peninsulas Skagi and Tröllaskagi. The North Iceland Volcanic Zone is 100 km to the east and the centre of the Iceland Hot Spot is somewhat farther to the southeast below the Vatnajökull glacier (Sæmundsson 1979).
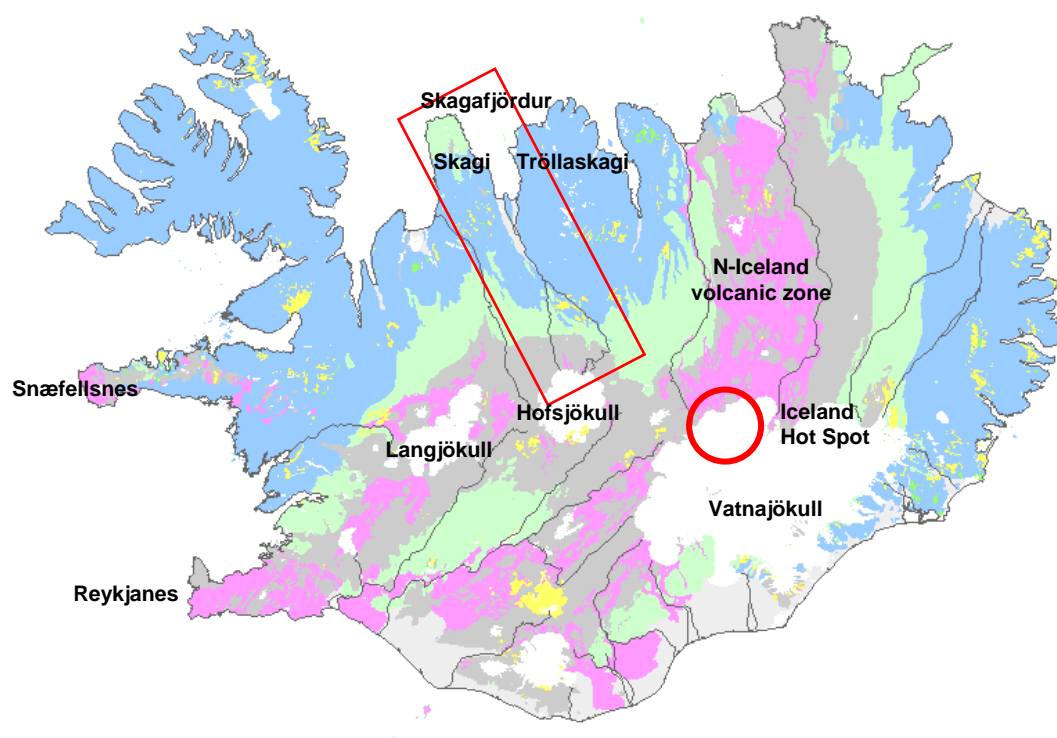
Fig. 1. Geological map of Iceland. The research area is indicated by a red frame. Blue = Neogene regions. Green = Plio-Pleistocene areas (3.3-0.8 Ma). Grey = Late Pleistocene. Violet = Holocene lavas. Yellow = rhyolite. (Modified from Jóhannesson and Sæmundsson 1998).

The recent volcanism of the Skagafjörður zone was recognised by Pjetursson (1905). By investigating the state of erosion and alteration he established a Pleistocene age for the basaltic lavas and plugs in the district. Líndal (1964) investigated the Skagafjörður valleys in 1938 and 1939 and discovered the Skagafjörður unconformity. He described the different dip of the strata above and below it and the thick coarse sedimentary layer at the unconformity.

Einarsson (1958, 1959, 1962) carried out intensive observations on the geology and geomorphology of the Skagafjörður district, and measured sections through the Plio-Pleistocene sediments. He also performed pioneering work in paleomagnetism and used it for dating the rock series (Einarsson 1959, 1962). Everts (1972) and Everts et al. (1975) published a geological map with a detailed description of the northern part of the Skagi Peninsula, including chemical rock analyses and K/Ar dates. Sæmundsson (1974) discussed the tectonics of Skagafjörður and their relation to the Tjörnes Fracture Zone. Sigurðsson et al. (1978) and Schilling et al. (1978) investigated the petrology and structure and geochemical variations in the area between Skagi and the Langjökull glacier.

Various publications on the geology of the Skagafjörður Valleys appeared between 1978 and 2000. Kaldal and Víkingsson (1978) and Harðarson and Guðmundsson (1986) described the stratigraphy, Karlsdóttir et al. (1991) discussed and mapped the geothermal activity, Jóhannesson (1991) published a general geological description, Hjartarson et al. (1998) published a bedrock map with a geological description of the area and Arnórsson and Sveinbjörnsdóttir (2000) described the hydrogeology.
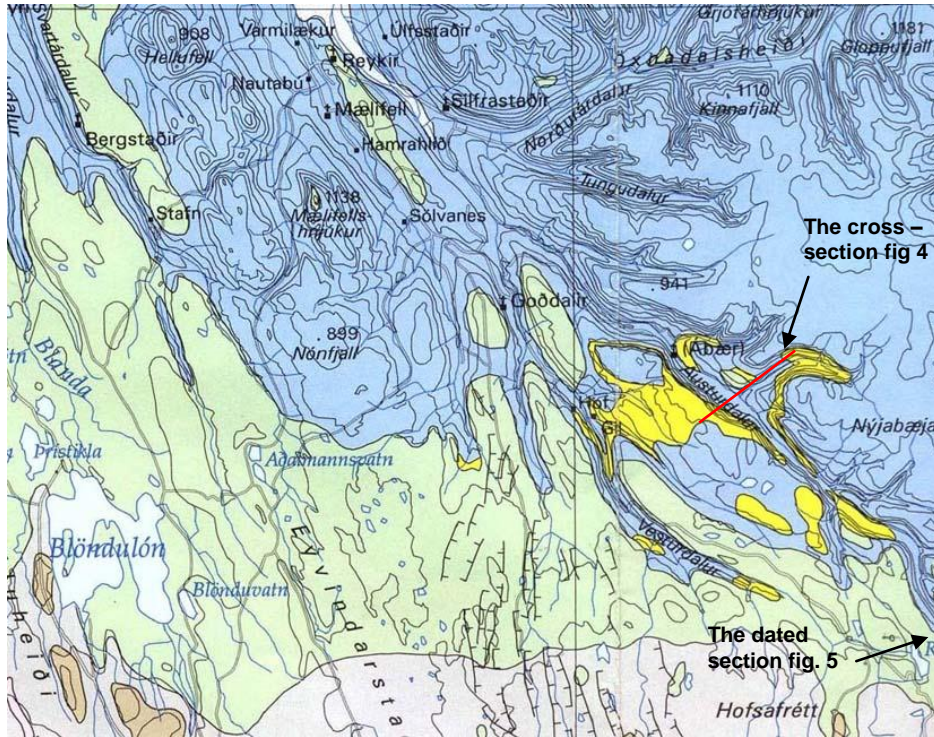
Fig. 2   Main features of the geology of the Skagafjörður valleys. Blue = Neogene succession. Green = Plio-Pleistocene. Grey = Late Pleistocene. Yellow = rhyolites of the Tinná Central Volcano. (From Jóhannesson and Sæmundsson 1998).

2. Rift jumps

The main features of the overall geology and tectonics of Iceland are explained as a complex interaction between the North Atlantic Mid-Ocean Ridge and the Iceland Mantle Plume. Although the North Atlantic region is spreading symmetrically away from the mid-ocean ridge, the ridge itself is drifting slowly to the NW with respect to the plume. About 55 Ma the ridge and the plume started to interact (Vink 1984, Lawver and Müller 1997, Bernstein et. al 1998, Jones 2003). The result was intense volcanism and increased heat flow into the crust. Later on Iceland was formed as a result of this interaction.

Table 1
Rift zones and rift jumps

Rift zone	Time of formation Ma	Ref.
Northwest Iceland rift zone	24	Hardarson et al. (1997)
Snæfellsnes-Húnaflói rift zone	15	Hardarson et al. (1997)
Reykjanes-North Iceland rift zone	6-7	Sæmundsson (1979)
Eastern rift zone	2-3	Sæmundsson (1979)
Skagafjörður rift zone	1.7	The present paper
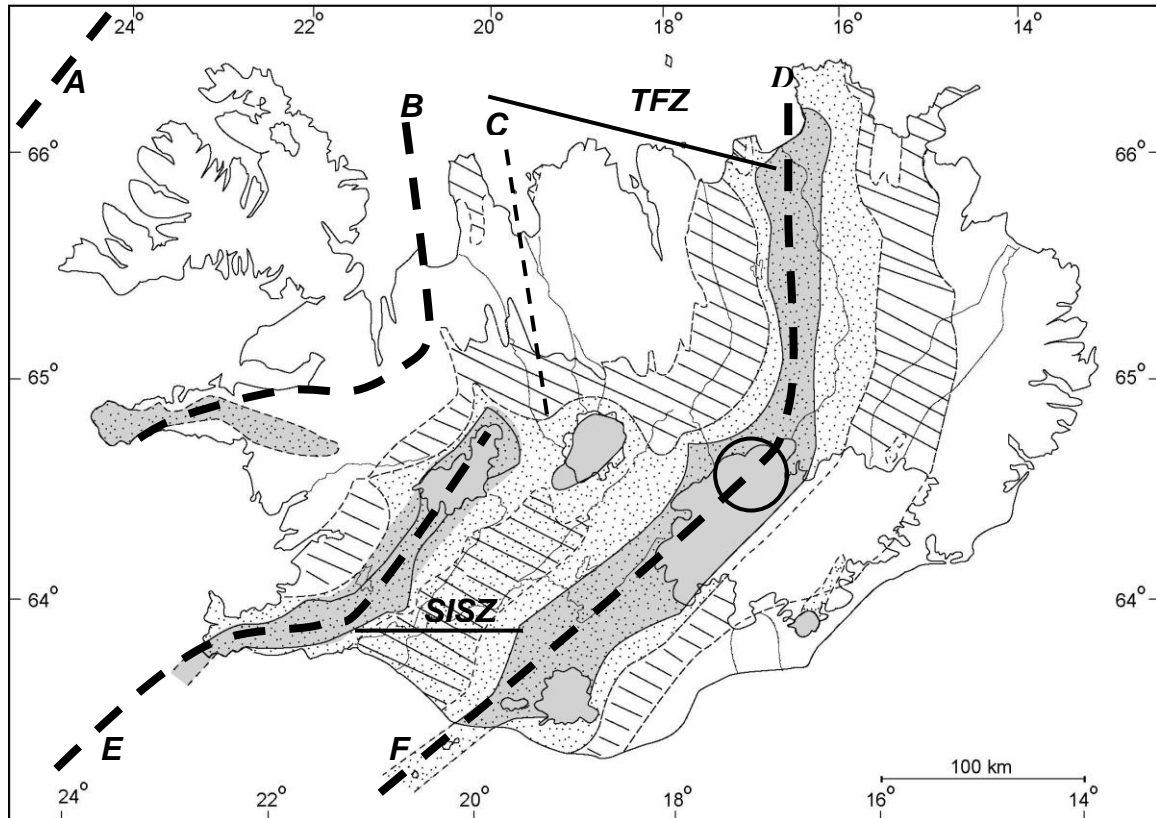
*Fig. 3. The neovolcanic zones and rift systems in Iceland. **A** = Northwest rift axis; **B** = Snæfellsnes-Húnaflói rift axis; **C** = Skagafjörður rift axis; **D** = North Iceland rift axes; **E** = Reykjanes-Langjökull rift axis; **F** = The propagating Eastern rift axis. **SISZ** = South Iceland seismic zone; **TFZ** = Tjörnes fracture zone. The circle represents the location of the Iceland plume.*

As the ridge migrated over the plume the latter is thought to have shifted the spreading axis repeatedly through rift jumping. Several jumps have been proposed in the geological history of Iceland (Table 1). The Northwest rift zone is thought to have formed about 24 Ma west of the Northwest Peninsula (Fig. 3) (Hardarson et al. 1997). It was active for 8-10 million years but then rift jump occurred and a new zone was established in the Snæfellsnes-Húnaflói region. There it was active for another 8-10 million years but c. 6-7 Ma it jumped again to form the North Iceland zone (Sæmundsson 1974,1979, Jóhannesson 1980, Kristjánsson and Jónsson 1998). In Southwest and South Iceland the picture is a little more complicated as there exist two parallel rift zones. The rift zones and rift jumps are listed in Table 1.

These jumps seem to be related to a tension between the mantle plume and the ridge. The rift axis crossing Iceland does not bridge the gap between the Reykjanes and Kolbeinsey ridges (Fig. 3). Instead it has been displaced eastwards and is connected to the main ridges by transform faults (Sæmundsson 1979, Hardarson et al. 1997). The plume tends to trap a segment of the ridge above itself by transform displacements. The force required for this is released in the earthquakes on the transform faults in South Iceland and off the north coast of the country.

3. The stratigraphic successions

The bedrock and strata pile of the Skagafjörður district can be divided into three main successions (Fig. 1):
1. Neogene volcanic succession
2. Sedimentary group of the unconformity
3. Pleistocene volcanic succession
The major Skagafjörður unconformity divides successions 1 and 2.

3.1. The Neogene succession

The Neogene volcanic succession below the unconformity originates in axial rift zones at the boundary between the North American and Eurasian crustal plates, the lower part (> 7 Ma) in the Snæfellsnes-Húnaflói rift zone but the upper part (< 7 Ma) in the North Iceland rift zone (Article 5 in this thesis). The Neogene succession is composed predominantly of basaltic lavas with thin fine-grained interbeds but there also exists a voluminous silicic centre, the Tinná Central Volcano (Article 2 in this thesis). Although this was not a stratovolcano, it rose above the environment and formed a mountainous area of rhyolite domes and other eruptive centres several hundred metres high. Paleomagnetic observations and Ar/Ar-dates indicate that the volcano was active 5-6 million years ago. A detailed description of the Neogene stratigraphy is given in the Appendix. Volcanism died out as the area drifted out of the volcanic zone.

3.2. The Skagafjörður unconformity

The unconformity between the Neogene succession and its sedimentary group is marked by an erosion surface that outlines the ancient valley system of the district. The unconformity is marked by a change in the strike and dip of the adjacent successions. Lava flows below it dip 5 - 15° S whereas flows above it dip 2-8° S, towards the Hofsjökull massif (Appendix, 6.1). This southward dip fluctuates east-west by as much as15-20°, but the main orientation is to the south.

   The sedimentary layers inside the successions below and above the unconformity differ, reflecting dissimilar climatic and environmental conditions. Below the unconformity, the stratigraphy is indicated by thin tuffaceous red-brown silt layers of aeolian origin and fine-grained, fluvial and lacustrine layers with regular bedding. The sediments seem to have been formed in a temperate climate. Above the unconformity, the sediments are much coarser and more irregular in structure and bedding. Till layers and glaciofluvial sediments have been identified (Appendix, 4). The eruptives also change, both petrologically and structurally. Composite olivine basalt layers forming shield volcanoes and hyaloclastite formations, both very rare below the unconformity, become dominant in the upper pile.
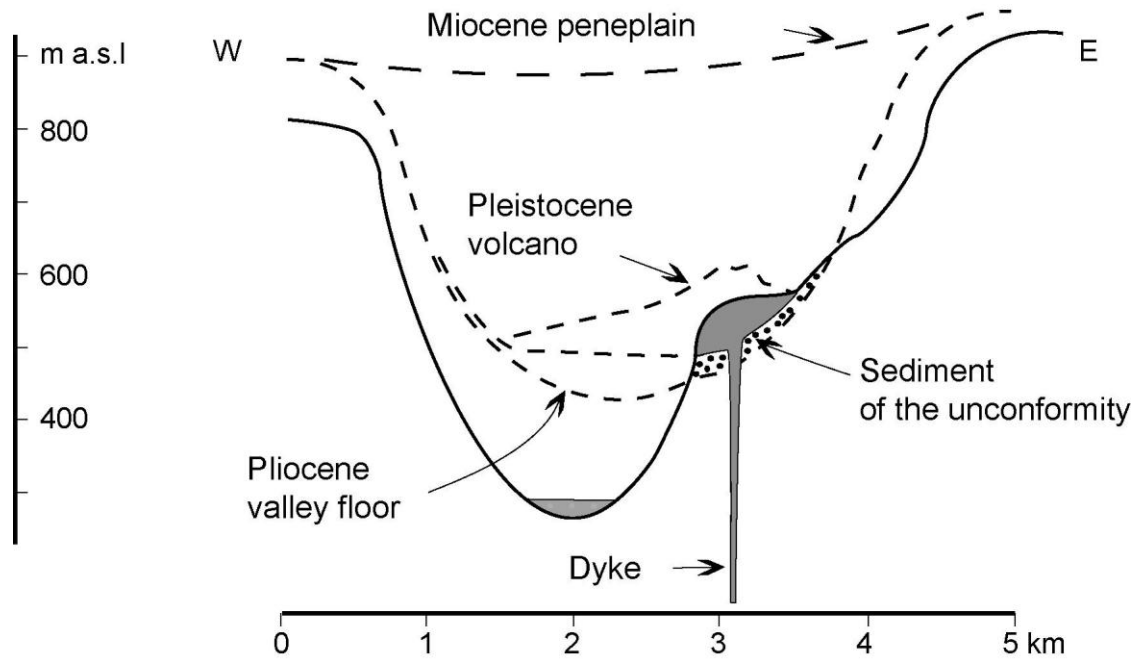
Fig. 4. Cross-section through the Austurdalur valley. Traces of the Pleistocene volcano can be seen in its eastern slopes along with the feeder dyke resting on the remains of old sediments. The dashed lines indicate some stages in the development of the valley; the Miocene peneplain (~5 Ma), the Pliocene valley and its sedimentary layer (~2 Ma). The Pleistocene volcano (~1.5 Ma) is also outlined. Today the valley is 600 m deep, in Pliocene it was 400-500 m deep.

The unconformity indicates that the area drifted away from the active volcanic zone. Accumulation of new extrusives halted and erosion took over, weathering the surface and carving out new landscape. The surface of the unconformity reflects the late Neogene - early Quaternary topography and the initial form of the Skagafjörður valley system. The valleys were mature and already several hundred metres deep, carved out of the primordial volcanic plateau (Fig. 4, see also Article 3 in this thesis). The forerunners of Austurdalur and Vesturdalur valleys were eroded on each side of the Tinná volcano. The principal valley of Skagafjörður district was eroded below the recent sea level but was narrower than it is today.

3.3. The sedimentary group

After the erosion of the valley system it started to fill up again for some unknown reason. Thick sedimentary layers were accumulated forming the sedimentary group of the unconformity that can be traced for long distances. It is thickest in the valleys, reaching tens or even hundreds of meters, but thinner in the highlands and completely absent in some places. In the central part of the Skagafjörður Valley it forms a 50-200 m thick uncapped layer. There the Pleistocene volcanic succession, if it ever existed at all, has been entirely eroded away. The sediments divide the Neogene and Quaternary successions on the Skagi Peninsula, as they do in the Skagafjörður valleys (Everts 1975). The southernmost locality is at the rivulet Geldingsá, where the sediments form a 55 m thick clastic layer (Fig. 5).

Tholeiite lava

Tholeiite
Ar/Ar=1.66 Ma

Tuff 0.5 m
Sandstone 0.2-0.4m

Tillite, dark gray
clastic silt, boulders up to 1 m across
robust layering

Erosion surface, unconformity

Olivine tholeiite 6 m
scattered phenochrysts << 1%
Conglomerate 0.5 m

Porphyritic basalt 14 m
compound Lava
plagloclase phenochrysts 10-15%

Ar/Ar=2.49 Ma

Black sandstone 0.5-1m
Tillite (?)

Tholeiite 10 m
scattered phenochrysts
1-2 % plagioclase

Porphyritic basalt 5 m, 10% plagioclase
Ar/Ar=2.84 Ma

Fig 5. A section across the unconformity at Austurdalur valley near Geldingsá, north of Hofsjökull. Here the unconformity spans 0.8 million years. The location of the section is shown in Fig. 2.

The internal structure of the sediments, and their thickness, vary from place to place. The layering is sometimes irregular but more often roughly horizontal. In many localities the grain size has an upward grading i.e. with a fine-grained lower part and a coarser upper part. It is most obvious in the slopes of Vesturdalur and its tributaries. The grain size distribution is unusual, with rounded boulders up to 1 m in diameter or more situated in a sandy and silty ground mass. Most of the sediment seems to be of fluvial and glaciofluvial origin, reflecting a glacial and periglacial environment and rivers with a high transport ability.

3.4. The Pleistocene succession

Volcanic activity started again in the Skagafjörður area in the early Pleistocene after a quiescence that had lasted for millions of years (Article 5 in this thesis). New lavas covered the land and filled up depressions and valleys forming the Pleistocene volcanic succession. This is divided into three series based on stratigraphical position and paleomagnetism. These consist of early Pleistocene formations with reverse

paleomagnetic polarity (the topmost lavas in Fig 5), mid-Pleistocene formations, also with reverse polarity but starting with a normal paleomagnetic subchron (Yaramillo) and late Pleistocene formations spanning the Brunhes paleomagnetic chron (Appendix, 5; see also the Bedrock map).

The oldest known formations of the Pleistocene succession are individual lavas of olivine and porphyritic basalts covering the sediments of the unconformity. They are they have been studied most thoroughly in the Vesturdalur valley both to the west and to the east of Goðdalir. These lavas seem to have flowed along the early Pleistocene valleys of the district, covering fluvial deposits. The Austurdalur Pleistocene Volcano is thought to belong to this formation. It was made of porphyric basalt lava erupted on a short fissure in the Austurdalur valley. Erosion has cut a prominent section through the lava, its feeder dyke and the sediments below (Fig. 4). Later on widespread tholeiites were accumulated in the southwest part of the field. They were followed by an early Pleistocene composite olivine basalt formation. All the lava flows have reverse paleomagnetic polarity (R).

The mid-Pleistocene formation begins with a short normal paleomagnetic subchron. It has been identified in a few places, comprising 1-3 layers in each locality. They form a discontinuous horizon inside the succession which is mostly made of porphyritic and olivine tholeiite lavas. Glaciogenic layers have not been found inside the formation. It has been suggested that the subchron represents the Yaramillo geomagnetic event, which is about 1 million years old (Hjartarson et al. 1997). Above this horizon, hyaloclastite formations become more common than before. Several hyaloclastite mountains have been mapped in the southern part of the area. They are thought to have been formed subglacially. There are also widespread composite olivine tholeiites and tholeiite lavas. Volcanic products of the northern part of the zone consist predominantly of interglacial lavas.

The most recent volcanic formations are thought to be from the late Pleistocene, less than 0.8 million years old and with normal polarity of the Brunhes magnetic chron (Appendix, 5.3). At this time the volcanism was declining and retreating to the southern part of the area. There is a continuous cover of recent formations, shield volcanoes and lava fields from the interglacials and pillow lavas and hyaloclastite ridges from the glacial periods. These are scattered over the northern part of the district. The best-known member of this formation is the island of Drangey (Jóhannesson 1991), a famous historical site depicted on the coat of arms of the Skagafjörður district (its location is shown in Fig. 5). Active volcanism during the Holocene was entirely restricted to the Hofsjökull central volcano (Jóhannesson and Sæmundsson 1998a).

4. Volcanic sites and the size of the Skagafjörður Zone

The length of the Skagafjörður zone, from Hofsjökull to the northern tip of the Skagi Peninsula, is 150 km. Its width is 50-60 km (Fig. 6). The area on land is therefore 8,250 km^2 but it is a reason to believe that the Skagagrunn bank north off Skagi is a continuation of the Pleistocene basalt area (Everts 1975). The thickness of the Pleistocene eruptives varies greatly from place to place, from about 100 m on the Skagi Peninsula in the north to 200-300 m in the central highlands north of Hofsjökull. All the Pleistocene lavas in the main Skagafjörður Valley have been eroded away, if indeed they ever existed there. The geological evidence indicates a low productivity of volcanic material.

Table 2 lists known Pleistocene volcanoes and volcanic vents in the Skagafjörður zone. The central part of the area, from the Norðurárdalur valley to Sauðárkrókur, is without any known eruptive centres (Fig. 6).
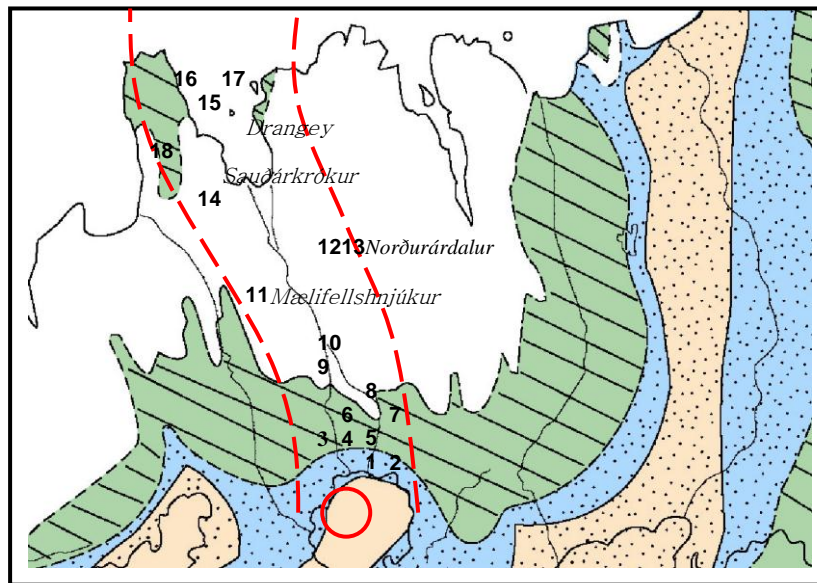
Fig 6. The Skagafjörður Volcanic Zone. Numbers = volcanic vents (see Table 2). Circle = the Hofsjökull caldera. Brown = Holocene volcanic zones. Blue = Late pleistocene. Green = Plio-Pleistocene (0.8-3.3 Ma) White = Neogene (>3.3 Ma).

Table 2
Pleistocene volcanoes and volcanic vents in the Skagafjörður Volcanic Zone

No.	Name or place	Type	Polarity	Estim. age Ma
1	Orravatn basalt	Crater row?	N	0.5
2	Orravatn	Shield volcano	N	0.5
3	Austurkvísl	Shield volcano	N	0.5
4	Sáta	Shield volcano	N	0.5
5	Hraunþúfa	Shield volcano	R	1
6	Bleikáluháls	Hyaloclastite mountain	R	1
7	Geldingsá	Dyke	R	1-2
8	Austurdur Pleist. volcano	Interglacial lava and dyke	R	1-2
9	Goðdalafjall	Dyke	R	1-2
10	Hlíðarfjall	Dyke	R	1-2
11	Mælifellshnjúkur	Hyaloclastite mountain	R	1-2
12	Kotagil, Norðurárdalur	Lava remains		
13	Heiðarsporður, Norðurárd.	Hyaloclastite remains		
14	Gönguskörð at Tindastóll	Subglacial pillow lava	N	< 0.8
15	Drangey	Submarine hyaloclastite ridge	N	< 0.8
16	Ketubjarg	Interglac. hyaloclastite + dyke	R	1-2
17	Þórðarhöfði	Interglacial lava and dyke	N	< 0.8
18	Selfjall	Craters	R	1-2

5. Tectonics

Faults and fissures of the Skagafjörður valleys can be divided into three main systems; the Neogene system the active Pleistocene - Holocene system and the Hofsjökull fissure swarm. The main trend of them all of them is N-S, with normal faulting (Karlsdóttir et al. 1991). The fourth system lies north of the mouth of Skagafjörður, the active transverse tectonics of Northern Iceland (the Tjörnes Fracture Zone), trending WNW (Garcia et al. 2002).

5.1. Neogene faulting

Neogene faulting is rather difficult to distinguish in the landscape. Erosion has wiped out the scarps of the faults and the fissures are filled with secondary minerals. This faulting is restricted to the Neogene regions and in some places it is covered by younger formations. Most of the faults are small (0-20 m throw) but faults with up to 120 m throw have been mapped. An important part of the tectonics is the caldera fault of the Tinná Central Volcano (Article 2 in this thesis, see also the Appendix, 6 and the Bedrock map). Neogene tectonics were active in the Late Miocene - Pliocene while the Neogene succession was accumulating but became extinct when the area had drifted out of the volcanic zone.

5.2. Pleistocene-Holocene tectonics

Pleistocene-Holocene tectonics are easily recognized in the area as they often left scarps and clear-cut lineaments in the landscape. They also cut through the whole pile, the Neogene and Quaternary successions and in some places the loose overburden is also disturbed. A prominent bundle of faults has been mapped near the 75 km long eastern border of the Skagafjörður district (Jóhannesson and Sæmundsson 1998b). In the eastern slopes of Vesturdalur valleys lies a system of faults that cuts the Pleistocene pile. It is composed of four normal faults trending NW - NNW. The throw is towards the east, 50 –100 m altogether. Lava has flowed into one of the fault indicating that its has the same age as the lava pile and the other faults are most likely of similar age. This faulting is believed to be related to rift tectonics.

   Weak and scattered seismic unrest indicates active tectonics throughout the area between Hofsjökull and the Skagafjörður mouth. Larger earthquakes occur occasionally. The largest recorded one was 5.0 on theRichter scale. It occurred on the sea floor, 7 km northwest off Drangey, on 11 July 1964 (Icelandic Meteorological Office, homepage).

5.3. The Hofsjökull fissure swarm

The Hofsjökull fissure swarm cuts through the highlands north of the Hofsjökull glacier forming long, prominent scarps and grabens. It seems to be connected to the huge ice-filled caldera of the volcano. The age of the caldera is not known. Surface disruption, weak seismicity and volcanism under the glacier indicate permanent activity. The length of the fissure swarm is somewhat unclear as it grades into the Pleistocene fault system. The main trend is N-NNW (Karlsdottir et al. 1991, Vilmundardóttir et al.1997).

   The active Pleistocene-Holocene tectonics and the Hofsjökull fissure swarm are in fact manifestations of the same phenomenon, i.e. an axial rift zone in the Skagafjörður

district. Although it is fading out it is not quite extinct, but the spreading is presumably slowly approaching zero.

6. Geothermal fields

The Skagafjörður district is known for its geothermal fields (Karlsdóttir et al. 1991). The activity is far higher than in the adjacent districts in terms of both heat and utilisation. Five municipal heating services operate in the district, togetherwith many small private heating centres for house and greenhouse heating. All the geothermal fields are in the Neogene bedrock but are connected to the young tectonics that broke up the strata pile, forming geothermal aquifer systems. No explanation has been given for this high discharge of geothermal heat in the Skagafjörður district but here it is assumed to be connected to the short lived Pleistocene axial rift zone of Skagafjörður.

7. Petrology

Sigurðsson et al. 1978 made an extensive petrological and chemical study of basalts in a 220 km long profile from the Langjökull area to the tip of the Skagi Peninsula. They found an abrupt change at 65º 10', about 20 km north of the glacier, that separates the areas petrologically. Basalts in the Langjökull area are relatively primitive MgO and Al_2O_3-rich olivine tholeiites. Basalts of the Skagafjörður zone, on the other hand, are relatively evolved iron-titanium-rich tholeiites. Rocks most comparable to the Skagi tholeiites are found among the transitional alkali basalts in the southern parts of the Eastern Volcanic Zone (Sigurdsson et al. 1978). Jakobsson (1979) found a similar abrupt change in the propagating Eastern Volcanic Zone where the petrology of the eruptives changes from tholeiitic to transitional alkali basalts.

8. The age of the Skagafjörður Zone

When the pioneer geologist Helgi Pjetursson (1905) discovered the young formations in Skagafjörður district at the beginning of the 20th century he proposed that they were of Pleistocene age. Later observations have not changed this estimate. It was improved by the geological investigation and paleomagnetic work of Einarsson (1959) and was further strengthened by K/Ar dates (Everts et al. 1972, Everts 1975). Everts's samples were collected on the Skagi Peninsula and from a recent formation on the east coast of Skagafjörður. The dates give the time interval 0.5 - 2.6 Ma but the margins of error are large in some cases.

In the present study seven samples were collected in the Skagafjörður Valleys for Ar/Ar dating. Four of them were taken from basaltic lavas directly above the unconformity and two from basaltic lavas directly below it. The last one was collected from the Skati rhyolite dome, a little further down in the Neogene pile. The distribution of the samples should reveal the supposed widening of the age gap of the unconformity from south to north, from the highlands near Hofsjökull and northwards to the lower Vesturdalur valley (Table 3) (Article 5 in this thesis).

The dates from above the unconformity indicate that the activity of the Skagafjörður Volcanic Zone started near the beginning of the Pleistocene, 1.6-1.7 Ma. According to the date no. 17177 in Austurdalur, near the confluence with Geldingsá, volcanism began at 1.66 Ma. A little later the first Pleistocene lava flows reached the lower part of Vesturdalur 40 km farther north, which is where dates nos. 17160 and 17169 are from. The lavas are 1.48 Ma old and lie directly on the sediments. The

dated samples have reverse polarity and fit with the mid Matuyama geomagnetic field prevailing 1.07-1.77 Ma (Cande and Kent 1995).

The date from Austari Jökulsá near Geldingsá (no. 17175) below the unconformity gave 2.49 Ma or upper Pliocene age. It has normal polarity and has the best fit with the Reunion II (polarity subchron C2r.2r-1).

Dates nos. 17175 and 17177 are from the same section and give the maximum time gap of the unconformity at its southernmost exposure. There it spans 0.8 million years. Towards the north it widens mostly due to the erosion of the Neogene succession and is about 4 Ma in lower Vesturdalur.

Table 3.
Ar/Ar dates from the Skagafjörður valleys*

Locality (m a.s.l.)	Relation to the unconformity	Pol- arity	No.	Ar/Ar-age Ma ± 1σ
Djúpagil innra, 386 m, Vesturdalur	Above	R	17160	1.482 ± 0.060
Hlídarfjall, 335 m, Vesturdalur	Above	R	17169	1.468 ± 0.017
Goðdalakista, the topmost lava, 585 m.	Above	R	17172	1.256 ± 0.016
Austurdalur, near Geldingsá, 600 m.	Below	N	17173	2.841 ± 0.043
Austari Jökulsá, 620 m, near Geldingsá	Below	N	17175	2.486 ± 0.068
Austari Jökulsá, 660 m, near Geldingsá	Above	R	17177	1.659 ± 0.024
Skati Dome, Tinnárdalur, 650 m	Below	R	17181	5.212 ± 0.016

* The dates are described in Article 5 in this thesis

9. Discussions

The recent volcanism in the Skagafjörður Valleys and on the Skagi Peninsula has long been a matter of discussion and its relation to the axial rift zone of North Iceland has been unclear. For a while it was supposed to be a forerunner of the present-day North Iceland rift zone (Sæmundsson 1967, 1974, Everts 1975) but later it has been regarded as an instance of intraplate volcanism (Sæmundsson 1979).

Sigurðsson et al. (1978) investigated the petrology and structure in the area between Skagi and the Langjökull glacier and found that the recent volcanism was rift-related. They suggested that 2.5 Ma rifting in the Reykjanes-Langjökull zone propagated northward through a 15 km thick crust, giving rise to an ephemeral Skagi zone. They also proposed that this rifting was related to a period of increased spreading in Northern Iceland 0.5-2.5 Ma or, alternatively, that the Skagi zone might have arisen owing to temporary cessation of rifting in the North Iceland zone and consequent rift jumping to the west. This latter assumption seems in good accordance with the geological evidence in the Skagafjörður district described in this paper.

The classification of Hofsjökull inside the volcanic belts of Iceland where it towers above the surrounding area in between the Reykjanes-Langjökull Volcanic Zone and the North Iceland zone, has been problematic. It has sometimes been claimed that it functions as a connection between the main volcanic areas, the Reykjanes-Langjökull zone, the North Iceland zone and the Eastern zone (i.e. Einarsson 1991). In many recent papers it is interpreted as an independent spreading zone, the Mid-Iceland zone (e.g. Hardarson et al., 1997). This opinion might be adapted to the idea of the Skagafjörður rift zone. Accordingly, the activity in Hofsjökull and its fissure swarm should be interpreted as the remains of activity in a deceasing rift zone.

The Skagafjörður Volcanic Zone was formed around 1.7 Ma. Then a rift jump seems to have occurred from the North Iceland Rift with a 100 km displacement

towards the west. This jump was in the opposite direction to previously recorded rift jumps in Iceland, away from the hot spot. No hiatus is known inside the Pleistocene succession of the North Iceland rift zone, so the Skagafjörður zone only partly took over the spreading activity. This is the explanation of its rather low productivity of volcanics. For one million years or so, two parallel rift axes were active simultaneously in North Iceland.

A reason for this event and a rift jump towards west, away from the plume, might be advanced. Sigurdsson et al. (1978) suggested that rifting in Skagi was related to either a period of increased spreading in North Iceland 0.5-2.5 Ma or to a temporary cessation of rifting in the North Iceland zone and consequent rift jumping to the west. Here the later assumption will be adopted and connected to ideas about a pulsating plume.

The plume has caused several rift jumps as the ridge drifts across it, as mentioned earlier. At certain times the activity of the plume culminates, but at other times it declines. V-shaped ridges oblique to the Reykjanes ridge are thought to reflect these pulses, and they are thought to indicate peaks in the plume's activity 15 and 6 Ma (Voigt 1971, Voigt and Avery 1974, Jones 2003). This would suggest that the energy brought to the crust differs from one time to another.

As shown by Abelson and Agnon (2001) the Iceland Plume seems to have been in a phase of low activity about 1.5 million years ago. Here it will be suggested that at that time it lacked the energy to control the North Iceland rift zone and keep it in place and thus it failed to maintain the transform faults and associated earthquake activity. A new rift zone without any transform faults started to form, bridging the shortest way between the Reykjanes and Kolbeinsey ridges. This became the Skagafjörður zone. Volcanism started in Skagafjörður and its valleys. Later on the power of the mantle plume increased again. It gained energy to keep control over the rift zone. Volcanic activity increased in the North Iceland Volcanic Zone but died out in the Skagafjörður area. The zone there was left incomplete. This discussion leads to the prediction that the Skagafjörður ridge jump was an overture for the next major rift jump in the country, which will be towards west, indicating a parting of the Mid-Atlantic ridge and the Iceland Hot Spot.

10. Conclusions

The geological setting, recent volcanic formations, tectonics and geothermal activity on a rather narrow and well-defined belt between Hofsjökull and Skagi, together with submarine formations in the Skagafjörður fjord and north of Skagi, leads to the conclusion that this is a short-lived and immature rift zone. For a while it bridged the shortest way between the Reykjanes-Langjökull ridge and the Kolbeinsey ridge. Spreading, together with spreading volcanism, started in the area around 1.7 Ma. The zone seems to have propagated from both ends, towards north from the Hofsjökull area and towards south from the submarine Kolbeinsey ridge. The central part of the area, between the propagating belts, perhaps never became volcanically active; at any rate, no eruptive centres are known there. The zone was uncompleted in the sense that the activity was short-lived: it did not form a tectonic syncline or a flexure or central volcanoes, with the exception of Hofsjökull, the only active volcano of a spreading zone with diminishing activity.

Acknowledgements

References

Abelson, M. and Agnon, A., 2001. Hotspot activity and plume pulses recorded by geometry of spreading axes. Earth and Planetary Science Letters 189, 31-47.

Arnórsson, S. and Sveinbjörnsdóttir, Á.E., 2000. Age and chemical evolution of groundwaters in Skagafjördur, Iceland. Journal of Geochemical Exploration 69-70, 459-463.

Bernstein, S., Kelemen, P.B., Tegner, C., Kurz, M.D., Blusztajn, J. and Brooks, C.K., 1998. Post-breakup basaltic magmatism along the East Greenland Tertiary rifted margin. Earth and Planetary Science Letters 154, 221-235.

Björnsson, H., 1988: Hydrology of Ice caps in Volcanic Regions. Societas Scientiarum Islandica 45, Reykjavik, 139 pp + maps.

Einarsson, P. and Björnsson, S., 1979. Earthquakes in Iceland. Jökull, 29: 37-43.

Einarsson, T., 1958. Landslag á Skagafjallgarði, myndun þess og aldur. Náttúrufræðingurinn, 28: 1-56.

Einarsson, T., 1959. Jarðeldasvæði um norðanverðan Skagafjörð. Aldursákvörðun á landslagi á Miðnorðurlandi. Náttúrufræðingurinn, 29: 113-168.

Einarsson, T., 1962. Upper Tertiary and Pleistocene rocks in Iceland. Societas Scientiarum Islandica XXXVI. Reykjavik.

Einarsson, Th., 1991. Myndun og mótun lands. [Formation and figuration of the landscape]. Mál og Menning, Reykjavík.

Everts, P., Koerfer, L.E. and Swartzbach, M., 1972. Neue K-Ar-Datierungen islandischer Basalte. Neues Jahrbuch Geol. Palaeontol. Monatsh., 5: 280-284.

Everts, P., 1975. Die Geologie von Skagi und der Ost-Küste des Skagafjörðurs (Nord-Island). Sonderveröffentlichung des Geologischen Instituts der Universität Köln.

Garcia, S., Angelier, J., Bergerat, F. and Homberg, C. 2002. Tectonic analysis of an oceanic transform fault zone based on fault-slip data and earthquake focal mechanisms: the Húsavík-Flatey Fault zone, Iceland. Tectonophysics 344, 157-174.

Harðarson, B.A. and Guðmundsson, Á., 1986. Stafnsvatnavirkjun. Mannvirkjajarðfræði. OS-86039/VOD-14, Orkustofnun, Reykjavik.

Hardarson, B.S., Fitton, J.G., Ellam, R.M. and Pringle, M.S., 1997. Rift relocation – a geochemical and geochronological investigation of a paleo-rift in NW Iceland. Earth and Planetary Science Letters, 153: 181-196.

Hjartarson, A., Friðleifsson, G.O. and Hafstað, Th.H., 1997. Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020 Orkustofnun, Reykjavík.

Hjartarson, A., and Hafstað, Th.H., 1999. Jarðfræði á Hofsafrétt í Skagafirði. OS-99031, Orkustofnun, Reykjavík.

Jakobsson, S.P., 1979. Petrology of recent basalts of the Eastern Volcanic Zone, Iceland. Acta Naturalia Islandica, 26: 103 p.

Jóhannesson, H., 1980. Jarðlagaskipan og þroun rekbelta a Vesturlandi. Natturufræðingurinn, 50: 13-31.

Jóhannesson, H., 1991. Yfirlit um jarðfræði Tröllaskaga (Miðskaga). Árbók FÍ, 1991: 39-56.

Jóhannesson,, H and Sæmundsson, K., 1998a. Geological map of Iceland 1:500.000. Bedrock Geology. Icelandic Institute of Natural History, Reykjavik (2nd edition)

Jóhannesson,, H and Sæmundsson, K., 1998b. Geological map of Iceland 1:500.000. Tectonics. Icelandic Institute of Natural History, Reykjavik.

Jones, S.M. 2003. Test of ridge-plume interaction model using oceanic crustal structure around Iceland. Earth and Planetary Science Letters 6564, 1-14 (in press)

Kaldal, I. and Vikingsson, S., 1978. Jökulsár í Skagafirði I. Jarðfræði. OS-ROD 7805, Orkustofnun, Reykjavík.

Karlsdóttir, R., Haraldsson, G.I., Ingimarsdottir, A., Guðmundsson, A. and Hafstað, Th.H., 1991. Skagafjörður. Jarðfræði, jarðhiti, ferskvatn og rannsóknarboranir. OS-91047/JHD-08, Orkustofnun, Reykjavík.

Kristjánsson, L. and Jónsson, G., 1998. Aeromagnetic results and the presence of an extinct rift zone in Western Iceland. J. Geodyn., 25: 99-108.

Lawver, L.A. and Müller, R.D., 1997. Iceland hotspot track. Geology 22, 311-314.

Líndal, J.H., 1941.Drangey og hvernig hún er til orðin. Náttúrufræðingurinn 11: 41-46.

Líndal, J.H., 1964. Með huga og hamri. Jarðfræðidagbækur og ritgerðir. Reykjavik.

Pjetursson, H., 1905. Om Islands geologi. Meddelser fra Dansk Geologisk Forening, 11: 1-104.

Schilling, J.-G., Sigurðsson, H. and Kingsley, R.H., 1978. Skagi and Western Neovolcanic Zones in Iceland: 2. Geochemical Variations. J. Geophys. Res., 83: 3983-4002

Sigurðsson, H., Schilling, J.G. and Meyer, P.S., 1978. Skagi and Langjökull Volcanic Zones in Iceland., 1. Petrology and Structure. J. Geophys. Res., 83: 3971-3982

Sæmundsson, K., 1974. Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. Geological Society of America Bulletin, 84: 495-504.

Sæmundsson, K., 1979. Outline of the geology of Iceland. Jökull 29, 7-28.

Vink, G.E., 1984. A hot spot model for Iceland and the Vöring Plateau. J. Geophys. Res., 89: 9949-9959.

Vilmundardóttir, E. G., Hafstað Th. H., Kaldal, I. and Víkingsson, S., 1997. Vestari-Jökulsá. Berggrunnur og jarðgrunnur. jarðfræðikortlagning 1997. OS-97068, Orkustofnun, Reykjavík.

Vogt, P.R., 1971. Astenosphere motion recorded by the ocean floor south of Iceland. Earth and Planetary Science Letters, 13: 153-16.

Vogt, P.R. and Avery, O.E., 1974. Detailed magnetic surveys in the Northeast Atlantic and Labrador Sea. Journal of Geophysical Research 79, 263-389.

^{40}Ar/^{39}Ar–Dates from the Skagafjörður Valleys, N–Iceland – Implications for Rift Relocations and the Deep-Sea Sedimentary Record

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Abstract

The Skagafjörður Valleys volcanic Pleistocene formations in North Iceland rest with a major angular unconformity on a Neogene volcanic pile. Here we discuss new ^{40}Ar/^{39}Ar dates from above and below the unconformity. The dates above the unconformity show that Pleistocene volcanism started at 1.7 – 1.6 Ma and became extinct in the late Pleistocene. A date from below the unconformity at its southernmost exposure north of the Hofsjökull glacier gave an age of 2.49 ± 0.07 Ma. The volcanic hiatus at this locality is therefore about 0.8 million years. The hiatus increases from south to north and in the lower Vesturdalur valley it is 7 million years. The Neogene pile is predominantly made of basaltic lavas with sedimentary interbeds and more rare layers of acid and intermediate rocks. The largest formation of the Neogene pile is the Tinná Central Volcano. The Skati rhyolite dome belongs to the volcano and it was formed during a very large explosive eruption accompanied by an extensive tephra fall. It is the most voluminous monogenetic rhyolite unit known in Iceland, corresponding to 18 km^3 of dense rock. A date from the Skati dome gave an age of 5.212 ± 0.016 Ma. The formation is correlated to a thick tephra layer found in ODP cores from site 907 NNE of Iceland. This is the first correlation between Icelandic Neogene eruptions and a deep-sea tephra layer. It is concluded that this date also reveals indirectly another unconformity lower in the pile, representing a rift relocation from the extinct Snæfellsnes – Húnaflói rift zone to the presently active North Iceland rift zone at 7Ma.

Introduction

The bedrock of the Skagafjörður valleys in central North Iceland have long been divided into two main successions (Kjartansson 1965). These are the Neogene and the Pleistocene successions, separated by a major Skagafjörður angular unconformity (Fig. 2). The geological history of the area is complicated because the lavas were erupted at different times in three volcanic rift zones. The earliest zone is the extinct Snæfellsnes–Húnaflói rift (Fig. 1). About 7 million years ago it was shifted eastwards by a rift relocation as the North Iceland rift zone became the main focus of spreading. In the early Pleistocene a premature or ephemeral rift zone became active in the Skagafjörður District (Fig. 1, Sigurðsson et al., 1978, Hjartarson, 2002). The volcanism partly filled up a Pliocene valley system that had been developing in the area, but became extinct in the late Pleistocene.
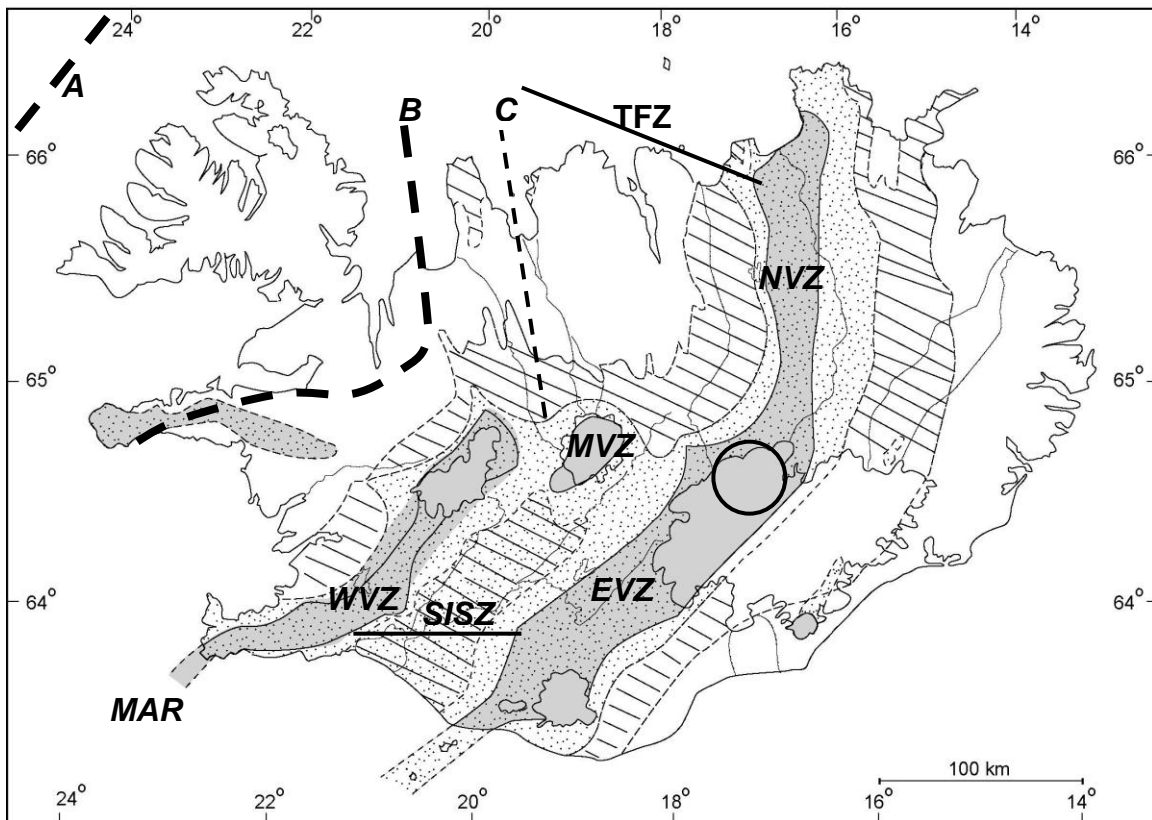
Fig. 1. The neovolcanic zones and rift systems in Iceland. A = Northwest rift axis; B = Snæfellsnes-Húnaflói rift axis; C = Skagafjörður rift axis; MAR = Mid-Atlantic Ridge; WVZ = Western Volcanic Zone; MVZ = Mid-Iceland Volcanic Zone; NVZ = Northern Volcanic Zone; EVZ = Eastern Volcanic Zone; SISZ = South Iceland Seismic Zone; TFZ = Tjörnes Fracture Zone. Circle represents the location of the Iceland plume. Shaded = The Neovolcanic zone. Dotted =Early Pleistocene rocks. Oblique lines = Plio – Pleistocene rocks (0.8-3.2 Ma).

Previous Studies

The age of the Skagafjörður unconformity and the duration of the volcanic hiatus it represents, has never been determined directly although several indirect assumptions have been made (Líndal 1964, Hjartarson et al. 1997). No radiometric dates on lava flows from the Skagafjörður valleys were available but a few from nearby areas in the district have been published. Because of the regional dip of the Neogene pile the age gap included in the unconformity is supposed to increase from south to north.
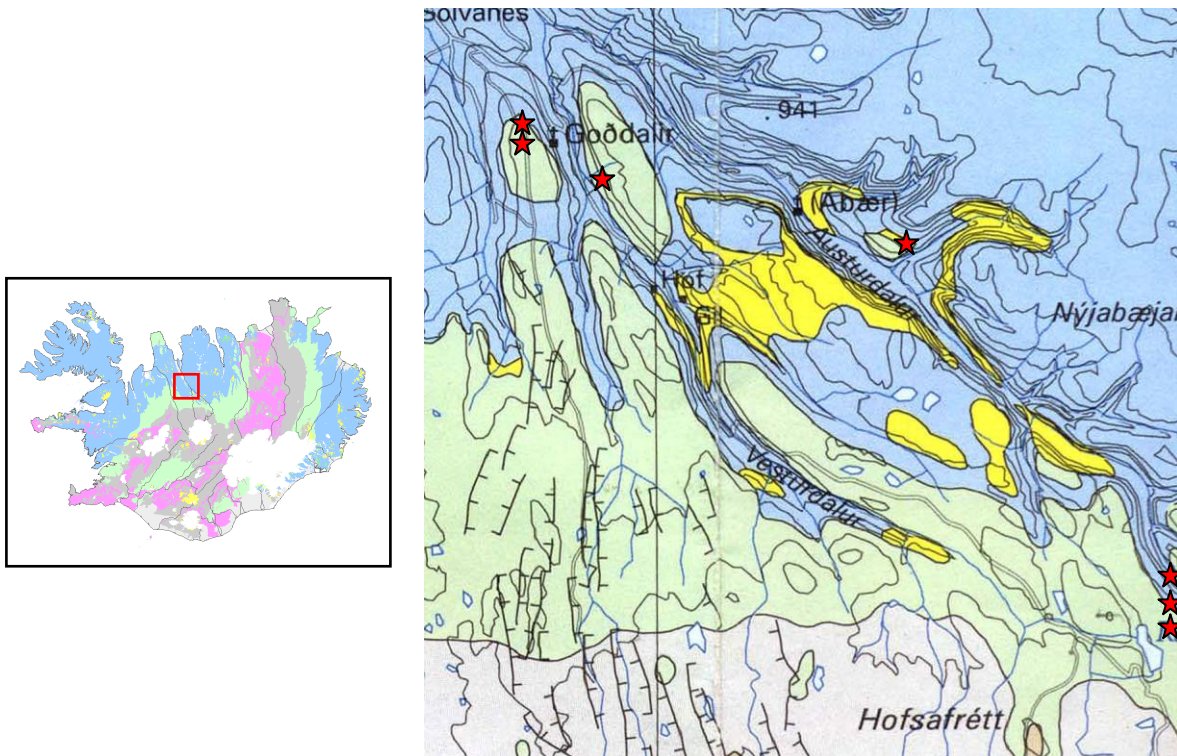
Fig. 2. Location map. Red stars mark the dated samples. Blue = The Neogene regions. Green = The Plio-Pleistocene areas (3.3-0.8 Ma). Gray = Late Pleistocene. Yellow = rhyolite. The unconformity is at the borderline between the Neogene and Plio-Pleistocene regions. (Modified from Jóhannesson and Sæmundsson, 1998).

Everts (1972) and Everts et al. (1975) published geological maps along with a detailed description of the northern part of Skagi peninsula and the east coast of Skagaförður, including geochemical analyses, paleomagnetic field measurements and K/Ar dates. The samples were collected from young bedrocks of the Skagi peninsula and by the east coast of Skagafjörður (Fig. 1). The dates show a time interval of 2.6 – 0.5 Ma, with wide errors for the older lavas (Table 1).

Sæmundsson et al. (1980) published a series of dates in an extensive geological and paleomagnetic study of Tröllaskagi, North Iceland (Fig.1). Their data set includes several dates from Mt. Sólheimafjall in central Skagafjörður District indicating an intensive and continuous build up of the Neogene lava pile between 9.7 – 8.7 Ma (Table 1).

Table 1. K/Ar dates from lava flows in the Skagafjörður area

Locality	Pol- arity*	Sample number.	K/Ar-age Ma	Ref.**
Króksbjarg	R	1	1.4 ± 0.4	2)
Tjarnarfjall	R	2	1.1 ± 0.5	2)
Tjarnarfjall	N	3	0.7 ± 0.4	2)
Ketubjörg	R	4	2.6 ± 0.8	2)
Hofsós	N	5	2.7 ± 1.2	2)
Þórðarhöfði	N	6	5 ± 5	2)
Sólheimafjall PF-48	N	78-1005	8.88 ± 0.12	1)
Sólheimafjall PF-43	N	78-1002	8.72 ± 0.10	1)
Sólheimafjall PF-39	A	75-144	9.04 ± 0.18	1)
Sólheimafjall PF-36	A	75-141	9.12 ± 0.12	1)
Sólheimafjall PF-34	R	75-140	9.16 ± 0.20	1)

* Paleomagnetic polarity; N = normal polarity, R = reverse polarity, A = anomalous polarity.

**1) Sæmundsson et al. 1980. 2) Everts 1975.

Sampling and Analytical Techniques

In this study several samples were collected from the Skagafjörður valleys and, after careful selection including microscopic and chemical analyses (Table 3), a subset of seven samples was selected for ^{40}Ar/^{39}Ar dating (Fig. 2). Four samples are from basaltic lavas located immediately above the Skagafjordur unconformity and two are from basaltic flow immediately below. One sample was selected from the Skati rhyolite dome, which is located deeper in the Neogene lava pile. The geographical distribution of the samples should reveal the predicted increase in age gap represented by the unconformity from south to north, namely from the highlands near Hofsjökull and northwards to the lower Vesturdalur valley (Fig. 2). The stratigraphical section across the unconformity at the Austurdalur valley is shown in Fig. 3.

Radiometric dates were obtained by ^{40}Ar/^{39}Ar incremental-heating analysis of holocrystalline whole-rock cores (5 mm in diameter) from which ~1 mm thick wafers were cut and subsequently irradiated. Details of analytical procedures are described by Singer and Pringle (1996). Reliability of the Ar-Ar ages is assessed using the following criteria, each of which involves a rigorous statistical test (Pringle *et al.*, 1996). We accept an apparent ^{40}Ar/^{39}Ar age as an accurate estimate of the crystallization age of a volcanic rock only if: (1) a well-defined, high temperature age spectrum plateau is formed by at least three concordant, contiguous steps representing at least 50% of the ^{39}Ar released,

(2) a well-defined isochron exists for the plateau points, (3) the plateau and isochron ages are concordant, and (4) the isochron $^{40}Ar/^{36}Ar$ intercept is not significantly different from the atmospheric composition. It may not be possible to determine why a particular experiment is unreliable, i.e., because of geological error, experimental artifact, or both. However, these internal tests can determine whether the K-Ar clock of a particular sample has been too disturbed to reveal a meaningful geological age. All the seven samples fulfill these requires. The main results are shown in Table 2 and the details in Table 5.
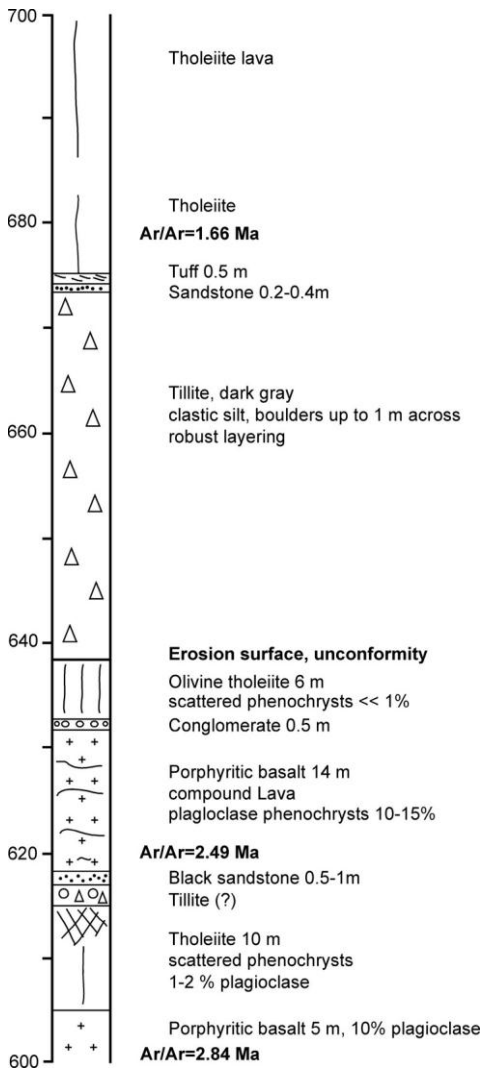
Fig. 3. A section across the unconformity at the Austurdalur valley near Geldingsá north of Hofsjökull. Here the unconformity represents a hiatus of about 0.8 million years.

Table 2. $^{40}Ar/^{39}Ar$ ages from the Skagafjörður valleys

No.	Locality (m a.s.l.)	Location	K_2O %	Pol-arity	$^{40}Ar/^{39}Ar$-age $Ma \pm 1\sigma$
17160	Vesturdalur, Djúpagil innra, 386 m. Above the unconformity	65°19.841' 19°06.764'	0.74	R	1.482 ± 0.060
17169	Vesturdalur, Hlídarfjall, 335 m. Above the unconformity	65°18.730' 190°3.540'	0.27	R	1.468 ± 0.017
17172	Goðdalakista, the topmost lava, 585 m. Above the unconformity	65°19.280' 19°08.054'	0.24	R	1.256 ± 0.016
17173	Austurdalur, near Geldingsá, 600 m. Below the unconformity	65°07.462' 18°27.610'	0.12	N	2.841 ± 0.043
17175	Austurdalur, near Geldingsá, 620 m. Below the unconformity	65°07.462' 18°27.610'	0.13	N	2.486 ± 0.068
17177	Austurdalur, near Geldingsá, 660 m. Above the unconformity	65°07.208' 18°27.238'	0.63	R	1.659 ± 0.024
17181	Skati dome, Tinnárdalur, 650 m. Below the unconformity	65°17.243' 18°44.291'	3.3	R	5.212 ± 0.016

Table 3. Chemical analyses of the dated samples

Lab. no	17160	17169	17172	17173	17175	17177	17181
Field no.	S-34	S-37	S-39	S-40	S-41	S-42	S-16
SiO_2	48.81	47.61	47.45	48.64	48.71	47.78	73.75
TiO_2	3.78	2.13	2,66	1.53	1,74	3.53	0.17
Al_2O_3	12.76	17.97	13,68	15.45	14.78	13.23	12.86
Fe_2O_3	16.63	12.81	14.76	11.33	13.26	17.16	2.21
MnO	0.27	0.18	0.22	0.18	0.21	0.24	0.05
MgO	4.75	5.4	7.13	7.8	6.91	5.27	0.05
CaO	9.06	11.35	11.71	13.18	12.32	9.29	1.2
Na_2O	2.84	2.65	2.31	2.16	2.33	2.82	4.54
K_2O	0.74	0.27	0.24	0.12	0.13	0.63	3.3
P_2O_5	0.65	0.17	0.21	0.12	0,14	0.38	0.02
LOI	0.09	0	0	0	0	0	1.9
Total	100.46	100.6	100.44	100.55	100.59	100.47	100.15

The age of the Skagafjörður volcanic zone

The dates from above the unconformity indicate that activity in the Skagafjörður volcanic zone started early in the Pleistocene, about 1.7 – 1.6 million years ago. The first two dates (no. 17160 and 17169 in Table 2) are from lavas resting directly on a sedimentary horizon within the unconformity in the outer Vesturdalur area. The results indicate that lavas were erupted along the valley near 1.50 Ma. According to date no. 17177 in Austurdalur (Table 2), 40 km farther inland, volcanism started there slightly earlier, or at 1.66 Ma. The dated samples show reverse polarity consistent with the mid-Matuyama geomagnetic field prevailing at that time (Cande and Kent 1995).

The two dates from Austurdalur below the unconformity, near the confluence with Geldingsá (no. 17175 and 17173 in Table 2), give 2.49 Ma and 2.84 Ma respectively or upper Pliocene age. They are normally polarized but seem to be from different magnetic subchrons. The upper one fits to the short Reunion II cryptochron (C2r.2r-1) and the lower one are from the end of the Gauss polarity subchron (C2An.1n) (Cande and Kent 1995).

These two layers have younger ages than previously accepted (Hjartarson et al. 1997). At this time the Snæfellsnes – Húnaflói rift zone had been extinct for several million years (Sæmundsson 1979, Jóhannesson 1980) and the Skagafjörður volcanic zone had not been initiated. This strongly suggests that these lava flows originate from the North Iceland volcanic zone and must have flowed about 30 km westwards out of the volcanic belt. This means that the lavas are rather large. A diamictite sedimentary horizon lies in between the dated lava flows. The sediment is interpreted as a tillite formation, suggesting glaciation in the area near to 2.5 Ma (Fig 3).

The Skati Rhyolite Dome

The date from the Skati rhyolite dome (no. 17181 in Table 2) gives an age of 5.21 ± 0.02 Ma. The Skati dome is the largest member of the so-called Tinná Central Volcano. The dome was formed during a very large explosive eruption associated with extensive tephra discharge and consists of an acidic lava and a tephra. It is the most voluminous monogenetic rhyolite unit known in Iceland, covering an area of 60 km², with a lava volume of about 8 km^3. The tephra layer corresponds to 10 km^3 of dense rock. The tephra, therefore, can serve as an important marker horizon in the Icelandic lava pile as well as in the deep-sea sediments (Fig. 4) around the island (Article 2 in this PhD-thesis).

There appears to be a discrepancy between the $^{40}Ar/^{39}Ar$-age of the Skati dome and its magnetic polarity. The polarity of the rock is reverse but at 5.21 Ma the Earth's magnetic field was normal, representing the Thverá subchron (C3r), the lowest normal interval of the Gilbert polarity chron. Before the Thverá subchron a rather long reverse subchron, representing the lowest part of Gilbert, 5.89 – 5.23 Ma existed (Cande and Kent 1995). The Skati dome is supposed to belong to the central or the lower part of this subchron (see Appendix, fig. 23) and its $^{40}Ar/^{39}Ar$-age is therefore slightly too low. A more credible age would be 5.5 Ma.

The Tinná volcano stayed active for at least three polarity subchrons and the Skati dome was formed during the middle one (C3r). The volcano was active during the polarity subchrons of C3An1n – C3r – C3n4n, between 6.14 and 4.98 Ma (Cande and Kent, 1995)

or for about 1.2 million years. This is consistent with the usual lifetime of central volcanoes in Iceland, which is in the order of 0.5 – 1 Ma. (Sæmundsson 1979, Guðmundsson 2000). After its extinction it was buried by younger lavas as it drifted away from the volcanic zone. The petrology and stratigraphy of the Tinná Volcano shows that it originated in a rift zone.
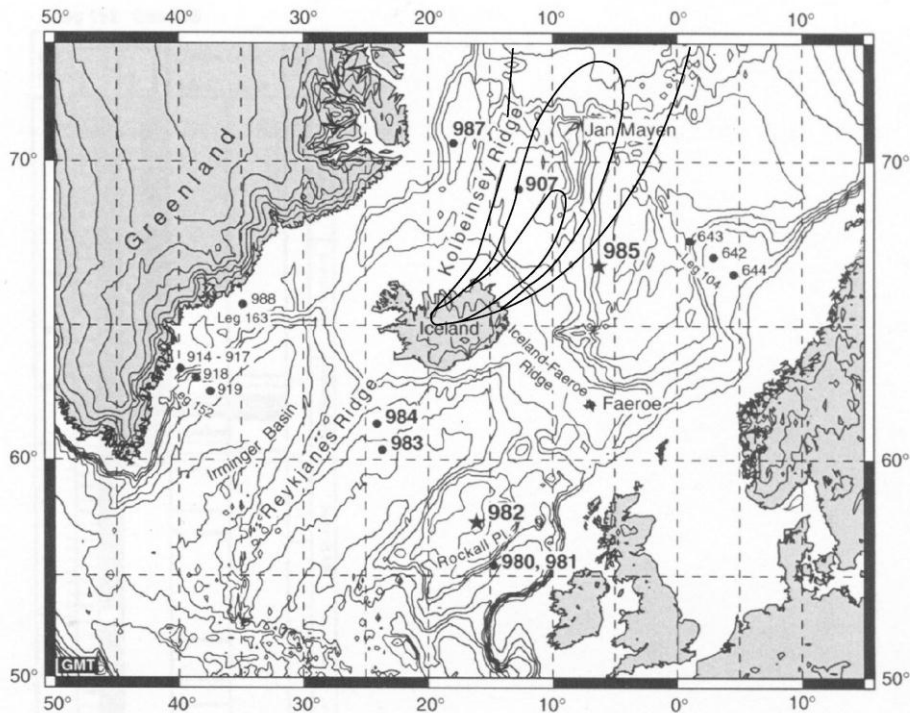
Fig. 4. Map of the North Atlantic showing the locations of ODP sites and suggested distribution of the Tinná tephra. The exact alignment of the tephra sector is unknown but the most probable alignment is shown here. The tephra sector is projected on a map of Iceland and the North Atlantic as they look like today, not as they were 5.5 Ma.

A possible identification of the Skati Tephra in ODP Cores

It is likely that the Skati tephra may be recognised in ODP cores from the ocean floor around Iceland, especially east and northeast of the island. Lacasse and Schönberg (2001) studied explosive volcanism in Iceland and the Jan Mayen area during the last 6 million years by its fingerprints in the deep-sea sedimentary record. They found that 90% of all identified tephra layers were recognised in ODP sites 907 and 985, which are located NNE and NE of Iceland in a downwind direction (Fig. 4). The cores have nearly 100% recovery and their description of the earlier part of the period is based on data from these sites. There should be a high probability of finding the Skati tephra there. One layer seems to be a possible candidate for the Skati tephra and if it occurs in the data collection at all, it is actually the one. The layer is the oldest tephra in the collection. Its age is

estimated to be 5.48 Ma from the paleomagnetic chrons and isotopic stage boundaries identified along the sequence of the ODP cores according to Lacasse and Schönberg (2001). They do not suggest where the tephra originates, but state that chemical characteristics strongly indicate that it came from a spreading zone in Iceland. They specifically refer to the tephra as: "Dispersal, thickness and grain size of the tephra layer clearly indicate that it was likely derived from one of the largest Icelandic explosive eruptions that ever occurred in the late Tertiary".

ODP site 907 is in the Arctic Ocean NNE of Iceland, 550 km away from the Tinná volcano. The distance has not changed considerably during the last 5.5 Ma. The holes are at 1800 m below sea level and the tephra layer is at 85 m depth in the bottom sediment (Fig. 4, see also Lacasse et al. 1996).

This is the first reliable correlation between Icelandic Neogene eruptions and a deep-sea tephra layer. Dates on lava and tephra layers in Iceland are likely to improve significantly interpretations of sediment cores recovered by ocean drilling projects and finally result in a deep-sea tephrachronological time scale (see e.g. Knudsen and Eríksson 2002 for Holocene and late glacial times).

The Hidden Unconformity

We have shown that the age of the Tinná Central Volcano is younger than expected. Based on stratigraphy and paleomagnetic studies, the Skati dome and the Tinná volcano were suggested to have been active at 7–8 Ma (Hjartarson et al., 1998). Sæmundsson et al. (1980) published K/Ar ages around a polarity reversal in Mt. Sólheimafjall in the central Skagafjörður District. Their results gave an age of approximately 9.1±0.2 Ma (Table 1) and the reversal can be correlated with the 9.3 Ma reversal C4Ar1n/C4Ar2r in the time scale of Cande and Kent (1995). The reversal is therefore roughly 3 million years older than the Skati dome. Only a 1000 m thick lava pile separates these formations. Hence, the average accumulation rate would be 330 m/Ma. That is in striking contrast to the growth rate of the lava pile in Mt. Sólheimafjall where it is 3850 m/Ma, or a factor of ten higher, as reported by Sæmundsson et al. (1980). This indicates there is a hidden hiatus in the lava pile between the reversal and the Skati dome. Palaeomagnetic results suggest the same. Published polarity time scales show 19 reversals during this interval, while only 9 are found in the field (Article 4 in this thesis). Moreover, if the history of rift relocations in Iceland has been interpreted correctly (Table 4), a major unconformity should be present in this 9–5 million years old lava pile.

Rift Relocations

A dominant process in the evolution of Iceland is the repeated eastward relocation of the spreading axis as it responds to the westward migration of the Eurasian-American plate system away from the Iceland mantle plume centre (Fig. 1). During rift relocations (or rift jumps) the active spreading axis at the plate boundary is displaced and shunted for tens or even over hundred km in a short period of time (Sæmundsson 1979).

Several rift relocations are known in the geological record of Iceland (Table 4). The Northwest rift zone is thought to have formed some 24 Ma west of the Northwest

Peninsula (Harðarson et al. 1997). It was active for 8-10 million years but then the focus of spreading was relocated to the east forming the Snæfellsnes – Húnaflói rift zone, which was active for another 8-10 million years. However, about 6 million years ago active spreading was relocated to its present location forming the Reykjanes – Langjökull zone and the North Iceland Zone (Sæmundsson 1974, 1979, Johannesson, 1980, Kristjánsson and Jónsson, 1998, see Fig. 1). The most recent rift relocation in North Iceland occurred when the Skagafjörður rift zone became active at 1.7 – 1.6 Ma and a temporary rift axis generated for about million years (Article 4 in this thesis). In South Iceland the picture is more complicated as the Eastern volcanic zone (Fig. 1) seems to be an evolving spreading axis (Sæmundsson 1979). It was initiated at 2-3 Ma and is slowly propagating to the southwest taking over the spreading of the Reykjanes – Langjökull segment.

Table 4. Rift zones and rift relocations

Rift zone	Time of initiation Ma	Ref.
Northwest Iceland rift zone	24	Hardarson et al. 1997
Snæfellsnes – Húnaflói rift zone	15	Hardarson et al. 1997
Reykjanes-Langjökull – N-Iceland rift zone	6 – 7	Sæmundsson 1979
Eastern rift zone	2 – 3	Sæmundsson 1979
Skagafjörður rift zone	1.6 – 1.7	Hjartarson 2003

Garcia et al. (2003) state that a paleo-rift, located some 60 km east of the Snæfellsnes – Húnaflói rift zone, was active in the Skagafjörður region between 8.1–2.9 Ma, and the latter one was only a short lived spreading centre, if it existed at all. This suggestion is based on new Ar/Ar dates of dykes but most age determinations in the Neogene and early Pleistocene regions of Iceland have been made on lavas or hyaloclastite formations. Garcia et al. suggest that when a study focuses on spreading rates and rift jumps, dates from dykes are more reliable than dates from lavas because the lavas might have flowed tens of km from its magmatic source. But, by using dykes they loose all stratigarphical control, which is a weakness in their method. Another disadvantage in their work is how their alternative spreading model contradicts the main patterns in the general stratigraphy of the country (Jóhannesson and Sæmundssom 1998) and the paleomagnetic time scale for North Iceland (Sæmundsson et al. 1980, Kristjánsson et al. 1992). Moreover they do not account sufficiently for the possibility that dykes can be injected into the crust outside the volcanic zones (Walker 1974, Jancin et al 1985). However, it is interesting to see very young dates (< 2 Ma) on dykes in the Skagafjörður region. They are problematic in the argumentation of Garcia et al. but fit well into the spreading history as it is introduced here because they seem to belong to the young Skagafjörður rifting (Table 4).

Discussion

The 9.3 Ma lavas at the reversal in Mt. Sólheimafjall were erupted before the North Iceland rift zone was initiated. They are thought to have originated in the Snæfellsnes – Húnaflói rift zone (Sæmundsson et al., 1980). The Skati dome and the Tinná Central Volcano, on the other hand, seem to originate in the North Iceland rift zone. This is supported by the fact that during the life span of the volcano, at 5-6 Ma, activity in the Snæfellsnes – Húnaflói rift zone had ceased, or was waning, 80 km to the west. The age results from the Skati rhyolite dome (no. 17181, Table 2) therefore require a significant unconformity within the lava pile between the Tinná volcano and the reversal in Mt. Sólheimafjall, indicating a major rift relocation from the Snæfellsnes-Húnaflói zone to the North Iceland zone. This unconformity would correlate with the major unconformities of Fnjóskadalur and the Borgarnes anticline (Sæmundsson, 1979). This unconformity has not been found in the field and here it will therefore be called the Hidden Unconformity. Two recognized sedimentary horizons in the area may represent a significant unconformity. These are the Merkidalur sedimentary layer and the Tinná lignite sediment one of which could represent the unconformity (See the included Geologic Map of the Skagafjörður Valleys). There is little difference between the strike and dip of the layering above and below the unconformity. The regional dip in the area is $5 - 10°$ S with variations towards SSE and SSW. In a limited area in the outer Austurdalur and Vesturdalur valleys dips up to $12°$ have been recognized. They may indicate a flexure zone related to the Hidden Unconformity, such flexures are known to accompany unconformities in East, North, West and Northwest Iceland (Sæmundsson 1979, Jancin et al. 1985, Jóhannesson 1980, Hardarson et al. 1997).
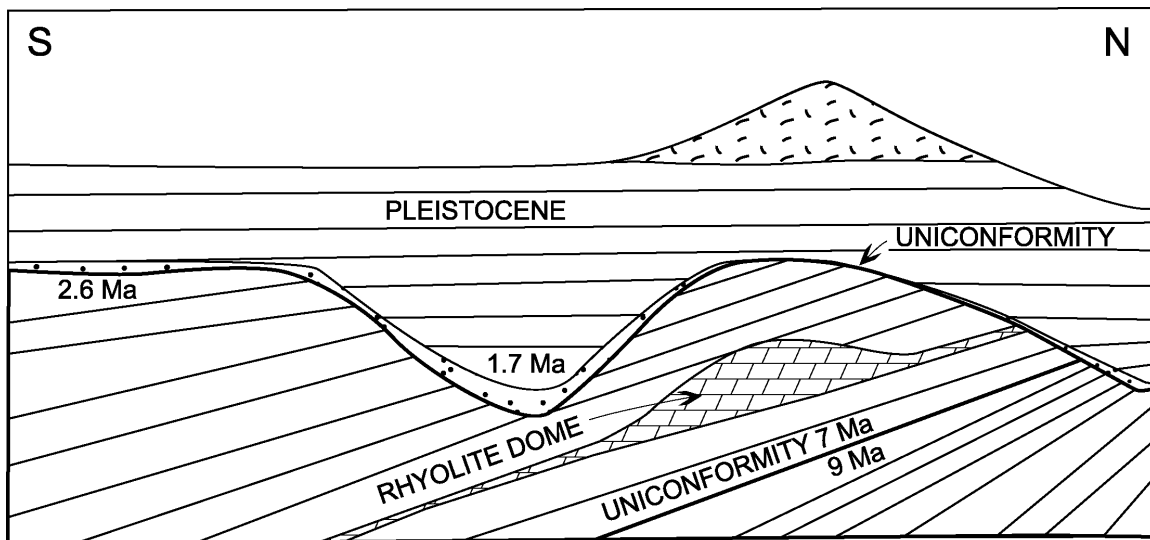
Fig. 5. A schematic cross section indicating the two unconformities in the stratigraphy of Skagafjörður Valleys. Horizontally layered Pleistocene succession on top. Skagafjörður Unconformity outlines an ancient valley system in the middle. The dipping Neogene succession is at the bottom including the Skati rhyolite dome and the Hidden Unconformity.

The time span represented by the Hidden Unconformity is unclear but by comparison with the magnetostratigraphy described by Hjartarson et al. (1998) it seems to span the interval 9–7 Ma in the outer Skagafjörður valleys. The hiatus represented by the Skagafjörður Unconformity is, however, more clear. It increases from south to north due to erosion and tilting of the underlying Neogene pile, as illustrated in Fig. 5.

The $^{40}Ar/^{39}Ar$ dates of lavas no. 17175 and 17177 (Table 2) are from the same section (Fig. 3) and give the maximum time gap of the unconformity at its southernmost exposure. There it is approximately 0.8 million years. In the outer Vesturdalur Valley, above the Hidden Unconformity it may be 5 Ma but immediately below it, it still increases to 7 Ma.

The main conclusions of the present study show that two major unconformities are found in the stratigraphy of the Skagafjörður valleys representing two dramatic events in the geological history of Iceland. The Hidden Unconformity represents the major rift relocation from the Snæfellsnes–Húnaflói rift axis to the North Iceland rift axis at 6 – 7 Ma. The Skagafjörður Unconformity, was formed during the incomplete rift relocation from the North Iceland axis to Skagafjörður that occurred at 1.7 Ma.

Acknowledgements

References

S.C. Cande and D.V. Kent, Revised calibration of the geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. J. Geophys. Res., 100 (1995) 6093-6095.

P. Everts, Die Geologie von Skagi und der Ost-Küste des Skagafjörðurs (Nord-Island). Sonderveröffentlichung des Geologischen Instituts der Universität Köln (1975) 114 pp + maps.

P. Everts, L.E. Koerfer and M. Swartzbach, Neue K-Ar-Datierungen islandischer Basalte. Neues Jahrbuch Geol. Palaeontol. Monatsh.5 (1972) 280.

S. Garcia, A.O. Arnaud, J. Angelier, F. Bergerat and C. Homberg, Rift jump process in Northern Iceland since 10 Ma from 40Ar/39Ar geochronology. EPSL 214 (2003) 529-544.

Á. Guðmundsson, Dynamics of Volcanic Systems in Iceland: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge System. Annual Review of Earth Planet. Sci., 28 (2000) 107-140.

B.S.Harðarson, J.G. Fitton, R.M. Ellam and M.S. Pringle, Rift relocation – a geochemical and geochronological investigation of a paleo-rift in NW Iceland. Earth Planet. Sci. Lett., 153 (1997) 181-196.

Á. Hjartarson, The Skagafjord Ephemeral Volcanic Zone. In: Jónsson, S. S. (ed.), The 25th Geological Nordic Winter Meeting. Januray 6th – 9th 2002, Reykjavík, Iceland. Abstract Volume.

Á. Hjartarson, The Skagafjörður Unconformity, North Iceland and its Geological History. PhD-thesis, University of Copenhagen (2003).

Á. Hjartarson, G.Ó. Friðleifsson and Th.H. Hafstað, Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020, National Energy Authority, Reykjavík, 55 pp (1997).

M. Jancin, D.K. Young, B. Voight, J.L. Aronson and K. Sæmundsson, Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. J. Geophys. Res., 90 (1985) 9961-9985.

H. Jóhannesson, Jarðlagaskipan og þróun rekbelta á Vesturlandi. Náttárufræðingurinn, In Icelandic with English Abstract, 50 (1980) 13-31.

H. Jóhannesson and K. Sæmundsson, Geological map of Iceland 1:500.000. Bedrock Geology. Icelandic Institute of Natural History, Reykjavik (2nd edition,1998).

K.L. Knudsen and J. Eiríksson, Application of tephrochronology to the timing and correlation of palaeoceanographic events recorded in Holocene and Late Glacial shelf sediments off North Iceland. Marine Geology, 191 (2002) 165-188.

L. Kristjánsson, H. Jóhannesson, and I. McDougall, Stratigraphy, age and paleomagnetism of Langidalur, Northern Iceland, Jökull, 42 (1992) 31-44.

L. Kristjánsson and G. Jónsson, Aeromagnetic results and the presence of an extinct rift zone in Western Iceland. J. Geodyn., 25 (1998) 99-108.

C. Lacasse, M. Paterne, R. Werner, H.-J. Wallabre-Adams, H. Sigurðsson, S. Carey and G. Pinte, Geochemistry and origin of Pliocene and Pleistocene ash layers from the Icelandic plateau, site 907. p. 309-331, in: J. Thiede et al. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results, Vol. 151, (1996) 309-331.

C. Lacasse and C.-D. Grabe-Schönberg, Explosive silicic volcanism in Iceland and Jan Mayen area during the last 6 Ma: sources and timing of major eruptions. J. Volcanol. Geotherm. Res., 107 (2001) 113-147.

J.H. Líndal, J. H., [1964]: Með huga og hamri. Jarðfræðidagbækur og ritgerðir. 410 p.

M.S. Pringle B.S. Singer, Recent advances in ^{40}Ar/^{39}Ar geochronology of Quaternary volcanic rocks. Eos, Transactions, American Geophysical Union, 77 (1996) 91

H. Sigurðsson, J.G. Schilling and P.S. Meyer, Skagi and Langjökull Volcanic Zones in Iceland., 1. Petrology and Structure. J. Geophys. Res., 83 (1978) 3971-3982.

B.S. Singer and M. Pringle, Age and duration of the Matuyama – Brunhes geomagnetic polarity reversal from ^{40}Ar/^{39}Ar incremental heating analyses of lavas. Earth Planet. Sci. Lett. 139 (1996) 47-61.

K. Sæmundsson, Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. Geol. Soc. Am. Bull., 84 (1974) 495-504.

K. Sæmundsson, Outline of the geology of Iceland. Jökull 29 (1979) 7-28.

K. Sæmundsson,L. Kristjánsson, I. McDougall and N.D.Watkins, K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. J. Geophys. Res., 85 (1980) 3628-3646.

G.P.L. Walker, The structure of Eastern Iceland. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, (1974) 155-164.

Table 5. Summary of Ar-Ar step-heating experiments on whole rock basalt samples from Northern Iceland.

Sample	K/Ca (total)	Total-fusion Age Ma	± 1σ	Increments (N of total)	^{39}Ar (%)	Plateau Age Ma	± 1σ	MSWD	Isochron Age Ma	± 1σ	^{40}Ar/^{36}Ar intercept	± 1σ	SUMS (N-2)
Blanda	0,07	1,27	0,02	11 of 15	92,9	**1,264**	**0,010**	0,49	1,282	0,016	291,1	4,4	0,45
Quarry	0,07	1,24	0,06	10 of 16	81,7	**1,253**	**0,013**	2,45	1,259	0,013	293,0	3,7	3,13
17172	0,06	1,21	0,02	9 of 16	79,5	**1,226**	**0,016**	3,12	1,120	0,051	319,3	10,8	2,46
	0,06	1,20	0,02	10 of 14	63,9	**1,285**	**0,029**	4,11	1,299	0,038	290,4	13,0	5,46
17169	0,07	1,41	0,03	9 of 14	77,1	**1,432**	**0,021**	1,21	1,432	0,027	295,3	3,3	1,37
	0,11	1,55	0,02	10 of 15	81,6	**1,503**	**0,026**	6,46	1,396	0,044	316,4	7,3	3,61
17160	0,50	2,27	0,09	8 of 16	60,1	1,664	0,045	3,36	**1,482**	**0,060**	299,6	1,4	2,51
	0,10	4,89	0,16	9 of 17	77,9	2,359	0,226	7,94	0,886	0,290	304,9	3,4	5,55
17177	0,23	1,63	0,10	14 of 19	89,9	**1,641**	**0,035**	10,41	1,642	0,049	297,0	2,5	17,30
	0,14	1,62	0,04	11 of 16	76,1	**1,677**	**0,032**	5,53	1,620	0,043	301,3	2,9	5,60
17175	0,02	2,23	0,07	9 of 14	78,3	**2,375**	**0,084**	6,05	2,675	0,163	291,6	1,8	6,10
	0,03	2,68	0,09	9 of 12	92,3	**2,597**	**0,107**	5,06	2,820	0,199	292,5	2,8	8,77
17173	0,01	2,95	0,18	7 of 14	78,3	**2,900**	**0,063**	1,99	2,771	0,075	297,0	1,6	3,10
	0,01	2,40	0,09	7 of 14	61,9	**2,782**	**0,060**	2,67	2,537	0,101	299,3	1,4	2,37
GR1	0,10	3,59	0,02	11 of 17	81,5	**3,644**	**0,030**	10,50	3,628	0,044	297,4	2,7	14,10
	0,05	3,32	0,05	9 of 19	56,0	**3,595**	**0,030**	8,66	3,529	0,090	299,9	5,0	11,30
17181	3,54	5,21	0,02	19 of 26	86,6	**5,212**	**0,016**	1,89	5,226	0,023	291,9	4,3	1,96

Postglacial lava production in Iceland

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Abstract. All known postglacial lavas of Iceland ≥ 1 km³ are listed and labeled by age and volume. A plot showing the lava production per millennium reveals two pulses. The first and larger of these occurred 9,000-7,000 BP and the second extends over the last 2,000 years. When eruptions of lava shields and of crater rows are examined individually it becomes evident that they have different periods of activity. Most of the largest lava shields in Iceland were erupted in the first millennia after deglaciation. During the last 4,000 years, only three large lava shields have been erupted. The crater rows, on the other hand, do not show any clear relations to the deglaciation. Their lava production is marked by two pulses, the first 9,000-8,000 BP and the other during the last 1100 years. Between the production highs there is a prominent low, the mid-Holocene low, a 2,000-year period when no large fissure eruptions took place. The explosive plinian eruptions in the central volcanoes of Iceland exhibit a similar pattern to the crater rows, with greater activity in the early and late Holocene than in the mid Holocene but in the long run their frequency has increased with time. Glacier unloading and crustal rebound seem to stimulate eruptions of lava shields but their influences on crater rows and central volcanoes are small if any.

Keywords: Lava volume – Lava shields (shield volcanoes) – Crater rows – Plinian eruptions – Production rate – Iceland

Introduction

The aim of this paper is to evaluate the lava production of postglacial volcanism in Iceland and examine possible relationships between the production rate and deglaciation and ice unloading. Stratigraphic investigations in the Skagafjörður Valleys, outside the active volcanic zones of Iceland, revealed that compound lavas and lava shields are rare in the Neogene strata pile, but become abundant in the Pleistocene pile after the onset of glaciation (Article 4 in this thesis). This led to the following discussion on possible interaction between volcanism and the glacial isostasy together with rapid ice unloading.

Basaltic volcanism in Iceland is mainly confined to two types of volcanoes: crater rows and lava shields. Both types erupt only once and thus differ from central volcanoes with their repeated activity and varying magma production. The crater rows vary in length, from only a short fissure with a single crater up to 80 km, comprising hundreds of eruptive vents. The eruptions are accompanied by rifting episodes and can be hazardous, especially at the beginning. Sometimes they are associated with tephra falls, but they usually diminish with time and are rarely long-lasting (more than 1 year).

Table 1. Iceland's largest postglacial lavas in volumetric order (≥ 1 km³)

Name	Type[1]	Age [2]	Estim. age [3]	Area km²	Volume km³	Ref.
Þjórsárhraun THb	F	7800 ^{14}C	8600	967	25	1
Eldgjá	F	AD 934 H, I	1070	780	19.6	2, 16
Stóravítishraun	S	11080-11980 T	11300	460	18.4	7, 21
Skjaldbreiður	S	9600 S	9600	200	17	3, 4, 20
Trölladyngja	S	> 7000 S	7250	340	15	4
Laki	F	AD 1783 H	220	600	14.7	5
Kollóttadyngja	S	> 4500 S	8000	69.1	14.5	6
Skildingahraun	S	>11980 T	12200	250	10	7,25
Bárðardalshraun	F	> 10300 T	10300	450	8	7
Búrfellshraun Thi	F	3200 T	3200	440	6	9
Lambahraun	S	3700 ^{14}C	4100	160	6	7, 20
Heiðin há	S	7500 S	7500	150	6	8
Kerlingardyngja	S	> 4000 ?	6000		6	10
Kjalhraun	S	7800 T	7800	150	6	7, 20
Þráinsskjöldur	S	~12500 S	13000	130	5.2	11, 23
Hólmsárhraun	F	6800 ^{14}C	7600		5	2
Sandfellshæð	S	~12500 S	12500	120	4.8	11, 23
Þingvallahraun	S	9130 ^{14}C	10200	200	4	3
Ketildyngja – Laxárhr. E.	S	4300 T	4300	300	4	12, 19
Kinnarhraun	F	>10300	10800	250	4	7
Tungnárhraun TH-d	F	6000-7000 S	7000	270	3.8	13
Sigölduhraun TH-f	F	4000 S	3900	200	3.4	13
Hallmundarhraun	S	1100 T, ^{14}C	1100	225	3.4	7, 22
Hrútagjárdyngja	S	4000-5000 T	4500	80	3.2	11
Leitahraun	S	4560 ^{14}C	5200	100	3	11
Flatadyngja	S	3500-4500 T	4400	110.5	2.6	6
Laxárhraun yngra	F	2000 ^{14}C, T	2100	220	2.5	12, 19
Selvogsheiði	S	8200 S	8200	50	2.2	11
Frambruni	F	700 T	700	200	2	18
Gjástykkisbunga	S	11080-11980TS	11500	50	2	28
Búrfellshraun (Mývatn)	F	>2900T	3500	100	1.8	7, 25
Hekla 1766	F (C)	AD 1766 H	240	65	1.7	14
Litladyngja	S	2900-3500 T	3200	85.1	1.7	6
Rauðhólar – Sveinar	F	9500S	9500	150	1.5	24
Hágönguhraun	F	> 7000 S	8500	100	1.5	7
Veiðivötn	F	AD 1477 H	500		1.5	13
TH-c	F	6800 S	6800	120	1.4	13
Vatnaöldur	F	1100 T, I	1100		1.2	13
Hellisheiðarhraun A	F	9240 ^{14}C	10440	32	1.1	26,27
TH-e	F	6400 S	6400	260	1	13
Skuggadyngja	S	>4500 T	5900		1	25
Svartadyngja	S	> 4500 S	6800	38.7	1	6
Útbrunahraun	S	>10300	10500	85	1	7
Kröfluháls	F	>10000 S	10600	60	1	28
Surtsey	S	AD 1964 H	30		1	15

1) F = Fissure eruption, S = Lava shield, (C = Lava from a central volcano). 2) Measured age, ^{14}C = radiocarbon, T = tephrachronology, S = stratigraphy, I = ice core, H = historical, 3) Corrected and estimated ages are in calendar years before AD 2000. **References:** see next page.
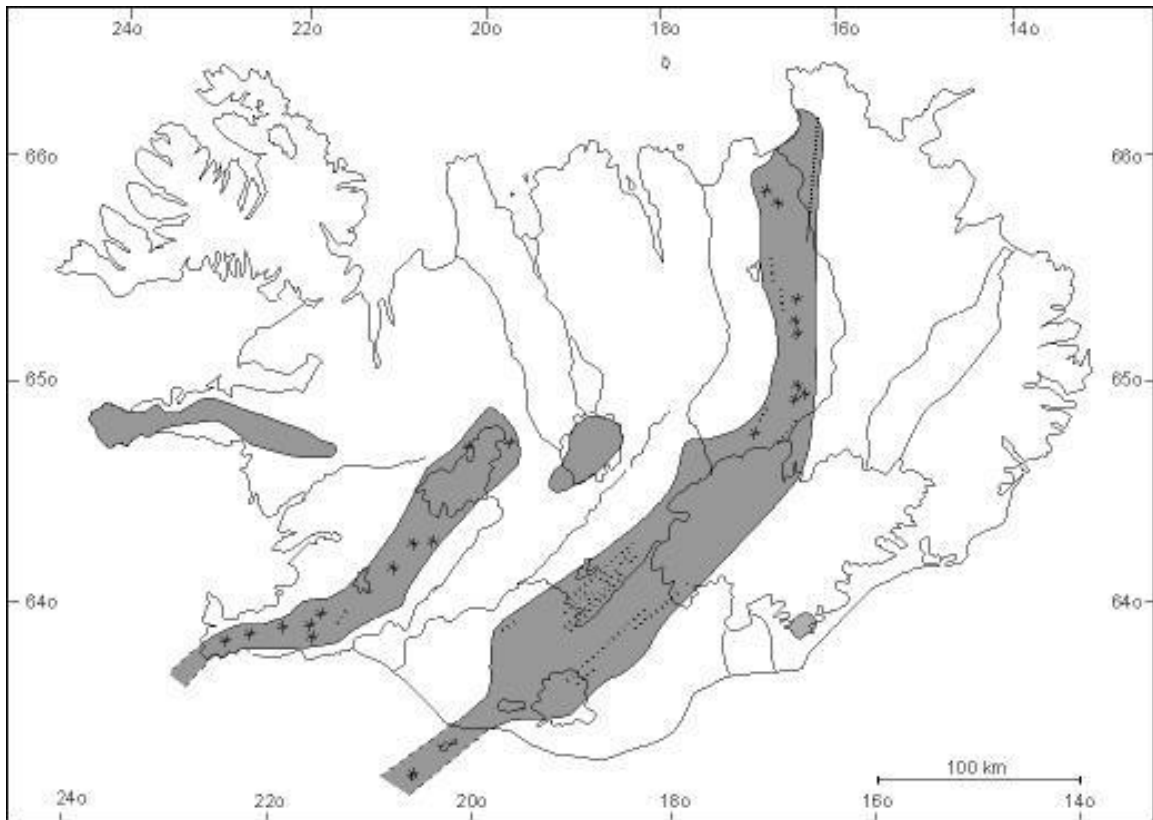
Fig. 1. Locations of the large lava eruptions (≥ 1 km³) in Iceland listed in Table 1. Lava shields are indicated by asterisks, crater rows by dotted lines. The grey areas are the active volcanic zones.

References for Table 1.

1 Hjartarson 1994, 2 Larsen 2000, 3 Sæmundsson 1992, 4 Kjartansson 1966, 5 Thordarson and Shelf 1993, 6 Sigvaldason et al. 1992, 7 Hjartarson unpubl., 8 Hjartarson 1999, 9 Vilmundardóttir et al. 1988, 10 Gudmundsson 2001, 11 Jónsson 1978, 12 Thorarinsson 1951, 13 Vilmundardóttir 1977, 14 Hjartarson 1995, 15 Thorarinsson 1968, 16 Thordarson et al. 2001, 17 Sigvaldason 2002, 18 Sigvaldason 1992, 19 Thorarinsson 1979, 20 Sinton and Grönvold 2001, 21 Sæmundsson 1973, 22 Jóhannesson 1989, 23 Sæmundsson 1995, 24 Sigurðsson 1975, 25 Sæmundsson pers. comm. 26 Jónsson 1989, 27 Kristjánsson et al. 2003, 28 Sæmundsson 1991.

Table 2. Holocene and late-glacial acid, plinian eruptions in Iceland

Eruption	Age	Age determination	Calendar age BP	Vol. (km³ magma)	Ref.
Askja 1875	125	Historical	125		1
Öræfajökull 1362	700	Historical	700	2	2
Hekla 1104	900	Historical	900		3
Snæfellsjökull	1750 +/- 150	^{14}C	1750		4, 7
Hekla H3	2820 +/- 70	^{14}C	2900	2.2	3
Hekla HSv	3600	Tephrachronology	3600		8
Snæfellsjökull	3960 +/- 100	^{14}C	4420		4, 7
Hekla H4	4030 +/- 120	^{14}C	4500	1.8	3
Hekla H5	6185 +/- 100	^{14}C	7000		3
Snæfellsjökull	8000	Tephrachronology	9000		4, 7
Askja (S)	11080 +/- 900	Tephrachronology	11080	1.5 +/- 0.5	6
Vedde (Z1)	11980 +/- 80	Ice core	11980	> 1	5

References: 1 Sigvaldason et al. 1992, 2 Thorarinsson 1958, 3 Larsen and Thorarinsson 1977, 4 Steinþórsson 1967, 5 Grönvold et al. 1995, 6 Sigvaldason 2002, 7 Flores et al. 1981, 8 Hjartarson 1995

The lavas consist of tholeiite (or transitional alkali basalt – alkali olivine basalt). The lava shields (shield volcanoes) are generally thought to be monogenetic (Walker 1971, Jakobsson et al. 1978, Rossi 1996) and most often emitted from a single circular crater. Rifting is rarely recognized. Their eruptions are thought to have been long-lasting but of low effusion rate. Practically no tephra is believed to have been produced. The lavas are of olivine tholeiite or picrite composition and occur mainly at the margins or outside the fissure swarms associated with the central volcanoes. It has been noted that many of the postglacial lava shields of Iceland are of early Holocene or late glacial age (Thorarinsson et al. 1959, Sigvaldason and Steinþórsson 1974, Jakobsson et al. 1978) and some relationship with glacial isostasy, together with rapid ice unloading, has been suggested (Gudmundsson 1986).

Earlier observations

Compound lavas from lava shields make up a considerable part of the total postglacial lava volume in Iceland. The ratio of the compound lavas to the total volcanic lava production has only been estimated for certain limited areas. Jakobsson et al. (1978) describe two volcanic systems on the western Reykjanes peninsula where they distinguish 56 lavas altogether: 42 fissure lavas and 14 lava shields. The volume of the fissure lavas is 3.2 km^3 and that of the lava shields 9.9 km^3. Consequently the crater rows are three times as numerous as the lava shields, and the lava shields are three times as voluminous as the crater rows. It is noteworthy that 67% of the total lava volume stems from two large eruptions, the Sandfellshæð and Þráinsskjöldur shield eruptions.

Gudmundsson (1986) examined 127 postglacial lavas on the Reykjanes Peninsula (including the area covered by Jakobsson et al. 1978). They belong to three volcanic systems and their total volume is 40-42 km^3. He found that the volume of the shields (29 km^3) is up to three times greater than that of the fissure lavas (11 km^3) and the largest

shields are of an order of magnitude larger than the largest fissure lavas. Thus, although the fissures are far more numerous than shields, their total volume is much smaller. The proportion of the shields is ¾ or 75%. Additionally he reports that the production rate appears to have diminished during the later part of the postglacial period. This is due to the early formation of most of the lava shields and a low lava production in historical times, i.e. only 2.3 km^3 (in several eruptions between AD 940 and 1340). However the fissure lava/compound lava ratio differs greatly between areas. In the Eastern Volcanic Zone and the Snæfellsnes off-rift zone, compound lavas are virtually non-existent.

Sigvaldason et al. (1992) studied the postglacial lava production rate of the Dyngjufjöll area, Central Iceland. From their areal and volumetric measurements, the ratio of the compound lavas seems to be close to 50%.

They also found a very high production rate for the Dyngjufjöll volcanic system, compared to the present productivity, in the first millennia following the last deglaciation. Magma production in the period 10,000-4,500 years BP was 20-30 times higher than in the period since 2900 years BP. This high production rate coincides with the deglaciation of the area and they assume that rapid isostatic rebound and vigorous crustal movements, together with decreasing lithostatic pressure, might have triggered intensive volcanism (See also Sigvaldason 2002).

Vilmundardóttir and Larsen (1986) found a similar pattern in the volcanic development of the Veiðivötn system where half of the lava volume erupted during postglacial time was produced in the period 10,000-8,000 years BP.

Fig 2. Volume distribution of 44 large Icelandic postglacial lavas (< 15,000 years) according to Table 1. The volume distribution between lavas from crater rows and lava shields seems to be even. Grey = fissure lavas from crater rows, white = compound lavas from lava shields.

The large Lavas

The studies mentioned above each cover one to three volcanic systems. If postglacial volcanism in Iceland is viewed as a whole, the picture is likely to become more general. The main problem is, however, that information about ages and lava volumes is in most cases uncertain and the uncertainty increases with increasing age. Large lavas have received more attention than small or intermediate ones and it is also, in general, easier to define their size, volume and age. The present investigation will focus on large lavas. Although such a focus inevitably leaves out some significant lavas which might affect the general trend, it is considered to give optimum results, and nonetheless a considerable fraction of total postglacial lava production is accounted for. Jakobsson (1972, 1979b) has estimated the total volume of postglacial extruded rocks in Iceland (lavas + tephra). His most recent estimate is 423 km^3.

Lava sizes are defined as follows:

> *Small lava*   *< 0.1 km^3*
> *Intermediate lava*   *≥ 0.1 – < 1 km^3*
> *Large lava*   *≥ 1 km^3*

Table 1 is the basis for the following discussion, and was prepared especially for this purpose. Information on lava ages and volumes is scarce and patchy in the geological literature, exceptions being the papers of Jónsson (1978) and Jakobsson et al. (1978) on the Reykjanes Peninsula. Information on the age and volume of a certain lava often comes from two different sources. The methods used for estimating lava sizes are as different and numerous as the authors. The same can be said about their age: the resolution is sometimes inconsistent, but Icelandic tephrachronology provides a valuable method of dating and makes possible reliable age estimates. Here, for the first time, a record of the volumes of the largest Icelandic lavas appears. Future observations will very likely revise the order and the ages slightly but the main trends are expected to hold.

Forty-five large postglacial lavas are listed in Table 1, totalling 247 km^3 of rocks, or 58% of the estimated total volume of eruptives. They are distributed fairly evenly over the volcanic belts of Iceland, with the exception of the Snæfellsnes off-rift zone where no large lavas are found (Fig. 1). Twenty-one of them are from fissure eruptions (108 km^3) and twenty-four from lava shields (139 km^3). Here the ratio of the shields makes up 139 km^3/247km^3 = 0.56 or 56% of the total volcanic production. The largest shields are similar in size (or a little smaller) to the largest fissure lavas (Fig. 2). This outcome is similar to that for the Dyngjufjöll region (Sigvaldason et al. 1992) but is considerably different from the pattern on the Reykjanes Peninsula (Jakobsson et al. 1978, Gudmundsson 1986).
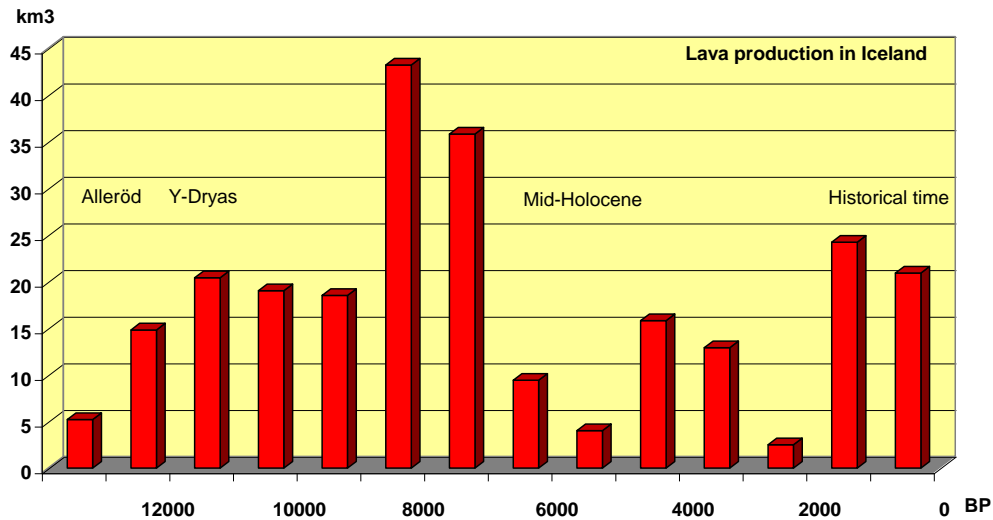
Fig. 3. Production of large lavas in Iceland per millennium. Two (or three) pulses appear, the first one in the early Holocene and the second in historical time with the great eruptions of Eldgjá and Laki.

Fig. 3 illustrates the production of large lavas per millennium. Some pulses appear, the largest one in the early Holocene and the second largest covering the last 2,000 years. However, the pulses constitute very few, though large, lavas and statistically they are not very convincing.

Although resolution of the age data in Table 1 is rather poor, the division between early and late Holocene is thought to be reasonably good. Early Holocene production ($\geq$ 6000 years BP) is 164 km^3 and late Holocene (and late glacial) production is 80 km^3, i.e. the ratio is roughly 2:1. This ratio does not agree with the immense eruptive pulse estimated by Jull and McKenzie (1996). They suggested a 30-fold increase in magma production in the first millennium after the deglaciation, as will be discussed later.

Production of the Shields

Fig. 4 shows individual large postglacial eruptions, with sizes plotted against age. Here the two main pulses are prominent and the high productivity in early Holocene and historical time is evident, as is a distinct Mid-Holocene minimum. The obvious difference between shields and fissure rows is seen more clearly in Figs. 5 and 6. The lava production of the lava shields started as soon as the glaciers retreated inland from the shore line and culminated already in late glacial times. In the Mid-Holocene the production of the lava shields decreased. During the last 4,000 years only three of them have erupted, which might be similar to the Neogene frequency of such volcanism, and no small shields have erupted. Therefore it is a fair statement that the period of the shields lasted from 13,000 to 4,000 BP, i.e. for 9,000 years.
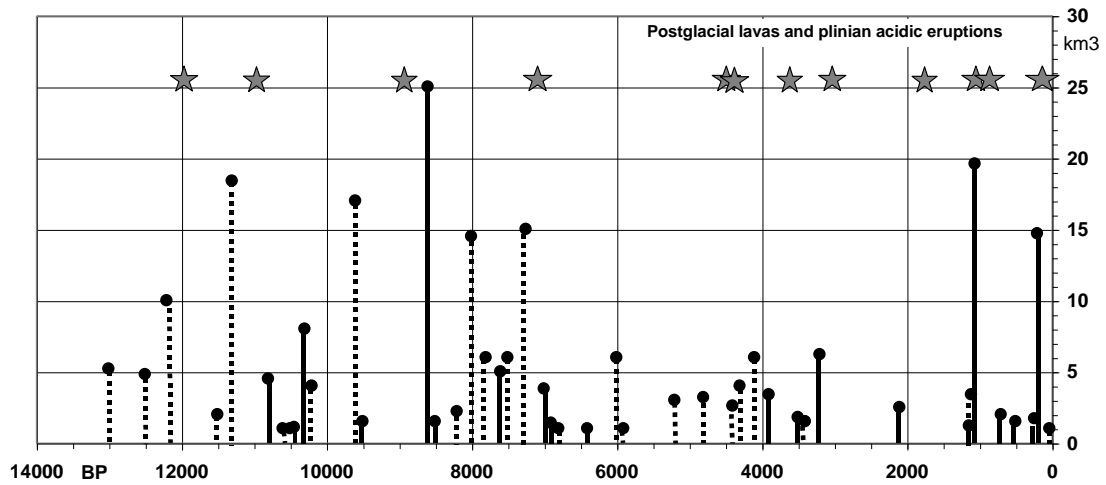
Fig. 4. Large lava eruptions in Iceland in postglacial time. Acidic plinian eruptions are also shown. Dashed lines = lava shields, solid lines = crater rows, stars = acidic plinian eruptions. Lava production is highest in the early Holocene but declines in the mid-Holocene, with a prominent peak in historical times. The frequency of acidic plinian eruptions seems to be increasing towards present times.
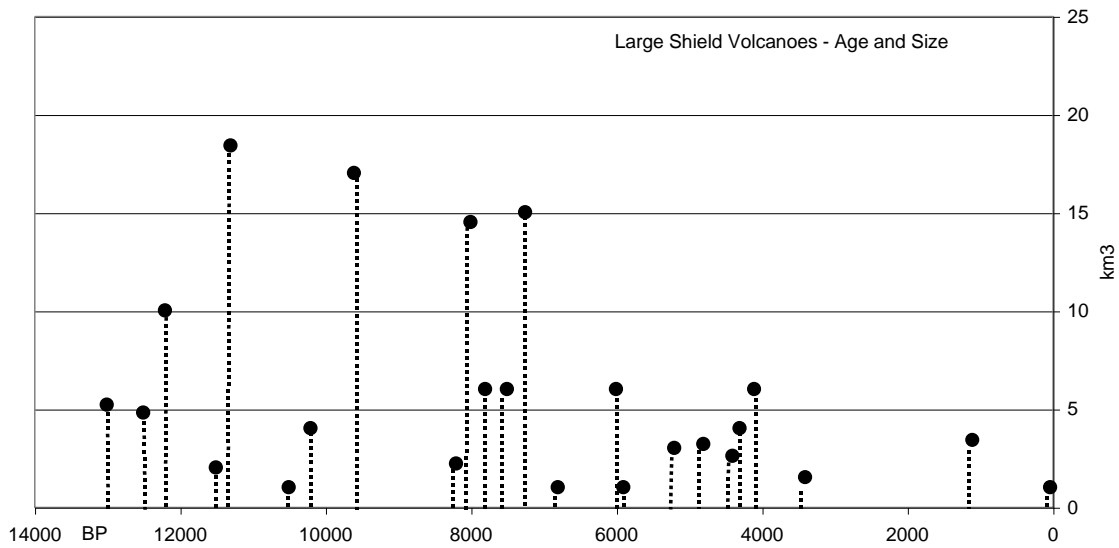
Fig. 5. Size and timing of large postglacial shields. Their productivity of compound lavas culminates in late glacial times with a constant decline during the Holocene.

A relationship between lava shields, glacial retreat and unloading might be proposed on the basis of this evidence. However, the occurrence of shield volcanism during and after deglaciation might also be fortuitous. Lava shields and compound lavas are rare in the Neogene lava pile of Skagafjörður but become abundant in the Pleistocene pile (Article 1

in this thesis; see also the Appendix 5.2.3.). This difference has also been recorded in the Plio-Pleistocene strata pile of Fnjóskadalur, North Iceland (Jancin et al. 1985).An increasing abundance of shields during late Pleistocene is also evident in the Reykjavík region (Torfason et al. 1997, 1999, 2000). An increasing number of lava shields in the volcanic pile of these regions, after the onset of glaciations, supports the idea about relationship between shields and deglaciations.
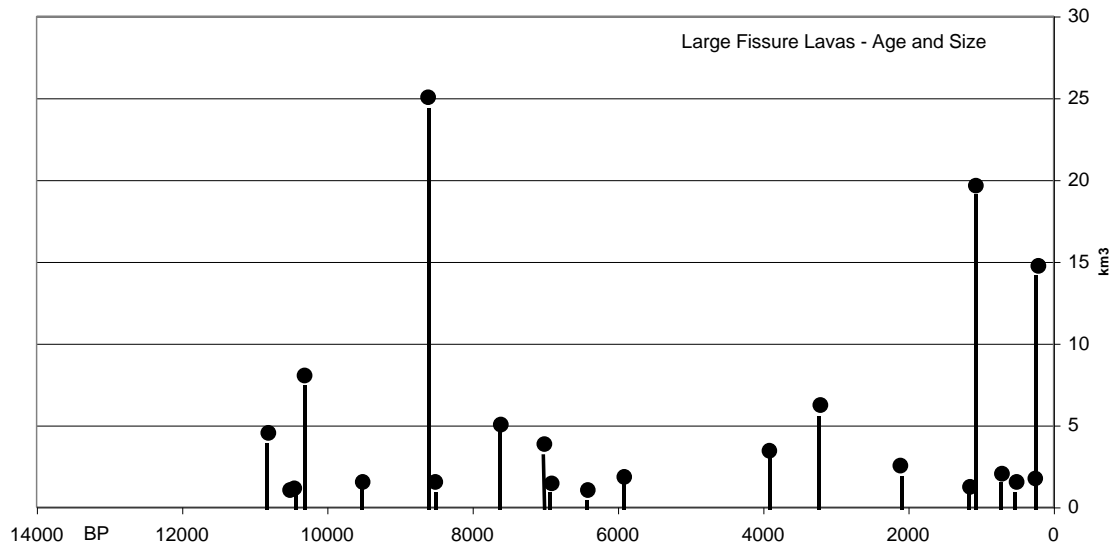
Fig. 6. Size and timing of large postglacial fissure lavas. Their productivity culminates in the early Holocene and in historical time with a distinct decline in the mid-Holocene.

Production of Crater Rows

The eruptions from crater rows show a different pattern from that of the lava shields (Fig. 6). No large lavas of late glacial age are known. Productivity culminates in the early Holocene, three thousand years later than the main pulse of the compound lavas, then a distinctive decline takes place between 6,000 and 4,000 years BP, after which a prominent peak appears during historical times (< 1,100 years BP). No effects of ice melt and glacial isostasy appear.

The best-known acidic plinian eruptions of the Holocene and late glacial times in Icelandic central volcanoes are shown in Table 2 and Fig. 3. These seem to exhibit a similar pattern to that of the fissure eruptions and do not seem to be influenced by the deglaciation. The longest interval between such eruptions coincides with the mid-Holocene decline. The last 2,000 years seem to have been exceptionally active as regards both large fissures and central volcanoes.
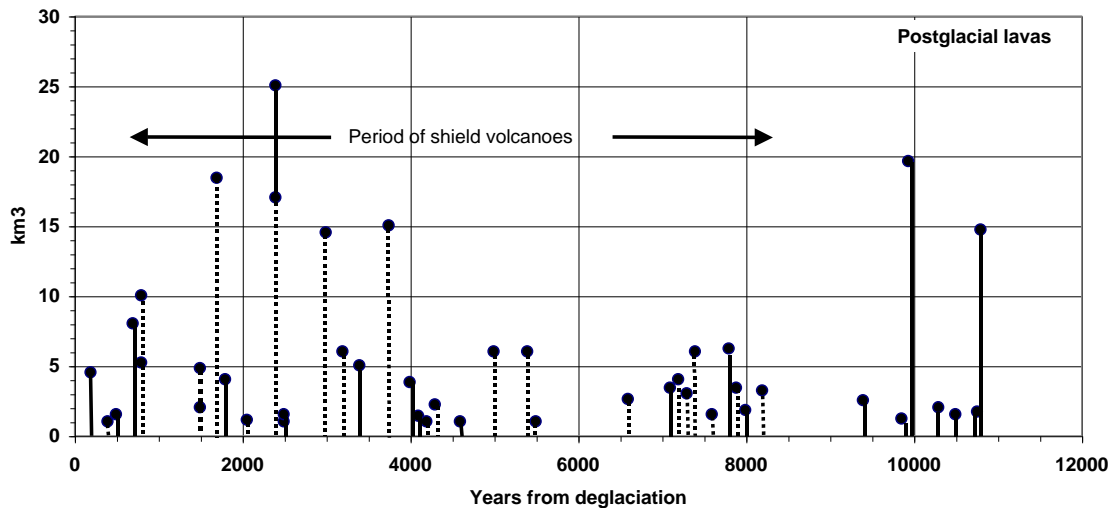
Fig. 7. Time of eruptions plotted against the time since the deglaciation of each area. The shield eruptions group within 8,000 years of the deglaciation. (The Surtsey eruption is omitted from the figure).

Discussion

In his study of the petrology of recent basalts in the Eastern Volcanic Zone, Jakobsson (1979a) plotted the frequency of basaltic eruptions during postglacial time. His diagram shows two periods of high frequency of eruptions, i.e. 0-2,000 BP and 5,000-8,000 BP. Jakobsson also noted that the eruption frequency of each volcanic system inside the Eastern Volcanic Zone conforms to this same pattern, and moreover, where accurate age determinations are available from volcanic systems outside the zone, they fall within either of the two periods. Jakobsson's observations indicate that the activity in the volcanic systems harmonizes with the fluctuations in the frequency and volumes of the large lavas discussed here in the present paper. These two independent observations support each other.

The conclusion of the present discussion is that the high lava production following the last deglaciation is essentially due to eruptions of lava shields. The interrelation between the shields and deglaciation becomes still clearer if allowances are made for the time difference between local deglaciations in different parts of the country. Four to five thousand years may have passed between the ice retreat from the outer Reykjanes Peninsula and the final retreat from the Central Highlands near Vatnajökull (Ingólfsson 1998). If a correction is made for this and the timing of an eruption of a lava shield is calculated as the time from the deglaciation of the area, the shield eruptions group within 8,000 years of the local deglaciation (Fig. 7). The overall lava production shows, in the same manner, a distinct pulse culminating c. 2,000 years after the ice retreat.

Glacier unloading and crustal rebound stimulate eruptions of lava shields but seem to have little effect on the crater rows and central volcanoes.

This outcome is supported by Gudmundsson's (1986) mechanical eruption model and investigation on the Reykjanes Peninsula. The late glacial-early Holocene eruption pulse of the lava shields is striking and if the production rate of the crater rows is examined it seems to be around 1 km^3/kyr on average during the postglacial period but over 2 km^3/kyr during the last millennium. No early Holocene pulse appears for the crater rows on the Reykjanes Peninsula.

On the other hand this contradicts the development in the Veiðivötn system (Vilmundardóttir and Larsen 1986), and the Dyngjufjöll system (Sigvaldason et al. 1992, Sigvaldason 2002) and partly contradicts the results of the calculations of Jull and McKenzie (1996) regarding the effect of deglaciation on mantle melting.

The Veiðivötn volcanic system is the most productive one in Iceland. Its production pattern is governed by the overwhelming size of the Great Þjórsá lava (25 km^3) and the timing of maximal production depends entirely on this single event. No lava shields have been recognized near Veiðivötn. Ignoring the Þjórsá lava, the production rate in the Veiðivötn system seems to be rather even during the Holocene.

The production pattern of the Dyngjufjöll system displays a clear maximum in the period 10,000-4,500 years BP, both with respect to shields and fissure lavas and also the production of acidic eruptives (Sigvaldason et al. 1992, Sigvaldason 2002). The age resolution of the lavas before 4,500 BP is very poor but even so it can be stated that the lava shields follow the general trend of maximum emission in early postglacial times. On the other hand, Sigvaldason's (2002) statements on the relationship between the great plinian eruption in Dyngjufjöll and glacial rebound in the area are not convincing. The distribution of plinian eruptions in time (Table 2, Fig. 4) does not support such a relationship in general. The timing of the great Askja eruption at the Pleistocene/Holocene boundary seems incidental. In the same manner it can be stated that the early Holocene production pulse of the fissure lavas in the Askja volcanic system was not caused by glacial rebound but rather by the activity of the central volcano combined with its colossal caldera collapse.

Jull and McKenzie (1996) calculated the effect of deglaciation on mantle melting beneath Iceland. They found that the average melting rate and magma production increased by about 30 times its steady-state value when a 2,000 m thick glacier disappeared in 1,000 years. They also state that this can only happen because melt generation is reduced for about 60,000 years after the ice load is applied. The observations and arguments presented above do not support the existence of such a high-intensity eruptive pulse. Part of the explanation is the short deglaciation period in the model of Jull and McKenzie. Instead of a 1,000-year period of glacier retreat, 5,000 years would have been more appropriate, which would then give a much smaller increase in the eruptive pulse.

Conclusions

Forty-five large postglacial lavas (≥ 1 km³) are listed in volumetric order in Table 1 together with the best available information on their age. These lavas make up 247 km^3 of rocks, or 58% of the estimated total postglacial volume of eruptives in Iceland. Twenty-one of them are from fissure eruptions (108 km^3) and twenty-four from lava shields (139 km^3). The largest shields are similar in size to the largest fissure lavas.

A diagram showing the lava production per millennium reveals two pulses. The first and larger one occurred between 9000 and 7000 BP (calendar years) and the second extends over the last 2,000 years. When eruptions of lava shields and crater rows are examined individually, it becomes evident that they have different periods of activity. Most of the largest lava shields in Iceland were erupted in the first millennium after deglaciation. During the last 4,000 years only three large lava shields have been erupted. It can be suggested that the period of the shields lasted from 13,000 to 4,000 BP, i.e. for 9,000 years. On the basis of this evidence, a relationship between lava shields, glacial retreat and unloading might be suggested; on the other hand, the occurrence of shield volcanism during and after deglaciation might also be fortuitous. Lava shields and compound lavas are rare in the Neogene lava pile of Iceland but seem more abundant in the Quaternary pile. An increasing number of lava shields in the volcanic pile after the onset of the Pleistocene glaciations supports statements about a relationship between shields and glacial unloading.

The crater rows, on the other hand, do not show any clear correlation to deglaciation. Their lava production is marked by two pulses, the first 9,000-8,000 BP and the other during the last 1,100 years. These pulses are primarily due to three exceptional eruptions, firstly the Great Þjórsá Lava (25 km³) and secondly the historical Eldgja and Laki eruptions (19.6 km³ and 15 km³ each). Between the production highs there is a prominent low, the mid-Holocene low, a 2,000-year period when no large fissure eruption took place. The explosive plinian eruptions in the central volcanoes of Iceland exhibit a similar pattern to that of the crater rows, with more activity in the early and late Holocene than in the mid Holocene, but in the long run their frequency has increased with time. Glacier unloading and crustal rebound seem to stimulate eruptions of lava shields but their influences on the crater rows and central volcanoes are small, if any.

References

Flores RM, Jónsson J and Jóhannesson H (1981) A short account of the Holocene tephrachronology of the Snæfellsjökull central volcano, western Iceland. (Ljósu öskulögin frá Snæfellsjökli), Jökull 31:23-30

Grönvold K, Oskarsson N, Johnsen SJ, Clausen HB, Hammer CU, Bond G and Bard E (1995) Ash layers from Iceland in the Greenland GRIP ice core correlated with oceanic and land sediments. Earth Planet Sci Lett 135:149-155

Guðmundsson AT (2001) Íslandseldar. Mál og menning, Reykjavík, pp 1-450

Gudmundsson A (1986) Mechanical Aspects of Postglacial Volcanism and Tectonics of the Reykjanes Peninsula, Southwest Iceland. J Geophys Res 91:12711-12721

Hjartarson Á (1994) Environmental changes in Iceland following the Great Þjórsá Lava Eruption 7800 14C years BP. In: Environmental Change in Iceland. Ed. J.Stötter & F.Wilhelm, Munchen, pp 147-155

Hjartarson Á (1995) Hekla and its surroundings (Á Hekluslóðum). Árbók Ferðafélags Íslands 1995, Reykjavík1-236

Hjartarson Á (1999) The geology of Ölfus, overview. (Yfirlit um jarðfræði í Ölfusi og grennd). ÁH 99/06, Orkustofnun, Reykjavík, pp 1-11 + map

Ingólfsson Ó (1998) Isavsmältning och israndlinjer på Island. In Andersen and Pedersen: Israndslinier i Norden. TemaNord 1998:584. Nordisk Ministeråd, København, pp 109-124

Jakobsson SP (1972) Chemistry and distribution pattern of recent basaltic rocks in Iceland. Lithos 5:365-386

Jakobsson SP (1979a) Petrology of recent basalts of the Eastern volcanic zone, Iceland. Acta Naturalia Islandica 26, pp 1-103

Jakobsson SP (1979b) Outline of the petrology of Iceland. Jökull 29:57-73

Jakobsson SP, Jónsson J and Shido F (1978) Petrology of the Western Reykjanes Peninsula, Iceland. J Petrol 19:669-705

Jancin M, Young KD, Voight B, Aronson JL and Sæmundsson K (1985) Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. J Geophys Res 90, 9961-9985

Jóhannesson H (1989) The age of Hallmundarhraun lava. (Aldur Hallmundarhrauns í Borgarfirði). Fjölrit Náttúrufræðistofnunar Íslands no 9. pp 1-12

Jónsson J (1978) Geological map of Reykjanesskagi. (Jarðfræðikort af Reykjanesskaga). OS-JHD 7831, Orkustofnun, pp 1-333 + maps

Jónsson J (1978) On the geology of Hveragerði and surroundings. (Hveragerði og nágrenni. Jarðfræðilegt yfirlit). Rannsóknarstofnunin Neðri Ás, skýrsla nr. 50, pp 1-55

Jull M and McKenzie D (1966) The effort of deglaciation on mantel melting beneath Iceland. J Geophys Res 101:21815-21828

Kjartansson G (1966) The volumes of lava shields. (Rúmmál hraundyngna). Náttúrufræðingurinn 36:125

Kristjánsson BR, Jónsson SS, Sæmundsson K and Hafstað ÞH (2003) Grunnvatnsborholur á Hellisheiði og nágrenni : greining jarðlaga í HK-holum 2001-2002, Orkustofnun, Rannsóknasvið, OS-2003/003 pp 1-60

Larsen G (2000) Holocene eruptions within the Katla volcanic system, south Iceland: Characteristics and environmental impact. Jökull 48:1- 28

Larsen G and Thorarinsson S (1977) H4 and other acid Hekla tephra layers. Jökull 27:28-45

Rossi MJ (1996) Morphology and mechanism of eruption of postglacial shield volcanoes in Iceland. Bull Volcanol 57:530-540

Sigurðsson O, Zóphóníasson S, Hannesdóttir L, Þorgrímsson S 1975: Dettifdossvirkjun, jarðfræðiskýrsla, OS-ROD-7526, Orkustofnun, Reykjavík, pp 1-62 + maps

Sigvaldason GE (1992) Recent hydrothermal explosion craters in an old hyaloclastite flow, central Iceland. J Volcanol Geotherm Res 54:53-63

Sigvaldason GE (2002) Volcanic and tectonic processes coinciding with glaciation and crustal rebound: an early Holocene rhyolitic eruption in the Dyngjufjöll volcanic center and the formation of the Askja caldera, north Iceland. Bull Volcanol 64:192-205

Sigvaldason GE and Steinþórsson S (1974) Chemistry of tholeiitic basalts from Iceland and their relation to the Kverkfjöll hot spot. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, pp 155-164

Sigvaldason GE, Annertz K and Nilsson M (1992) Effect of gracier loding/deloding on volcanism: postglacial volcanic production rate of the Dyngjufjöll area, central Iceland. Bull Volcanol 54:385-392

Sinton J and Grönvold K (2001) Post-Glacial Eruptive History of the Western Volcanic Zone, Iceland. (Poster)

Steinþórsson S (1967) Two new C14 dates on Snæfellsjökull tephra layers.(Tvær nýjar Cl4-aldursákvarðanir á öskulögum úr Snæfellsjökli). Náttúrufræðingurinn 37:236-238

Sæmundsson K (1973) Straumrákaðar klappir í kringum Ásbyrgi. Náttúrufræðingurinn 43:52-60

Sæmundsson K (1992) Geology of the Thingvallavatn area. In: ed. Jónasson 1992, Ecology of oligotrophic, subarctic Thingvallavatn, pp 1-434

Sæmundsson K (1995) Um aldur stóru dyngnanna á utanverðum Reykjanesskaga. In: Björn Hróarsson, Dagur Jónsson og Sigurður Sveinn Jónsson (eds) 1995: Eyjar í eldhafi. Gott mál, Reykjavík, pp.165-172

Thorarinsson S (1951) Laxárgljúfur and Laxárhraun a tephrochronological study. Geografiska annaler 1951:1-89

Thorarinsson S (1958) The Öræfajökull eruption of 1362. Acta Naturalia Islandica II, pp 1-100

Thorarinsson S (1968) The last stages of the Surtsey eruptiuon. (Síðustu þættir Eyjaelda). Náttúrufræðingurinn 57:113-135

Thorarinsson S (1979) The postglacial history of the Mývatn area. Oikos 32:17-28

Thorarinsson S, Einarsson T. and Kjartansson G (1959) On the geology and geomorphology of Iceland. Geografiska Annaler 41:143-153

Thordarson Th and Self, S (1993) The Laki (Skaftar fires) and Grimsvotn eruptions in 1783-1785. Bull. Volcanol. 55:33-63

Thordarson Th, Miller DJ, Larsen G, Self S. and Sigurdsson H (2001) New estimates of sulfur degassing and atmospheric mass-loading by the 934 AD Eldgja eruption, Iceland. J Volcanol Geotherm Res 108:33-54

Torfason H, Hjartarson Á, Jóhannesson H, Jónsson J and Sæmundsson K (1997) Bedrock map, Viðey 1613 III NV-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason H, Hjartarson Á, Jóhannesson H, Jónsson J and Sæmundsson K (1999) Bedrock map, Vífilsfell 1613 III SA-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason H, Hjartarson Á., Jóhannesson H, and Friðleifsson IB (2000) Bedrock map, Mosfellssveit 1613 III NA, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Vilmundardóttir EG (1977) Tungnárhraun. OS ROD 7702, Orkustofnun, pp 1-156 + map

Vilmundardóttir EG and Larsen, G (1986) Productivity pattern of the Veiðivötn fissure swarm, southern Iceland,in postglacial time. Preliminary results. 17. Nordiska Geologmötet , Helsingfors

Vilmundardóttir EG, Snorrason SP, Larsen G and Guðmundsson Á (1988) Bedrock map. Sigalda – Veiðivötn 1:50.000

Walker GPL (1971) Compound and Simple Lava Flows and Flood Basalts. Bull Volcanol 35:579-590

Crustal spreading in Iceland
The dilemma between spreading rate and rock age

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Abstract

Iceland is spreading towards the east and the west, away from rifting axes that transect the country. Consequently the oldest rocks, which date from roughly 14 million years BP, are found on the east and west coasts, 480 km apart. This indicates a spreading rate of 3.4 cm/yr, which differs widely from the generally accepted North Atlantic spreading rate of 1.8 cm/yr. A spreading model for North Iceland is presented, and several example calculations of the spreading rate are shown. They all give similar results. Trajectory calculations of the oldest rocks in East and West Iceland, from their origin near the rift axis to their present locations, yield spreading rates of 3.5 cm/yr and 3.4 cm/yr, respectively, the drift of Anomaly 5 towards the east gives 3.5 cm/yr, and the spreading of the Plio-Pleistocene belt in Northern Iceland indicates 3.5 cm/yr. Other examples give similar values, even for historical time. Although the spreading is strikingly uniform along a reference line from east to west, it varies considerably from north to south, reaching a maximum at the centre of the Iceland hot spot, where the spreading rate may exceed 6 cm/yr. The excess spreading decreases gradually along the rift axes, and reaches the normal value of 1.8 cm/yr some 300-400 km from the centre of the hot spot. The given examples of excess spreading rate are strikingly similar in spite of the different time intervals and types of formations. They seem to evince a steady and long-lasting geological process that has been operating during the last 15 Ma of the geological history of Iceland.

Introduction

During an expeditions to the central highlands west of Vatnajökull glacier in 1924 and 1927, Danish geologist Niels Nielsen observed firsthand the extended volcanic structures and crater rows that dominate the landscape there. Inspired by Wegener's ideas about continental drift, Nielsen proposed that Iceland was being split apart by tensional forces on a regional scale, giving rise to the fissural volcanism and associated linear tectonics that characterize the country (Nielsen 1933). His hypothesis was rejected at the time and remained out of favour for decades, only to reappear in the 1960's within the context of plate tectonics. Since then, it has been widely accepted that Iceland is spreading apart.

The first estimates of the rate of crustal spreading in the Iceland region appeared in a landmark paper by Tuzo Wilson (1963). Since 1966 the spreading rate has been considered to be in the range of 1.6-2.0 cm/yr (Vine 1966). This value was originally determined from the spacing of magnetic anomalies adjacent to the ocean ridges to the north and south of the island, and from radiometric dating. More recently, satellite methods based on the Global Positioning System (GPS) have been used to measure directly the spreading rate and its variations (Sigmundsson et al. 1995, Sella et al. 2002). The magnitudes and directions of the spreading vectors vary slightly from northern Iceland to the southern part of the country, averaging 1.8 cm/yr in the direction N105°E (DeMets et al. 1990, Guðmundsson 2000).

Iceland is a volcanic island positioned astride the North Atlantic Mid-Ocean Ridge. It is almost entirely built up of eruptive materials, which originate mainly in the volcanic belts that extend across the country from the southwest to the north. These materials drift in opposite directions, away from the volcanic zone. The oldest rocks are accordingly found at the eastern and northwestern extremities of the 480 km wide island. Radiometric determinations, paleomagnetic investigations, and indirect geological estimates have revealed that the oldest surface rocks date from the mid-Miocene. They are around 15 million years old in Northwest Iceland (Hardarson et al. 1997) and 13 million years old in the Gerpir promontory on the eastern coast (Moorbath et al. 1968, Ross and Musset 1976, Watkins and Walker 1977, Hardarson et al. 1997).
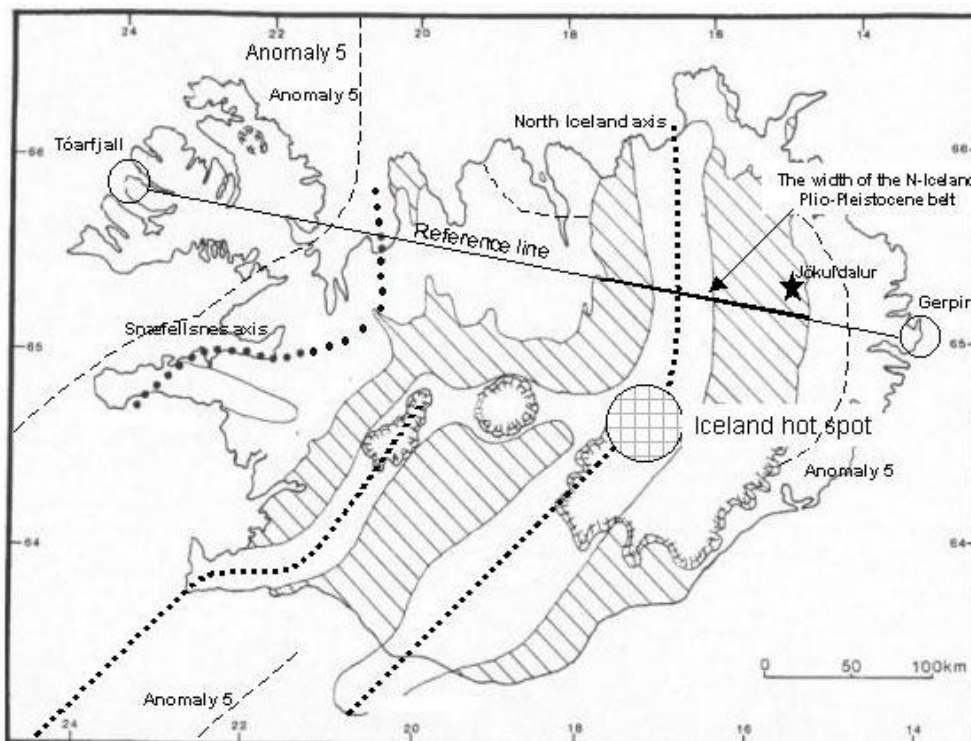
Fig. 1. Map of Iceland displaying the current rift axis, the extinct Snæfellsnes - Húnaflói rift axis, the neovolcanic zones, the Iceland hot spot, Tóarfjall and Gerpir and the reference line between them, the Jökuldalur sequence, and finally the alignment of Anomaly 5.

These figures immediately present a problem. If we divide the width of the island (480 km) by the average of the rock ages in the northwest and the east (14 Ma), we obtain the spreading rate 34.3 km/Ma or 3.4 cm/yr. This does not agree with the generally accepted spreading rate of 1.8 cm/yr. If the age is accurate, Iceland would seem too large. If the spreading rate is correct, Iceland would appear too young.

That the crustal spreading rate appears to be higher on land in Iceland than offshore has been pointed out and discussed previously. Walker (1975, 1976) noted that the aggregate width of the active volcanic zones is greater in South Iceland than in the northern part of the country. Assuming this width to represent the total extent of spreading, he inferred that the spreading rate in South Iceland might be several times higher than in North Iceland and on the adjacent mid-ocean ridges. Searle (1976) voiced what he called "a serious objection to this idea." Such excess spreading would result in either a zone of plate consumption or in major compressive deformation of at least one of the lithospheric plates. Neither has been observed. Pálmason (1981) regarded the inconsistency between spreading and rock ages as an unsolved problem in Icelandic geology. Helgason (1985) discussed the "excess width" of the country and tried to solve the problem by suggesting complex interactive ridge jumps. Bott (1985) on the other hand tried to resolve the matter in his reconstruction of the tectonic evolution of the Icelandic transverse ridge by assuming that ancient crust were hidden somewhere below the surface.

Rift jumps

The geology and tectonics of Iceland may be explained as a complex interaction between the North Atlantic Mid-Ocean Ridge and the Iceland Plume. While the North Atlantic is spreading symmetrically away from the mid-ocean ridge, the ridge itself is drifting slowly to the northwest with respect to the plume.

Table 1. Rift zones and rift jumps

Rift zone	Time of formation Ma	Reference
Northwest Iceland rift zone	24	Hardarson et al. 1997
Snæfellsnes – Húnaflói rift zone	15	Hardarson et al. 1997
North Iceland rift zone	6 – 7	Sæmundsson 1979
Reykjanes rift zone	6 – 7	Sæmundsson 1979
Eastern rift zone	2 – 3	Sæmundsson 1979
Skagafjörður rift zone	1.5	Article 4 in this thesis

As the ridge migrates over the plume, it repeatedly shifts its spreading axis in a process referred to as rift jumping. Several jumps have been identified in the geological history of Iceland (Table 1), but for the purposes of the following discussion only two jumps and two rift zones need to be considered, the Snæfellsnes - Húnaflói rift zone and the North Iceland rift zone. The former zone was established in the Snæfellsnes - Húnaflói region after a rift jump 15 million years ago. It remained active for 8-10 million years, but 6-7 millions years ago another jump gave rise to the North Iceland zone (Sæmundsson 1974, 1979, Johannesson 1980, Hardarson et al. 1997, Kristjánsson and Jónsson 1998).

A rift jump is not a sudden event. The transfer of the rifting activity from the Snæfellsnes - Húnaflói zone to the North Iceland zone here is suggested to have taken ca. 2 million years. Two volcanic zones were simultaneously active in North Iceland during this transition period, mirroring the present-day situation of South Iceland. The old Miocene eruptive rocks in West and East Iceland mostly originate in the extinct Snæfellsnes-Húnaflói zone, but the Plio-Pleistocene and Holocene lavas were erupted in the still active volcanic zone (Sæmundsson 1979).

Garcia et al. (2003) state that a paleo-rift, located some 60 km east of the Snæfellsnes-Húnaflói rift zone, was active in the Skagafjörður region between 8.1 – 2.9 Ma. They consider that the Snæfellsnes-Húnaflói zone was only a short-lived spreading centre, if it existed at all. Their suggestion is based on new Ar/Ar dates of dykes, but most age determinations in the Neogene and early Pleistocene regions of Iceland have been made on lavas or hyaloclastite formations. Garcia et al. suggest that when a study focuses on spreading rates and rift jumps, dates from dykes are more reliable than dates from lavas because the lavas might have flowed tens of km from their magmatic sources. By using dykes however, they lose all stratigraphical control, which is a Shortcoming of their method. Another weakness of their work is that their alternative spreading model contradicts the main patterns in the general stratigraphy of the country (Jóhannesson and Sæmundsson 1998a) and the paleomagnetic time scale for North Iceland (Sæmundsson et al. 1980, Kristjánsson et al. 1992). Moreover they do not account sufficiently for the possibility that dykes can be injected into the crust outside the volcanic zones (Walker 1974, Jancin et al 1985). This possibility reduces the reliability of the dataset they use in the spreading discussion. However, it is interesting to see very young dates (< 2 Ma) on dykes in the Skagafjörður region. They are problematic in the argument of Garcia et al. but fit well into the spreading history as it is introduced here because they seem to belong to the young Skagafjörður rifting (Table 1). The overall spreading rate in their rift model also favours of the main conclusions in this paper, as will be discussed later.

Spreading model

At this point, a model of the spreading in North Iceland must be introduced. This model is represented as a simple time - distance diagram, with time on the y-axis and distance on the x-axis (Fig. 2). The distance is measured along the so-called reference line. Its alignment is roughly parallel to the spreading vectors, and it connects the oldest dated rocks in the east and the west (Fig. 1). The origin of the x-axis is at the intersection of the line with the extinct Snæfellsnes - Húnaflói rift axis.

This model will be used to trace the trajectories of some well-dated rock formations in Iceland, from their origin in the volcanic belt to their current locations. The examples given all lie close to the reference line. The model is based on the following assumptions:

1. The spreading rate in Iceland is 1.8 cm/yr.
2. The reference sites in East and West Iceland, Tóarfjall and Gerpir, are 15 and 13 million years old, respectively.
3. Tóarfjall and Gerpir originate in the same volcanic zone on opposite sides of the Snæfellsnes - Húnaflói axis (Sæmundsson 1979). This means that no

ancient crust (>15 Ma) exists between the Snæfellsnes - Húnaflói and the North Iceland rift axes.

4. The spreading activity in the Snæfellsnes - Húnaflói zone ceased during a transition period 7-5 million years ago, and activity was transferred to the North Iceland zone.

5. The spreading is symmetric about the rift axis.

The spreading axis is a theoretical concept that represents the centre line of the volcanic zone. The rate of spreading is zero at the axis, but increases towards the margins of the volcanic zone, where it reaches a maximum value and remains constant thereafter. Volcanic materials erupted at the axis are buried under younger formations and disappear deep down into the crust. Materials formed at the margin of the volcanic zone will, on the other hand, drift outwards on the surface and soon erode away (Pálmason 1980). Eruptive formations found far away from the volcanic zone originate somewhere midway between the axis and the margin of the zone. For simplicity, neither the slower rate of spreading inside the volcanic zone nor the origin of volcanic formations at some distance from the axis will be taken into account in this discussion. These two effects partly cancel each other.

Examples on the spreading rates

Example 1 has already been mentioned, the width of Iceland versus the average rock ages of the oldest formations in the northwest and the east:
480 km/14 Ma = 34.3 km/Ma = 3.4 cm/a.

Example 2: The Tóarfjall series in Northwest Iceland is one of the oldest dated rock formations in the country (Fig. 2). This series originated near the Snæfellsnes - Húnaflói rift axis roughly 15 million years ago. It drifted westwards at a rate of 0.9 cm/yr, the standard half-spreading rate. About 7 million years ago it began to slow down, and 5 million years ago it stopped drifting, when the Snæfellsnes - Húnaflói rift axis became extinct. At this point it had drifted 81 km, as Fig. 2 shows, still about 70 km short of the 150 km needed for it to reach its current location. For the formation to cover the entire distance in the time available, a half-spreading rate of 1.7 cm/yr would have been required.

Example 3: This is the trajectory of Gerpir promontory, the oldest dated formation in East Iceland. It erupted close to the Snæfellsnes - Húnaflói rift axis 13 million years ago, but at present it lies 330 km east of this axis. One may ask how it is possible that Gerpir has covered a distance from the spreading axis that is more than twice that of Tóarfjall, in spite of its younger age. The example provides the explanation. After its formation, Gerpir drifted eastwards at the half-spreading rate until 7 million years ago (Fig. 2). At this time the North Iceland zone became active between Gerpir and the Snæfellsnes - Húnaflói rift axis. This increased the drift rate, which gradually reached its full value over the next 2 million years as the activity was being transferred to the North Iceland zone. This rate has remained constant to the present day. As in the previous example, the standard value of 1.8 cm/yr is far too low. For Gerpir to have reached its present location, the half-spreading rate must have been 1.75 cm/yr and the full spreading rate consequently 3.5 cm/yr.
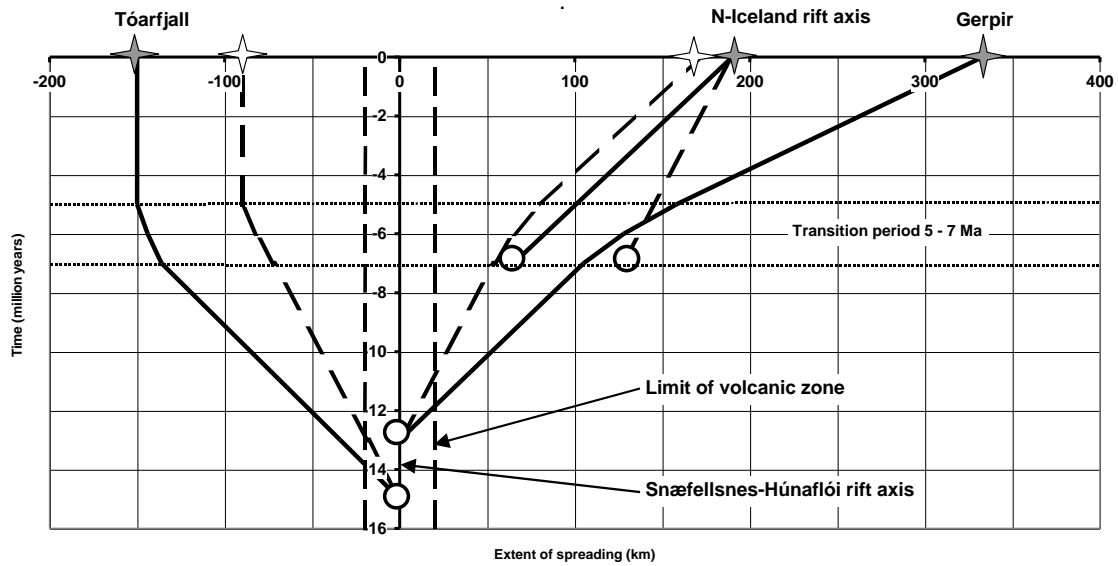
Fig. 2. Spreading model for North Iceland (time-distance diagram). Solid lines indicate the trajectories of Tóarfjall series, Gerpir rocks, and the North Iceland Volcanic Zone from their origins to recent locations, with respect to the Snæfellsnes – Húnaflói rift axis. Dashed lines indicate the trajectories for the 1.8 cm/yr spreading rate. The transition period indicates the rift jump from the Snæfellsnes – Húnaflói rift zone to North Iceland rift zone. See text for further details.

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Example 4: Figure 2 shows the postulated trajectory of the North Iceland rift axis, specifically its intersection with the reference line. Here the calculation must be carried out backwards in time, because the axis was formed by a rift jump 7 million years ago somewhere outside the active volcanic zone, and far away from the old axis. At present, the rift axes are 190 km apart. If the standard rate of spreading (1.8 cm/yr) is assumed, the new axis must have originated 130-140 km east of the Snæfellsnes - Húnaflói axis. But then Gerpir would have ended up to the west of the new axis and would thus be somewhere in Central Iceland and not on the east coast. If, on the other hand, a spreading rate of 3.5 cm/yr is assumed, this discrepancy does not arise. In this case the new axis initially appears 70 km east of the Snæfellsnes - Húnaflói rift axis, and Gerpir is on the right side.

Three more examples will be given, although these are not illustrated in Fig. 2.

Example 5: A magnetic anomaly in East Iceland, Anomaly 5 (C5n), has been very well established (Moorbath et al. 1968, McDougall et al. 1976b, Watkins and Walker 1977, Duncan and Helgason 1998, Jóhannesson and Sæmundsson 1998b). According to Cande and Kent (1995) it was formed during the period 9.9-10.9 million years BP. Application of the model presented above shows that a spreading rate of 1.8 cm/yr is far too low. A spreading rate of 3.5 cm/yr would put the centre of Anomaly 5 about 40 km west of Gerpir - where it is.

Example 6: The width of the Plio-Pleistocene belt constitutes a different observation. Where the reference line crosses it north of Mývatn, the belt is found to be 115 km wide if 10-20 km of far-reaching lavas on each side are disregarded (Jóhannesson and

Sæmundsson 1998b). The belt was formed inside the North Iceland Volcanic Zone and represents the width of the crust that has formed during the last 3.3 million years. This implies a spreading rate of 3.5 cm/yr.

Example 7: The Jökuldalur succession represents the youngest rocks discussed here, 1-2 Ma. This example differs from the others in that an entire sequence of layers is observed. The stratigraphy of Jökuldalur is well established because of a long-running debate on the existence of the Gilsá polarity subchron (Wensink 1964a, 1964b, McDougall and Wensink 1966, Watkins et al. 1975, Udagawa et al. 1999). More than 50 radiometric dates of carefully mapped and sampled sections in the valley have been published. Data from Udagawa et al. (1999) are used here to determine the spreading rate for the Jökuldalur rocks. The sections are 50-60 km east of the axis of the North Iceland Volcanic Zone. The numerical data and results are presented in Table 2.

Because of its high stratigraphic relief, the Jökuldalur succession is considered to have been erupted near the eastern margin of the North Iceland Volcanic Zone, maybe 30 km east of the hypothetical axis, which at present lies just west of Kollóttadyngja. The directions from the eruption vents to the points of sampling are not known, so all are presumed equally probable. Hence, no correction is made for the distance to the eruptive vents, and the succession as a whole is supposed to be of local origin. The half-spreading rates in Table 2 range from 1.6 to 2.4 cm/yr with an average of 2.0 cm/yr. Consequently, the full spreading rate is 4.0 cm/yr, which is higher than for the other cases examined here. This is not unexpected. The stratigraphic relief of the Jökuldalur succession is higher than that of the other formations considered, and it should therefore have the highest spreading velocity by the model of Pálmason (1980).

Table 2. Jökuldalur succession. Ages from Udagawa et al. (1999)

Strata	Age (Ma)	Displace-ment (km)	Half spreading rate (cm/yr)
TH14	0.91	22	2.4
TH13	0,95	-	2.3
TH11	1.10	-	2.0
TH10	0.91	-	2.4
TH8	0.91	-	2.4
TH7	0.98	-	2.2
TH6	1.25	-	1.8
TH5	1.26	-	1.7
TH 4	1.22	-	1.8
TH2	1.26	-	1.7
BD5	1.33	28	2.1
BD4	1.36	-	2.0
BD3	1.46	-	1.9
KG2	1.40	30	2.1
KG1	1.62	-	1.9
KG00	1.61	-	1.9
KG0A	1.85	-	1.6

All these seven examples indicate much higher spreading rates than expected. These rates are summarized in Table 3. They are strikingly similar in spite of the different time intervals and formation types. They seem to evince a steady and long-lasting geological process that has been operating during the last 15 Ma of the geological history of Iceland.

If the model is shunted north or south, different spreading rates are obtained, lower towards the north, higher towards the south. The maximum is somewhere near the centre of the Iceland hot spot, where the width of the Plio-Pleistocene belt in the direction of spreading exceeds 200 km. If this width represents the extent of spreading, the corresponding rate would appear to be >6 cm/yr. Farther south, the spreading rate decreases again. But here the data are less reliable and the spreading history is more confused, so the calculations are not as trustworthy. Nevertheless, the excess spreading rate clearly dwindles, and the rate reaches the normal value of 1.8 cm/yr some 300-400 km from the centre of the hot spot along the rift axis, as evidenced by the regular pattern of the marine magnetic anomalies at the Reykjanes and Kolbeinsey ridges. The average value for the Iceland spreading anomaly might be around 3.4 cm/yr.

Table 3: Spreading rates along the Tóarfjall – Gerpir reference line.

	Type of formation	*Time span* 10^6 *years*	*Spreading rate cm/yr*
Northwest Iceland (Tóarfjall)	Basalt lava	9	3.4
East Iceland (Gerpir)	Central volcano	13	3.5
Anomaly 5 in E-Iceland	Magnetic anomaly	10.5	3.5
North Iceland Plio-Pleistocene zone	Volcanic zone	3.3	3.5
East Iceland (Jökuldalur)	Basaltic succession	0.94	4.0
All of Iceland (Tóarfjall–Gerpir axis)		14	3.4

From their dates and alternative rift model, Garcia et al. (2003) obtain the total opening rate for North Iceland as 27.7 +/- 4.1 km/Ma during the time period 8 – 3 Ma. They have doubts about this result and argue that a propagation of various uncertainties might explain why this value differs from the present-day velocity of the plate divergence in the North Atlantic region, 18 km/Ma. However, they also point out that an increase in the plate divergence during this period cannot be excluded. After all, and despite my criticism on the method of Garcia et al. (2003), their spreading rate is in line with the main conclusions of this paper.

Dates, ages and ancient crust

The basic assumptions of the model may now be revisited. It is generally agreed that the spreading during the last 15 million years was exclusively between Tóarfjall and Gerpir. Assumptions no. 4 and 5 will therefore be dropped from the model. The timing of the rift jump and the question of symmetric spreading do not affect the outcome of the spreading rate calculations. Assumptions no. 1, 2, and 3 remain: the dates, the spreading rates, and the common origin. One of them, at least, appears to be wrong.

Table 4. Rock dates in Iceland (K/Ar and Ar/Ar dates)*

Reference	Number of dates	Method	Remarks
McDougall and Wensink 1966	5	K/Ar	Jökuldalur
Gale et al. 1966	21	K/Ar	Intrusions in E and W Iceland
Dagley et al. 1967	9	K/Ar	East Iceland
Moorbath et al. 1968	18	K/Ar	Diverse localities
Everts et al. 1972	13	K/Ar	Skagafjörður
Watkins et al. 1975	8	K/Ar	Jökuldalur
Aronson and Sæmundsson 1975	18	K/Ar	Diverse localities
McDougall et al. 1976a	12	K/Ar	Fljótsdalur, East Iceland
McDougall et al. 1976b	17	K/Ar	Anomaly 5, East Iceland
Ross and Mussett 1976	18	Ar/Ar	East Iceland
Bagdasaryan et al. 1976	31	K/Ar	Diverse localities
Watkins and Walker 1977	5	K/Ar	Gerpir
McDougall et al. 1977	24	K/Ar	Borgarfjörður W-Iceland
Albertsson 1978	7	K/Ar	Tjörnes
Sæmundsson et al. 1980	34	K/Ar	Tröllaskagi
McDougall et al. 1984	71	K/Ar	Northwest Iceland
Jancin et al. 1985	35	K/Ar	Flateyjarskagi – Fnjóskadalur
Albertsson and Eiríksson 1989	4	K/Ar	Flatey
Kristjánsson et al. 1992	6	K/Ar	Langidalur
Hardarson et al 1997	5	Ar/Ar	Northwest Iceland
Kristjánsson et al. 1998	5	K/Ar	South Iceland
Duncan and Helgason 1998	10	Ar/Ar	East Iceland
Udagawa et al. 1999	38	K/Ar	Jökuldalur
Garcia et al. 2003	37	Ar/Ar	North Iceland

*The table is in chronologic order. It does not include all K/Ar or Ar/Ar dates from Iceland. There are more than 450 dates in the table.

The diagram in Fig. 2 seems logical in itself. The statement that the spreading rate is several times higher in Iceland than on the Reykjanes and the Kolbeinsey ridges might, however, seem unacceptable on its face. But if the standard spreading rate for the country, 1.8 cm/yr, is correct, something must be wrong with the geological time scale of Iceland. The rocks should be older.

One possible explanation is that the dated rocks in Tóarfjall, Gerpir, Anomaly 5, and Jökuldalur are not of local origin, but from lavas that have flowed tens of kilometers along the spreading direction, away from their sites of eruption. If such is the case, 20-30 million year old local rocks should be found in Northwest and East Iceland, for example in dykes or inside central volcanoes. This would contradict paleomagnetic data, polarity time scales, and all the geochronology of Iceland that is based on extensive radiometric data (Table 4). In any case, this argument does not apply to the dates of the Gerpir rocks. They have been carefully determined and rechecked many times because of the key stratigraphic position of these rocks (Dagley et al. 1967, Moorbath et al. 1968, Ross and Mussett 1976, Watkins and Walker 1977, Mussett et al. 1980). The samples were taken from lavas inside the Gerpir-Barðsnes central volcano and are therefore of local origin.

Another objection is that Gerpir could be a part of an intraplate or off-rift volcano such as Snæfellsjökull or Snæfell. The stratigraphy and petrology do not support this.

The calculations for Anomaly 5, Jökuldalur, and the Plio-Pleistocene belt also show that errors in the dates, or in the interpretations of dates, are unlikely reasons for the high apparent spreading rate.

A way to resolve the discrepancy between the dates and rates in examples 1-5 is to include some ancient oceanic crust (>15 Ma), or a hidden piece of continental crust, somewhere below the surface, between the spreading axes of the Snæfellsnes - Húnaflói zone and the North Iceland zone. This would explain the high apparent, but false, spreading rates. This "piece" would have to be more than 200 km wide, though. The spacing of the axes today is 190 km, and it was only around 120 km at the time of the rift jump 6 million years ago (using 1.8 cm/yr). Hence, this suggestion does not seem tenable. It is also very difficult to see how substantial regions of older crust could be completely submerged by younger lavas. Furthermore, the spreading rates in examples 6 and 7 cannot be explained by ancient crust.

It may, of course, be argued that the spreading model is in some way fundamentally wrong. The model presented here, however, is rather simple and based on very few assumptions. It provides a satisfactory agreement with available geological evidence, and it fits the geological history of the country, as it is now understood after decades of research, including hundreds of rock age determinations (Table 4). The model is also rather insensitive to details such as the total number of rift jumps or the exact timing and locations of such jumps as long as they are inside the country; and the question of asymmetric vs. symmetric spreading is not of the essence. Thus, the model and its assumptions seem rather trustworthy. The spreading rate is the only part that does not fit. Even the alternative spreading model of Garcia et al. (2003) does not change this.

Various workers have assumed the North Iceland Volcanic Zone to have been active during the whole of the geological history of Iceland (Ross and Mussett 1976, Mussett et al. 1980, Bott 1985). In their discussion, Jancin et al. (1985) called this early (15-7 Ma) and enigmatic zone the Proto North Iceland Spreading Zone. If it ever existed, Gerpir might have originated there. But here a familiar problem arises. Gerpir has moved too far away from this hypothetical protozone. Covering this distance would have required the entire half-spreading rate of the last 13 million years. Thus nothing would have been left over for any spreading at the Snæfellsnes - Húnaflói axis, and the geological history of West Iceland would be thrown into confusion.

Continuous GPS Measurements

GPS measurements can provide accurate and detailed information on the spreading and movement of crustal plates. Obtaining unequivocal results for a geological process, such as the one discussed here, will probably require decades of continuous measurement, however.

Since 1999, the Icelandic Meteorological Office has operated a network of continuous Global Positioning System stations to monitor crustal deformation and spreading (Geirsson 2003). The results are displayed on the Office's Web page (www.vedur.is). EUREF (Europian Reference Frame) is running an international network of GPS station displaying its results on the website http://epncb.oma.be. The time series in question (Fig. 3 and 4) are still too short for any long-term trends to be evaluated.

Between March 1999 and August 2003, however, the distance between the stations in Reykjavík (North American plate) and Höfn (Eurasian plate) increased corresponding to a spreading rate of 20.3 mm/yr (after removing the coseismic movements of the earthquakes of June 2000, Halldór Geirsson, private communication).
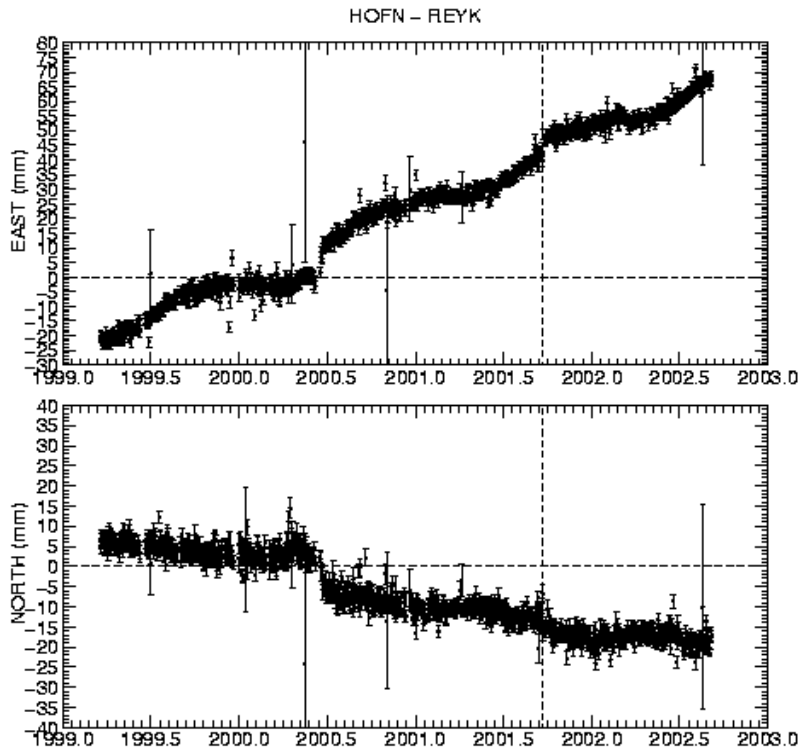

Fig 3. The crustal spreading across Iceland along the GPS-baseline Reykjavík – Höfn. Since March 1999 the distance between the stations in Reykjavík has increased corresponding to a spreading rate of 20.4 mm/yr. The break in the general trend in June 2000 is due to large earthquakes. (The picture comes from the website of the Icelandic Meteorological Office).

Sella et al. (2002) have presented a global model for recent plate velocities (REVEL) based on publicly available space geodetic (primarily GPS) data for the period 1993 - 2000. Their spreading rate for Iceland is 20.3 +/- 0.2 mm/yr, calculated at the point where the baseline between the GPS stations in Reykjavík and Höfn crosses the plate boundary (64.2°N, 18.8°W). This is slightly higher than the NUVEL velocity (1.85 mm/yr, DeMets et al. 1990), but lower than the high rates discussed above.

If Iceland is spreading as suggested above, its eastern part should be moving towards Europe. In the EUREF Permanent Network data this should appear at the GPS-station in Höfn. According to their improved data this is actually the fact (Fig. 4), since 1998 Höfn has been approaching the continent by 4 mm/yr in the average (http://epncb.oma.be/ organisation.html).

The steadiness of the rifting is noteworthy, both for large distances such as between Reykjavík and the European mainland, and for short stretches such as between Reykjavík and the nearby GPS-station in Hveragerði, 45 km towards the east across the Reykjanes volcanic belt (www.vedur.is). It appears as a constant movement in the

geophysical dataset of the Icelandic Meteorological Office and also in the EUREF time-series. This might be interpreted as a steady spreading of the continental plates.
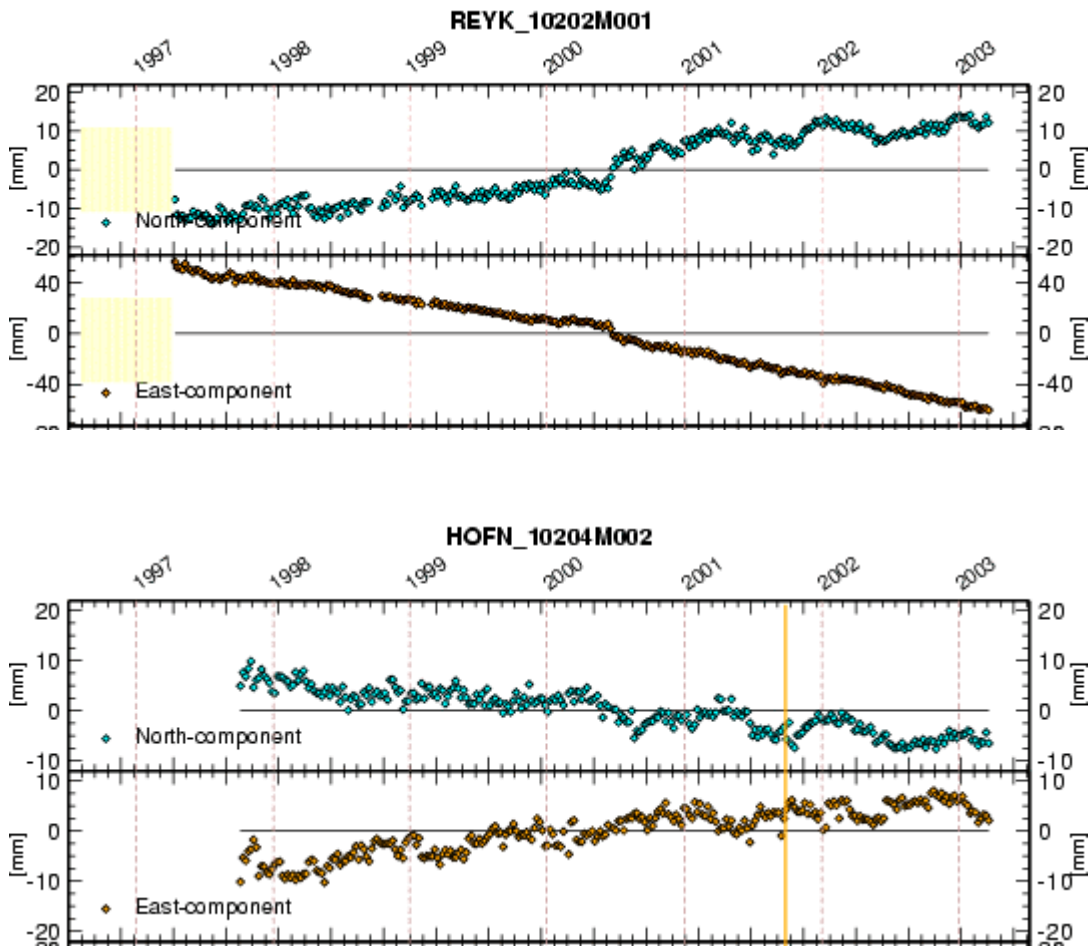

Fig. 4. Improved time series from EUREF Permanent network. The upper part shows the Reykjavík station drifting away from the European mainland. The lower part indicates a slower movement of the Höfn station towards the mainland. Note the different vertical scales. (The pictures are taken from the EUREF website http://epncb.oma.be/organisation.html).

In Iceland an additional opening also seems to be taking place, perhaps giving rise to the excess spreading of the country. It occurs in the so-called rifting episodes accompanying fissure eruptions, and implicitly in the large earthquakes of the transition zones. In the Krafla fires of 1975-1984 in North Iceland the opening of the Krafla fissure swarm was found to be 7.5 m near the centre of the swarm (Wendt et al. 1985, Einarsson 1991, Völksen 2000) and around 5 m on the average for 80 km distance along it. Two other major rifting episodes are known to have occured in North Iceland in historical times, the Askja − Sveinagjá eruptions in 1875 and the Mývatn fires of 1724-1729. These were similar or larger events than the Krafla fires. No direct estimates have been made for the rifting accompanying these later two eruptions. A comparison with the Krafla fires suggest that the total opening of the three events should have been at least 15 m in the Mývatn region. Hence, during the last 1000 years the opening in North Iceland, near the reference line (Fig. 1), has been

18 m of constant spreading (1.8 cm/year) plus 15 m in rifting episodes, or altogether 33 m (3.3 cm/year on the average). How these possible extra movements are transferred across the lithosphere to the edges of the insular shelf remains an unsolved problem.

Conclusion

The primary conclusion is that assumption no. 1 of the model is not valid. The spreading rate in Iceland, along the Gerpir-Tóarfjall line, really appears to be around 3.4 cm/yr. Parts of the Eurasian and North American plates are thus spreading faster than their main bodies. This appears paradoxical in the context of plate tectonics: Iceland seems to be a fast-spreading island on a slow-spreading floor. This process seems to have been operating at least during the last 15 Ma of the geological history of Iceland.

References

Albertsson, K. 1978: Um aldur jarðlaga á Tjörnesi. Náttúrufræðingurinn 48, 1-8.

Albertsson, K. and Eiríksson, J. 1989. K/Ar ages of rocks from the Flatey borehole in the offshore Skjálfandi basin, North Iceland. Jökull 38, 55-60.

Aronson, J. L. and Sæmundsson, K. 1975. Relatively old basalts from structurally high areas in central Iceland. Earth and Planetary Science Letters 28, 83-97

Bagdasaryan, G. P., Gerasimovkiy, V. I., Polyakov, A. I. and Gukasyan, R. K. 1976. New data on the absolute age of Icelandic volcanic roks. Geokhimya 9, 1333-1339. (In Russian).

Bott, M. H. P. 1985. Plate tectonic evolution of the Icelandic transverse ridge and adjacent regions. J. Geophys. Res. 90, 9953-9960.

Cande, C.S. and Kent, D.V. 1995. Revised calibration of the geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. JGR 100, 6093-6095.

Dagley, P., Wilson, R. L., Ade-Hall, J. M., Walker, G. P. L., Haggerty, S. E., Sigurgeirsson, T., Watkins, N. D., Smith, P. J., Edwards, J. and Grasty, R. L. 1967. Geomagnetic Polarity Zones for Icelandic Lavas. Nature 216, 25-29.

DeMets, C. Gordon, R.G., Argus, D.F. and Stein, S. 1990. Current Plate motions. Geophys. J. Int. 101, 425-478.

Duncan, R.A. and Helgason, J. 1998. Precise dating of the Holmatindur cooling event in eastern Iceland: Evidence for mid-Miocene bipolar glaciation. J. Geophys. Res. 103, 12397-12404.

Einarsson, P. 1991: Umbrotin við Kröflu 1975 – 1989. In (ed.) Garðarsson and Einarsson: Náttúra Mývatns. Hið íslenska náttúrufræðifélag, Reykjavík, 97-139.

Everts, P., Koerfer, L.E. and Swartzbach, M. 1972. Neue K-Ar-Datierungen islandischer Basalte. Neues Jahrbuch Geol. Palaeontol. Monatsh.5, 280.

Gale, N. H., Moorbath, S., Simons, J. and Walker, G. P. L. 1966. K-Ar ages of acid-intrusive rocks from Iceland. Earth and Planetary Science Letters 1, 284-288.

Garcia, S, Arnaud, A.O., Angelier, J, Bergerat, F and Homberg, C. 2003. Rift jump process in Northern Iceland since 10 Ma from 40Ar/39Ar geochronology. EPSL 214, 529-544.

Geirsson, H. 2003. Continuous GPS measurements in Iceland 1999 – 2003. Report 03014, Icelandic Meteorological Office, Reykjavik.

Guðmundsson, Á. 2000. Dynamics of Volcanic Systems in Iceland: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge System. Annual Review of Earth and Planetary Sciences 28, 107-140.

Harðarson, B.S., Fitton, J.G., Ellam, R.M. and Pringle, M.S. 1997. Rift relocation – a geochemical and geochronological investigation of a paleo-rift in NW Iceland. Earth and Planetary Science Letters 153,181-196.

Helgason, J. 1985. Shifts of the plate boundary in Iceland: Some aspects of Tertiary volcanism. JGR 90, 10084-10092.

Jancin, M., Young, K. D., Voight, B., Aronson, J. L. and Sæmundsson, K. 1985. Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. Journal of Geophysical Research 90, 9961-9985.

Jóhannesson, H. 1980. Jarðlagaskipan og þróun rekbelta á Vesturlandi. Náttúrufræðingurinn 50, 13-31.

Jóhannesson, H. and Sæmundsson K. 1998a. Geological map of Iceland 1:500.000. Bedrock geology. Icelandic Institute of Natural History, Reykjavik.

Jóhannesson, H. and Sæmundsson K. 1998b. Geological map of Iceland 1:500.000. Tectonics. Icelandic Institute of Natural History, Reykjavik.

Kristjánsson, L., Jóhannesson, H. and McDougall, I. 1992. Stratigraphy, age and paleomagnetism of Langidalur, Northern Iceland. Jökull, 42, 31-44.

Kristjansson, L., A. Guðmundsson and Haraldsson, H. 1995. Stratigraphy and paleomagnetism of a 3-km-thick Miocene lava pile in the Mjoifjördur region, eastern Iceland. Geol. Rundsch. 84, 813-830.

Kristjánsson, L. and Jónsson, G. 1998. Aeromagnetic results and the presence of an extinct rift zone in Western Iceland. J. Geodyn. 25, 99-108.

Kristjánsson, L., Duncan, R. A. and Guðmundsson, Á. 1998. Stratigraphy, paleomagnetism and age of volcanics in the upper regions of Þjórsárdalur valley, central southern Iceland. Boreas 27, 1-13.

McDougall, I., Watkins, N.D. and Kristjansson, L. 1976a. Geochronology and paleomagnetism of a Miocene-Pliocene lava sequence at Bessastaðaa, eastern Iceland. American Journal of Science 276, 1078-1095.

McDougall, I., Watkins, N.D., Walker, G.P.L. and Kristjansson, L. 1976b. Potassium-Argon and Paleomagnetic Analyses of Icelandic Lava Flows: Limits on the Age of Anomaly 5. Journal of Geophysical Research 81, 1505-1512.

McDougall, I., and Wensink, H. 1966. Paleomagnetism and geochronology of the Pliocene-Pleistocene lavas in Iceland. Earth and Planetary Science Letters 1, 232-236.

McDougall, I., Saemundsson K., Johannesson, H., Watkins, N.D. and Kristjansson, L. 1977. Extension of the magnetic polarity time scale to 6.5 m.y.: K-Ar dating, geological and paleomagnetic study of a 3,500-m lava succession in western Iceland. Geological Society of America Bulletin 88, 1-15.

McDougall, I., Kristjansson, L. and Saemundsson K. 1984. Magnetostratigraphy and geochronology of Northwest Iceland. Journal of Geophysical Research 89,7029-7060.

Mussett, A. E., Ross, J. G. and Gibson I. L. 1980. ^{40}Ar/^{39}Ar dates of eastern Iceland lavas. Geophysical Journal of the Royal Astronomical Society 60, 37-52.

Moorbath, S., Sigurdsson, H. and Goodwin, R. 1968. K:Ar ages of the oldest exposed rocks in Iceland. Earth and Planetary Science Letters 4, 197-205.

Nielsen, N. 1933. Contribution to the physiography of Iceland with particular reference to the highlands west of Vatnajökull. Det Kongelige Danske Videnskabs Selskabs Skrifter. Naturvidenskab og Matematik. Afd. 9, Række IV. 5. 105 pp. Copenhagen.

Pálmason, G. 1980. A continuum model of crustal generation in Iceland: Kinematic aspects. Journal of Geophysics 47, 7-18.

Pálmason, G. 1981. Crustal rifting and related thermomechanical processes in the lithosphere beneath Iceland. Geol. Rundsch. 70, 244-260.

Ross, J. G. and Mussett, A. E. 1976. ^{40}Ar/^{39}Ar dates for spreading rates in eastern Iceland. Nature 259, 36-38.

Sella, G. F., Dixon, T. H. and Mao, A., 2002: REVEL: A model for Recent plate velocities from space geodesy. Journal of Geophysical Research 107, EGT 11 1-32.

Searle, R. 1976. Spreading rate in Iceland. Nature 261, 75.

Sigmundsson, F., Einarsson, P., Bilham, R. and Sturkell, E. 1995. Rift transform kinematics in south Iceland: Deformation from Global Positioning System measurements 1986 to 1992. JGR 100, 6235-6248.

Sæmundsson, K. 1974. Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. Geological Society of America Bulletin 84, 495-504.

Sæmundsson, K. 1979. Outline of the geology of Iceland. Jökull 29, 7-28.

Sæmundsson, K., Kristjánsson, L., McDougall, I. and Watkins, N. D.1980: K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. Journal of Geophysical Research 85, 3628-3646.

TuzoWilson, J. 1963. Evidence from Islands on the Spreading of Ocean Floors. Nature 197, 536-538.

Udagawa, S., Kitagawa, H., Gudmundsson, A., Hiroi, O., Koyaguchi, T., Tanaka, H., Kristjansson, L. and Kono, M. 1999. Age and magnetism of lavas in Jökuldalur area, Eastern Iceland: Gilsá event revisited. Physics of the Earth and Planetary Interiors 115, 147-171.

Vine, F. J. 1966. Spreading of the Ocean Floor: New evidence. Science 154, 1405-1415.

Völksen, C. 2000. Die Nutzung von GPS für die Deformationsanalyse in regionalen Netzen am Beispiel Islands. Wissenschaftliche Arbeiten der Fachrichtung Vermessungswesen der Universität Hannover, nr. 237.

Walker, G.P.L. 1974. The structure of Eastern Iceland. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, 155-164

Walker, G. P. L. 1975. Excess spreading axes and spreading rate in Iceland. Nature 255, 468-470.

Walker, G. P. L. 1976. Spreading rate in Iceland. Nature 261, 75-76.

Watkins, N. D., Kristjansson, L. and McDougall, I. 1975. A detailed paleomagnetic survey of the type location for the Gilsa geomagnetic polarity event. Earth and Planetary Science Letters 27, 436-444.

Watkins, N. D. and Walker, G.P.L. 1977. Magnetostratigraphy of Eastern Iceland. American Journal of Science 277, 513-584.

Wendt, K., Möller, D: and Ritter, B. 1985: Geodetic measurements of surface deformations during the present rifting episode in NE-Iceland. JGR 90, 10163-10172.

Wensink, H. 1964a. Secular variation of Earth magnetism in Plio-Pleistocene basalts of eastern Iceland. Geol. Mijnbouw 43, 403-413.

Wensink, H. 1964b. Paleomagnetic stratigraphy of younger basalts and intercalated Plio-Pleistocene tillites in Iceland. Geol. Rundsch. 54, 364-384.

Árni Hjartarson

Appendix

Skagafjörður Valleys – Geological Description

Ph.D. Thesis
Geological Museum
University of Copenhagen

2003

Content

Tables

1   INTRODUCTION

1.1   Organization of the thesis

The following chapters and enclosed geological map form the background for my PhD-thesis at the University of Copenhagen bearing the name *The Skagafjörður Unconformity, North Iceland, and its Geological History*. The whole work is divided into 9 main parts:

- Article 1: The Geology of Skagafjörður Valleys
- Article 2: Tinná Central Volcano
- Article 3: The Skagafjörður Zone – An abandoned Rift Zone
- Article 4: The Pliocene Valleys of Skagafjörður District. (With J. Bonow)
- Article 5: ^{40}Ar/^{39}Ar–Dates from the Skagafjörður Valleys, N–Iceland – Implications for Rift Relocations and the Deep-Sea Sedimentary Record. (With Harðarson and Pringle)
- Article 6: Lava Production in Iceland
- Article 7: Spreading Rate in Iceland
- Geologic map: Skagafjörður Valleys – Bedrock Map
- Appendix: Skagafjörður Valleys – Geological description

The geological investigation introduced in this appendix and the included bedrock map, has been worked out during the last years mainly associated with hydropower planning. The map sheet, Skagafjörður Valleys – Bedrock Map, will be referred to in the following chapters simply as the Bedrock Map. Some of this material has already been published in rapports from the National Energy Authority in Iceland (Hjartarson et al. 1997, Hjartarson and Hafstað 1999).

The appendix contains all the details of the investigation, historical overview, stratigraphic descriptions, definitions, methodology, analysis and various ideas and hypothesis. The articles, on the other hand deal, with the main scientific yields of the investigations. Their material has been extracted from the appendix in later stages of the PhD-project and developed further and increased.

1.2   The study area

In the project the main emphasis is on the geology of the study area, i.e. the area covered by the Bedrock Map. But in many cases the description and discussion go beyond those limits, and then the whole district from Hofsjökull glacier to the tip of Skagi Peninsula is in focus and in some cases the whole country. The study area framed by the geological map covers an area of about 820 km². The region is a mountainous terrain with highest

peaks reaching over 1000 m. Exposures are in general good and in most cases the lithologies can be studied in canyons and gullies. The study area is inside the watershed of Héraðsvötn, the main river of Skagafjörður District. The deep valleys were inhabited much further inland in earlier centuries than today but a few farms are still left. The main farmlands of Skagafjörður valleys are north of the map borders while towards south is the glacier Hofsjökull, an active central volcano with an ice-filled caldera and a summit reaching over 1800 m a.s.l. The climate is considerably mild. The area is located in the precipitation shadow of the glacier and the highest growth limit of birch wood in Iceland is to be found there.
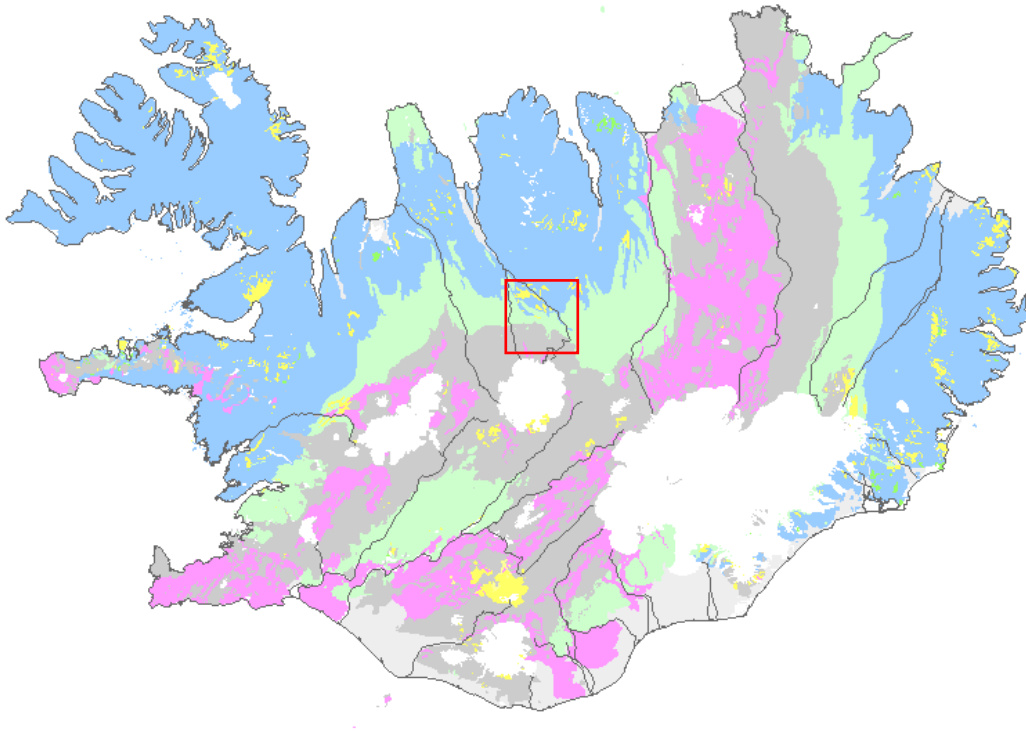
Fig. 1. A Geological map of Iceland. Blue = The Neogene regions. Green = The Plio-Pleistocene areas (3.3-0.8 Ma). Gray = Late Pleistocene. Violet = Holocene lavas. Yellow = rhyolite. The research area is indicated by a red frame. (Modified from Jóhannesson and Sæmundsson 1998)

 The North Iceland volcanic belt is 100 km to the east and the centre of the Iceland Hot Spot is somewhat farther to the southeast. The volcanic pile described here is spanning some 9 million years from late Miocene up through Pliocene and Pleistocene.

The Skagafjörður strata pile is divided into three successions of which the lowest one is separated from the others by a major unconformity:

1. The Quaternary succession
2. The sedimentary succession of the unconformity
3. The Skagafjörður unconformity
4. The Neogene succession
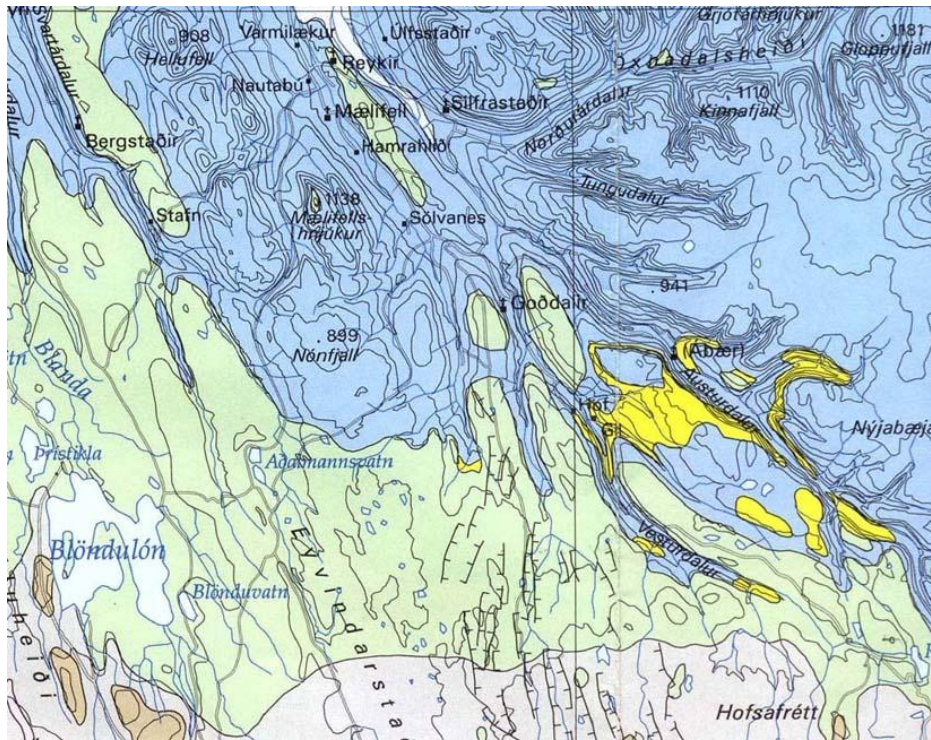
Fig. 2   Skagafjörður valleys, the main features of the geology. Blue = Neogene succession. Green = Plio-Pleistocene. Grey = Late Pleistocene. Yellow = rhyolites of the Tinná Central Volcano. (From Jóhannesson and Sæmundsson 1998).

Briefly, the geological history is as following:

The Neogene volcanic succession below the unconformity originates in an axial rift zone at the boundary between the North American and Eurasian crustal plates. The volcanic belt at the plate margins is 40 – 50 km wide. When the area drifted out of the volcanic zone erosion took over weathering the surface and carving out new landscape. The late Neogene – early Quaternary topography and the initial form of the Skagafjörður valley system can still be seen below younger formations in the District. Later on a sedimentary horizon was deposited in the valleys, tens or even hundreds of meters thick in some places. After that a temporary volcanism started again in the district and lavas of the Quaternary succession were erupted. They flowed along the old valleys, filled some of them up and covered extensive areas. In late Pleistocene the volcanism dwindled out and the glacial erosion became the leading factor creating the current landscape of the district.

1.3 Questions and topics

In the beginning of the work several unsolved questions were put on the agenda for discussion. The Skagafjörður unconformity is the main theme and object of this thesis, as the title indicates, and the questions are aimed at it:

- What is the reason for the Skagafjörður unconformity?

- Is the Skagafjörður Pleistocene succession remains of an axial rift zone, off-rift zone or some other kind of intra plate volcanism?

- How does the Skagafjörður volcanic zone fit into the configuration of spreading axis and ridge jumping in Iceland?

- What is the age or age span of the Skagafjörður unconformity?

- When and for how long time was the Skagafjörður volcanic zone active?

- What does the paleo topography below the unconformity tell about the development of the landscape in Skagafjörður?

- How old are the lowest tillites in the Skagafjörður pile?

- Is it possible to correlate proposed Miocene tillites in East Iceland with tillites in the Skagafjörður stratapile?

- Where does the Neogene volcanic succession in Skagafjörður originate? In the Snæfellsnes – Húnaflói volcanic zone or in the North Iceland volcanic zone or somewhere else?

Answers to most of the questions could be found in the course of the study but not to all of them, as presented in the following chapters. Unexpected dilemmas and puzzles turned up and numerous new questions appeared so at the end of the project more problems were unsolved than could be foreseen in the beginning.

1.4   Previous work

Geological research in the Skagafjörður valleys began already in the 18th century, but has all the time been sporadic both in space and time. The pioneers in Icelandic geography and geology, *Eggert Ólafsson and Bjarni Pálsson,* came to Skagafjörður in 1755. They investigated lignite layers in the Austur- and Vesturdalur valleys. In their Book of Travels they conclude that the best lignite of North-Iceland is in Hofsgil gorge (Jökulsárgil) in Vesturdalur. The lignite from there was used in iron-work in the valley. In Austurdalur they investigated Tinnárfjall mountain where they found a 5 feet thick layer of "black jasper" (Ólafsson 1943, Ferðabók II, p. 25,37). In those days great attention was paid to the lignite in spite of its poor quality because many thought it might give indications to hidden coal deposits in the mountains.

Fig. 3. Sveinn Pálsson

Sveinn Pálsson (1762 – 1840), naturalist and doctor, visited Skagafjörður valleys in 1792. In his diary he describes sandstone and coarse conglomerate layers in Tungusveit District (north of the enclosed map) later known to be part of the sediments of the Skagafjörður unconformity. In Austurdalur he climbed Tinnárfjall to check out the investigations of Ólafsson and Pálsson. There he describes a prominent layer of columnar basalt and underlying sediments. He finds the "black jasper" but thinks it might as well be obsidian. In Ábær gorge he sees a thick layer of "white clay (Argiella)". He also mentions the lignite in Hofsgil and says that it is used in a small scale like fire-wood. (Pálsson 1983, p. 118-123). Now we know that the columnar basalt is the Austurdalur Quaternary Volcano, the "black jasper" the obsidian base of the Skati rhyolite dome and the "white clay" is fine grained acid tephra at the base of Ágúll rhyolite dome.

Fig. 4. Jónas Hallgrímson

 Jónas Hallgrímsson (1807 – 1845), Iceland's beloved poet, was an educated naturalist with highest degrees in geology and mineralogy from the University of Copenhagen. He visited Skagafjörður County in August 1839 to observe supposed coal seams in Austurdalur and Tinnárdalur valleys for the Danish Government. There he took some rock samples and measured a section through the lignite seams and the adjacent layers. In his diary he writes: "Although these findings are probably not going to be of any practical use, as the coals are of low quality as fuel, then I however hope that they will be considered to be noticeable from scientific point of view because inside the lower two seams imprints can be seen of marine plant remains (among others *Zostera marina?*) – at high elevations, 8 miles from the shore – indicating a recent upheaval of the land. It is also remarkable that the

uppermost layer is very unlike the lower ones (wood remains, birch?) with still younger voluminous columnar lava resting on top of it" (Hallgrímsson 1989, p. 353-356). Hallgrímsson never developed these ideas any further but in his report to the government the seam is suggested to be a thin lignite layer of poor quality and too little quantity to be of any practical value. After this no geological investigations were made in Skagafjörður Valleys for more than hundred years.
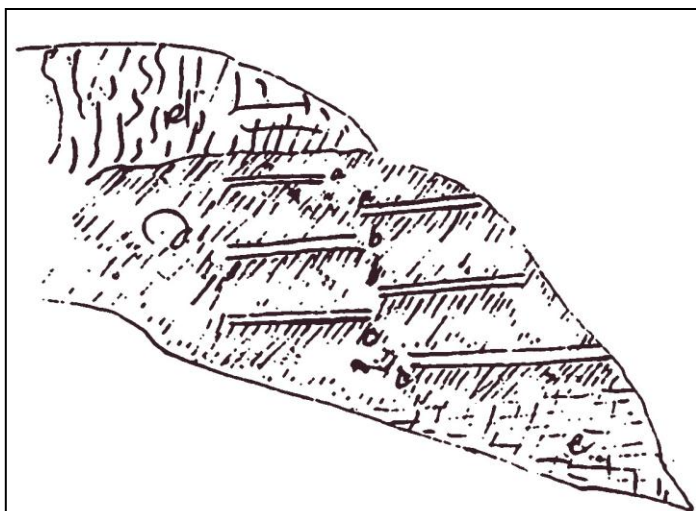
Figure 5. Section across the lignite layer in Illagil gorge in Tinnárdalur. From Hallgrímsson's diary in august 1839 (Hallgrímsson 1989 p. 356).

a, b, c. lignite

d. sedimentary layer

e. columnar basalt

Helgi Péturss (1872 – 1949) was one of the first educated geologists in Iceland. He finished a degree in geology at the University of Copenhagen 1897 and defended a doctorial thesis "Om Islands Geologi" in 1905. There is a whole chapter about the geology of Skagi peninsula and the east coast of Skagafjörður but he never visited Skagafjörður valleys. He describes the sedimentary horizon of the unconformity and notes the relatively young age of the basaltic lavas and plugs in the district. He recognized till horizons in between the lava layers and concluded that the ice age could be divided into glacial and interglacial periods. That was a scientific innovation in those days when most geologists supposed the ice age to have been a period of stable cold climate and permanent glaciation.

Jakob H. Líndal (1880 – 1951), a farmer and a well known self-educated geologist in Iceland. He visited Skagafjörður valleys in 1938 and 1939 and investigated strike and dip, lignite layers and stratigraphy and discovered the unconformity. In his notebook he described the different dip of the strata above and below it and the thick, coarse sedimentary layer at the unconformity. He studied the origin of the sediments and concluded that they were partly of fluvial and partly of glacial origin. He found the Pleistocene volcanic sites in Mælifellshnjúkur summit and in the island of Drangey (Líndal 1940, 1941, 1964).

Fig. 6. Jakob H. Líndal

Trausti Einarsson (1907 – 1984) was a professor in physics and geophysics at the University of Iceland. He made important investigations on the geology and germorphology of the Skagafjörður District and he also performed a pioneering work in paleomagnetic measurements (Einarsson 1958, 1959, 1962). He came to the conclusion that an evolved valley system had already developed in the area in late Tertiary times before the onset of the glaciation. He found out that the valleys had an erosion level near a plane that now is at 300 m a.s.l. He concluded that this was the late Tertiary sea level. At this time a short volcanic period was ignited and lavas flowed along many of the valleys and filled some of them up. At the end of Tertiary he supposed that a major uplift took place and during Quaternary a new and deeper valley system were formed, mostly by glacial erosion. Trausti Einarsson was in no way a mainstream scientist and his theories, and especially his uplift hypothesis, were never taken into account by Icelandic geologists. But as will be seen here in later chapters he might have been underestimated.

Peter Everts, a German geologist, published a geological maps (black and white) along with a detailed description of the northern part of Skagi Peninsula and the east coast of Skagaförður, including chemical rock analyses and K/Ar dates (Everts 1972 and Everts et al. 1975). This is a basic work on the geology of the northernmost areas of the Skagafjörður district with special emphasis on the petrology and geochemistry.

Haraldur Sigurðsson and his coworkers described the petrology and structure and geochemical variations in the area between Skagi and Langjökull glacier and proposed that 2.5 million years ago rifting in the Reykjanes − Langjökull Zone extended northward through 15 km thick crust, giving rise to an ephemeral Skagi Zone. They also proposed that rifting in Skagi was related to a period of increased spreading in Northern Iceland 0.5-2.5 million years ago (Sigurðsson 1970, Sigurðsson et al. 1978 and Schilling et al. 1978). This idea has not been taken in account by later authors. The main trend has always been to regard the Skagi volcanic area as an intra plate volcanism. In this work the idea of ephemeral rift zone in Skagafjörður district is adapted and developed.

Kristján Sæmundsson, Leó Kristjánsson, Ian McDougall and Norman D. Watkins published in 1980 an article on K/Ar dating and geological and paleomagnetic study in Tröllaskagi Peninsula. The southwest end of their area overlaps the northern limit of Skagafjörður Valleys. This work has served as the base for the chronology of the Tertiary region in North Iceland.

Between 1970 and 2000 diverse research was performed on the stratigraphy, local tectonics and geothermal activity of the Skagafjörður valleys (Sæmundsson 1974, Kaldal and Vikingsson 1978, Hardarson and Gudmundsson 1986, Karlsdottir et al. 1991, Jóhannesson 1991, Hjartarson et al. 1997, Hjartarson and Hafstad 1999).

1.5   The field work

In recent decades plans have been made to harness the hydropower of the main rivers of Skagafjörður; Héraðsvötn, Austari Jökulsá and Vestari Jökulsá. Various ideas and implementations have been introduced. The basic element in all the harnessing ideas is a reservoir with water level at 700 m a.s.l. in the Bugar area north of Hofsjökull and a power station somewhere in the Skagafjörður valleys at ca. 100 m a.s.l. In a geological exploration study for National Energy Authority in Iceland I and my colleagues started

investigations on the geology of Skagafjörður valleys in the summer of 1992. The first field trip was to observe the hydrogeology of the area. But the aim of the later work was to make geological maps and sections and rock evaluation in possible tunneling sites in the district. One to two weeks were spent in the field each year. In the summer of 1993 Nýjabæjarfjall and lower Austurdalur valley were observed and mapped in cooperation with Þ.H. Hafstad and G.Ó. Friðleifsson. In the summers of 1995 and 1996 the upper Austurdalur valley was mapped along with G.Ó. Friðleifsson. In the summer of 1997 lower Vesturdalur valley was investigated and mapped. In the summer of 1998 the Hofsafrétt highlands were researched in cooperation with Þ.H. Hafstad and in the summer of 2000 the Bugar area was mapped in cooperation with Þ.H. Hafstad. In all this process and fieldwork the main emphasis was on the geotechnology, rock evaluation, engineering problems and mechanical characteristics of the stratigraphy.

1.5.1   The PhD-project

The PhD-project started formally in January 2000. Then the first outlines of the thesis were defined. I intended to use the Skagafjörður data, collected during last few years, to dig deeper into the geology of the area. Instead of the geotechnology of the rocks and strata pile the focus should be on the Skagafjörður unconformity, its nature and geological history. My mentors should be Asger Ken Pedersen, docent at the Geological museum in Copenhagen as supervisor, Kristján Sæmundsson, geologist at Orkustofnun as supervisor abroad and Svante Björck, then professor at the University of Copengagen, but now at the University in Lund, as vice supervisor. The project was divided into three main components. First the Neogene succession should be treated, then the Pleistocene succession and finally the section in between, the unconformity and its related sediments. According to this the lecture in the first PhD-day, in May 2000, dealt with the most prominent formation of the Neogene pile, Tinná Central Volcano.

1.5.2   Activities in 2000

Two field trips were made to the research area in the summer of 2000. In the first the sedimentary layer at the unconformity in Vesturdalur Valley and Goðdalafjall Mountain were studied. There an unexpected and important discovery was made when shell fragments of balanus sp. were found at 350 m a.s.l in the mountain slope near Goðdalir farm (unfortunately not in situ). This is by far the highest and farthest inland marine fossil site in Iceland and could possibly throw new light on the geological history of Skagafjörður Valleys. Repeated attempts to find more fragments or other fossils have been in vain.

In this field trip dykes cutting the unconformity and its sedimentary succession in Vesturdalur were found, indicating Pleistocene volcanic activity in valley. Along with the research in the field rock samples, mainly from lavas sandwiching the sediments of the unconformity, were sampled for Ar/Ar dating in cooperation with B.S. Harðarson. Chemical analyses and thin sections were used to choose the appropriate samples for the dating.

In August I participated the Iceland Summer School on Plume-Ridge Interaction held in cooperation of RIDGE and NORDVULK at Mývatn, August 20-30. There I exhibited a

poster on the Tinná Central Volcano and gave a short lecture on the Neogene geology of Skagafjörður Valleys. Expeditions were made to Krafla and Askja and the terrestrial part of the Tjörnes fracture zone.

Figure 7.

The NORFA-course Igneous and low – T metamorphic processes at volcanic rifted margins. Porsild leaves Godhavn July 3. 2001

1.5.3   Activities in 2001

In May 2001 I stayed at the Museum of Geology in Copenhagen for three weeks and gave the lecture *The Skagi Zone – An Aborted Rift Zone* at the PhD-day. It dealt with the Pleistocene succession, the initiation, culmination and sudden decline of the volcanism and spreading

In June 27th – July 10th I participated the NORFA-course *Igneous and low – T metamorphic processes at volcanic rifted margins.* It started with lectures in Copenhagen that were carried on in Godhavn after arriving Greenland via Iceland 30th of June. After several days in Godhavn the expedition went up north to Ubekendt Ejland to investigate the well known volcanic formations there, picrite lavas of the Vaigat formation, layered intrusion and evolved eruptive rocks. After some days in Ubekendt Ejland the expedition sailed to Svartenhuk peninsula where traces of oil have been found in a pillow-lava. After that the famous but abandoned mines in Maarmorilik were visited and the historical site where Wegener started his final expedition to Camp Ismitte. The NORFA course was well planned and very successful, informative lectures, important geology, good company and wonderful weather. For me the comparison between the Greenlandic and Icelandic volcanic piles and their striking difference were the most interesting thing.

A field trip to Skagafjörður valleys in July together with Kristján Sæmundsson and Hafdís Eygló Jónsdóttir proved to be important. Kristján had read the manuscript of my article *The Skagi Volcanic Zone – an abandonned Rift Zone,* and after the trip he could make valuable critical comments and give stimulating advice. Several things were investigated during the trip. The most important ones were the tectonics of the Pleistocene succession and we came to the conclusion that they were clearly rift related. We also investigated some geothermal fields and measured fissures and studied their

connection to the proposed Pleistocene rifting of the Skagafjörður Zone. At last we studied the fossil site in Goðdalafjall and tried to find more shell fragments but without success.

Fig. 8. Longyearbyen, Spitsbergen in late August 2002

1.5.4   Activities in 2002

The year of 2002 was a period of continuous study and writing. Then I quit my full-time job at Orkustofnun and took a year off to finish the PhD project. Most of the writing was completed at the Geological Museum in Copenhagen but the field season in Iceland also was important. While writing the thesis several ideas came up that required a field investigations in the summer 2002. Some of the ideas turned out to be pure visions but others proved to be more stable. New sites were also visited as Mælifellsknjúkur peak and the Pleistocene formations in Norðurárdalur. A new project was started in cooperation with Leó Kristjánsson on the paleomagnetic stratigraphy of the Neogene pile. Preliminary results came already at the end of September and could be included in the thesis.

In fall, the last one of the necessary courses along with project was completed at UNIS in Spitsbergen. It had the name Artic Terrestrial and Marine Quaternary Stratigraphy and was organized by Ólafur Ingólfsson, Jan Mangerud and Michael Houmark-Nielsen. An expedition for 7 days was made on Lance, the vessel of the Norwegian Polar Research. Several key sites of the Quaternary geology of Spitsbergen were visited and on the way to the small research town Ny Ålesund the ship crossed the latitude 79°N.

The Ar/Ar dates (preliminary results) from Björn S. Hardarson were received in September and they threw in many respects new light on the matter and older ideas about the age and origin of the strata pile had to be reviewed.

Fig. 9. The canyon of Austari Jökulsá river

1.6 Methods

The classification of the bedrock, as it is presented on the geological map, is mainly based on hand specimens studied in the field during geological mapping. But chemical analyses and thin sections have also been used for checking the classification. The field classification used has been modified from those of Walker (1959) and Jancin (1984) (table 1). The prevalent rock type in the Skagafjörður area, as in Iceland as a whole, is basalt. A three-fold subdivision of the basalt is used; tholeiite basalt, porphyritic basalt and olivine basalt. In the porphyritic basalt the phenochrysts of plagioclase and/or pyroxene must make up at least 3% of the rock volume. Lavas with phenochrysts lower than 3% are called scattered or sparsely porphyritic and they can be classified as tholeiite or olivine basalt with scattered phenocrysts. Gradations exist between the three main types but as a rule they are distinctive enough in the field to provide a mappable stratigarphic units. The olivine basalt can be divided in two main groups; regional olivine basalt flows and compound olivine basalt. The latter are thought to originate in shield volcanoes. They are made of olivine tholeiite magma of low viscosity that flows along in thin lobes and build up lavas composed of several lava bands.

The more silicic lavas are associated with central volcanoes. In the field, basaltic andesite may be qualitatively distinguished from andesite on the basis of the latter's smaller wavelength of flow folds, darker fresh color, flinty fracture with a hammer and typically

aphyric petrography. The dacite and rhyolite lavas have more abundant phenocrysts than andesite. Rhyolite has a high vesicularity, black pitchstone, ore obsidian rich zones may occur at the base and top of the flow.

The distinction between lava types often is difficult because a complete gradation is seen from the porphyritic basalts through rhyolite. Transitional lavas are common.

Table 1. Field characteristics of lava types (Modified from Walker 1959 and Jancin et al. 1985)

Rock type	Grain size of groundmass mm	Phenocryst mineralogy	Flow planes	Vesicles	Weathered color	Color on fresh surface	Weathering morphology
Porphyritic basalt	Variable >0.25	> 3% PL, ol, cpx	poor-none	usually oval, abundant at base and top	brown, reddish-brown	medium dark gray	subrounded to rounded, RESISTANCE
Scattered porphyritic basalt	0.1-1.0	0.5-3% pl, ol, cpx	poor-none	spherical-oval, common	reddish brown-greenish gray	medium dark gray	subrounded, blocky
Olivine tholeiite	0.05-1.0	OL > 1%, pl, cpx may be present	NONE	spherical-oval, pipes common	dark green-greenish brown, reddish brown-medium gray	DARK GREY WHEN ALTERED, medium or dark gray	SPHEROIDAL, subrounded, blocky
Tholeiite	0.02-0.1	< 1% pl, cpx	moderate to WELL DEVELOPED planes >1 mm width	usually elongate or irregular, not abundant	reddish brown-greenish gray	light to dark gray	BLOCKY, subrounded, flaggy
Basaltic andesite	0.05	pl, ol < 2%	WELL DEVELOPED, somewhat contorted	irregular, rare	variable, VARNISHED APPEARANCE	medium dark to gray	BLOCKY, subrounded, flaggy
Andesite	0.005-0.05	rare	WELL DEVELOPED < 5 mm WIDTH, flow folding in outcrop	rare	dark brown-red gray	very dark gray	BLOCKY TO PLATY
Dacite	0.005-0.05	TOTAL>1% ol, cpx, opx, pl, Fe-Ti ox	WELL DEVELOPED, flow folding in hand specimen possible	rare	reddish brown-dark brown	medium gray	PLATY TO BLOCKY
Rhyolite	<0.01	similar to dacite	WELL DEVELOPED, flow folding in	abundant, irregular, 1 mm across	yellow brown-pinkish brown	very light gray	FLAGGY PILES

Graine size estimates by inspection in thin sections cut from samples representative of central portion of flow. Most distinctive characteristics in ALL CAPITALS. pl = plagioclase, ol = olivine, cpx = clinopyroxene (augite), opx = orthopyroxene (hyperstene), ox = oxide

Sedimentary layers are divided in three classes; siltstone - sandstone, conglomerate and breccia (including till). The most common ones are the thin red interbeds. They are made of sandy or silty tuffaceous material. The origin of these interbeds is thought to be aeolian, i.e. wind blown ash that has suffered chemical weathering towards laterite. Their thickness is most often 0 - 1 m. They represent the old soil that formed on top the lavas in between the eruptions and are most often too thin to be shown on the geological map.

Dykes are mapped but neither subdivided according to rock type, thickness nor paleomagnetism.

Magnetic measurements have been carried out with fluxgate magnetometers. Usually three samples from each lava layer are measured but more if the magnetic polarity is weak or fluctuating. Such measurements have been performed for over 30 years in Iceland and have been very helpful in local mapping projects all over the country. However experience has shown that they should be handled with care because the field method does not always yield the correct primary magnetic polarity of the rock. One section inside the map was cored and the paleomagnetism observed in a laboratory (Bakkadalur, section PG, Sæmundsson et al. 1980). A new paleomagnetic project was established in the summer 2002 in a teamwork with Kristjánsson and Guðmundsson.

On the Bedrock Map the rocks have been classified according to rock type, paleomagnetic polarity and age. The colors are according to the standard legends of Orkustofnun, National Energy Authority. Blue, green and blue-green colors show the basalt types, the pink colors indicate the andesites and the yellow ones the rhyolitic rocks. Hyaloclastites and sedimentary horizons are shown in brown and orange. Dykes and intrusions are red. Unconsolidated superficial deposits are only shown where they have considerable thickness and extension. They are in gray colors (Vilmundardóttir and Sigurðsson 1984).

1.7   Stratigraphic Systems and Terminology

Nomenclature and use of stratigrapic terms in the bedrock geology in Iceland has been rather subjective. No strict rules or standards have been established. In the beginning of the 20th century Icelandic geologists divided the whole country into three formations, the Blue-Basalt Formation (blágrýtismyndun), the Older Grey-Basalt Formation (eldri grágrýtismyndun) and the Younger Gray-Basalt Formation (yngri grágrýtismyndun).

G. P. L. Walker, a pioneer in the study of the Tertiary stratigraphy of Iceland, used the terms succession or sequence as his biggest units and introduced them for the whole strata pile of the area. He divided the succession into groups such as Gerpir – Barðsnes acid group, Víkurvatn olivine basalt group, Vindás porphyritic group, Hómatindur basalt group and so on (Walker 1959). As a whole his nomenclature is informal. Most Icelandic geologists have also been very informal in their terminology using successions and sequences for thick regional strata piles and series, suites, groups, formations and members for their sub-division in a free style. Formal stratigraphic terminology and classification according to international standards as the International Stratigraphic Guide (Salvador 1984) or the North American Stratigraphic Code (NACSN 1983) has more or less been avoided.

Table 2. Classification of stratigraphic units*

A. Material units

Lithostratigraphic Units	Lithodemic Units	Magnetopolarity
Supergroup	Supersuite	
Group	Suite	Polarity Superzone
Formation	Lithodeme	Polarity Zone
Member (Lens, Tongue)		Polarity Subzone
Bed(s) or Flow(s)		

B. Chronostratigraphic Units

Chronostratigraphic Units	*Geochronological Units*	*Polarity Chrono-stratigraphic Units*	*Polarity Chrono-logical Units*
Eonothem	Eon	Superchronozone	Polarity Superchron
Erathem	Era		
System	Period	Chronozone	Polarity Chron
Series	Epoch		
Stage	Age	Subchronozone	Polarity Subchron
Chronozone	Chron		

*The tables are adapted from the North American Stratigraphic Code published by North American Commission on Stratigraphic Nomenclature (NACSN 1983). They are slightly modified to harmonize with the IUGS standards (Salvador 1984).

The only strata sequence in Iceland that has been given a formal status, with type localities and stratotypes, is the Tjörnes sequence (Eiríksson 1981, Eiríksson et al. 1990). The uppermost units are Höskuldsvík Group and Breiðavík Group. The latter one is 600 m thick, made of lava flows and sediments. It is divided into six formations that are further subdivided into members. Compared to the nomenclature and stratigraphic terms that have been introduced in the Tertiary regions in East and West Greenland (Larsen et al. 1989, Pedersen 1985, Pedersen et al. 2002) the Icelandic units are small. The formations in the Greenland basalt successions are similar to the Icelandic groups. In this thesis the stratigraphic classification listed in table 2 will be used. Informal stratigraphic categories, as succession, sequence, assemblage, units and horizon will also appear in the text.

Fig 10. The North American Commission on Stratigraphic Nomenclature was established 1946. The chief publication of the Commission is the North American Stratigraphic Code, the latest complete version of which was published in 1983. It has served as a guide to all stratigraphic practices throughout North America.

IUGS (International Union for Geological Sciences) operates the International Subcommission on Stratigraphic Classification. It published the International Stratigraphic Guide (Salvador 1984). It is expected gradually to supplant number of national and regional stratigraphic codes. These two stands are similar in nature, content, and stratigraphic philosophy and serve as a basis for the stratigraphic nomenclature in this work.

Lithostratigraphic units are the basic units of general geologic work and serve as the foundation for delineating strata, local and regional structures and economic resources in regions of stratified rocks. They are recognized and defined by observable rock characteristics; boundaries may be placed at clearly distinguished contacts or drawn arbitrarily within a zone of gradation.

Formation is the fundamental unit in lithostratigraphic classification and the base for describing and interpreting the geology of a region. It is a body of rock classified by lithic characteristics, distinctive physical and chemical features and stratigraphic position. It should not be defined by geological time. It is prevailingly, but not necessarily, tabular and is mappable at the Earth's surface or traceable in the subsurface. Its limits normally are those surfaces of lithic change that give it the greatest practicable unity of constitution. A formation may represent a long or short time interval, may be composed of materials from one or several sources and its thickness can range from less than a meter to several thousand meters. Names of formations will be taken, according to international tradition, from geographical names of places where they were originally described and combined with names of the predominating rock comprising it, e.g. Skati rhyolite formation. Names of groups are only geographical.

Stratotype, or type section, is a particular sequence of strata chosen as a standard of reference for a stratigraphic unit. In the following chapters stratotypes will be defined for most of the groups in form of *reference profiles* with precisely defined location and upper and lower boundary. In spite of this formality the strata pile in Skagafjörður valleys will not be given formal lithostratigraphic definitions for the moment. Informal stratigraphy will remain in this thesis.

The Skagafjörður main successions (the Neogene succession and the Quaternary succession) are divided into groups. Each group is further divided into formations according to common lithic characteristics as lava type, stratigraphic position and polarity (table 2). The formations are split into members where necessary, especially in the Quaternary succession. Each group, formation and member has been given a name. On

the Bedrock Map indicated formations and members have also their own identities, two or three letters. The names should be permanent but the identities might vary from a map to map. The Fossárdalur group is an example. It is divided into three formations:

ge   Geldingsá formation

hö   Hölkná formation

fo   Fossárdalur formation

Geldingsá formation is split up in two members, Geldingsá clastic member and Geldingsá porphyritic member (fig. 11). Geldingsá porphyritic member can be divided into beds f. ex. scoria and dense rock and so on.

Fig. 11. Lithological terminology for the rock stratigraphy as used in this work (example); succession (supergroup) – groups – formations – members. Based on the International Stratigrapic Guide (Salvador 1984).

The Skagafjörður succession is summarized in a geological column on the included geological map showing an ideal section through the whole pile. There the thickness of groups and formations are indicated, the paleomagnetic polarities and reversals and the relative age of the rocks. The following description of the strata pile will start in the oldest formations of the area. The lithostratigraphical groups will be discussed one by one until the youngest formations, near the southern limits of the area, has been treated.

2 THE NEOGENE SUCCESSION

The term Neogene is here used instead of the more common term late Tertiary. Tertiary is divided into two sub-periods; Palaeogene (Palaeocene, Eocene, Oligocene) and Neogene (Miocene, Pliocene). The terms Palaeogene and Neogene have been preferred by IUGS as the correct terms in formal stratigrapic description. The Neogene succession was mapped and described by Hjartarson et al. (1998). The geological names and identifications of formations are mostly taken from their work. At the end of each chapter a location of a reference profile or stratotype for each formation will be defined. Table 11 gives an overview on the names of groups, formations and members of the Neogene succession.

2.1 Sólheimar Group

Sólheimar group is composed of two formations, Sólheimar formation and Merkidalur formation. Its name derives from the Sólheimar farm and the Sólheimar mountain several kilometers north of the included Bedrock Map (Sólheimar means "Sunny World"). The group rests on the prominent Bóla sedimentary layers and forms a 1200-1300 m thick pile predominantly made of tholeiite lavas. Only the upper part of the group is inside the field of investigation. Bóla sedimentary layers mark the top of marine magnetic anomaly 5 (epoch 9) (Sæmundsson et al. 1980) and according to the paleomagnetic time scale of Cande and Kent it is 9.7 Ma.

2.1.1 Sólheimar Formation

At the northern border of the research area (shown on the Bedrock Map) the rocky canyons of the two great glacier rivers, West- and East-Jökulsá (Vestari og Austari Jökulsá), join to form Héraðsvötn, the main stream of Skagafjörður District. The oldest rocks in the area are to be found in the canyon at the confluence. They belong to the upper part of the Sólheimar formation. Only the uppermost part of the formation is inside the Bedrock Map, dominated by thin-layered tholeiite lavas with thin sedimentary interbeds. The sediments are most often red or red-brown in colour, made of sandy or silty tuffaceous material of aeolian origin, i.e. wind-blown ash that has suffered chemical weathering towards laterite. Their thickness is most often 0 – 1 m. They represent the old soil that formed on the lavas in between the eruptions. A small basic intrusion is found near the top of the formation in the north end of Hlíðarfjall mountain.

The lavas have reverse polarity (R) but at the top of the formation the magnetic direction becomes irregular.

Three K/Ar-age determinations have been made on lavas at the upper limit of the formation at 900 m a.s.l. in Sólheimar mountain (table 21). They gave around 9 Ma or late Miocene age (Sæmundsson et al. 1980).

Reference profile for the Sólheimar formations is the profile PF in Sólheimafjall mountain described by Sæmundsson et al. (1980). The base is at N/R magnetic reversal at 360 m a.s.l. Farther south the Bóla sedimentary layers are seen at the base of the formation. The top is at R/N magnetic reversal at 970 m a.s.l. These sites are north of the Bedrock Map. The entire profile has been carefully measured and sampled, paleomagnetic profile and radiometric dates have been published.

2.1.2   Merkidalur Formation

The Merkidalur formation is divided into three members. The lowest part (me1) is a direct continuation of the Sólheimar formation. It is 180 m thick, made of thin-layered tholeiite lavas with red interbeds. The polarity is normal (N).

The central part (me2) is 170 m thick. It is more diverse than the lowest one with olivine and porphyritic basalts and sedimentary layers along with the tholeiites. In Merkidalur valley thick layers of basaltic scoria indicate a close neighbourhood to a volcanic vent. Polarity measurements in the field show reverse polarity (R) as a main trend in the member with some irregularity near its middle part. It was detected in the field and laboratory measurements gave reverse direction with one normal layer in the middle (Sæmundsson et al. 1980).

The sedimentary member, Merkidalur sediment (me3), is made of 5 - 40 m thick, horizontally layered, brown sandstone, rich in cobbles and pebbles. It is found in Merkidalur and in the canyon of Austari Jökulsá river. It is thickest in outer Merkidalur valley (40 m) and there it is among the thickest sedimentary layers of the Neogene pile, only the Skati tephra layer is thicker. This layer is supposed to be related to an unconformity (the Hidden Unconformity) that represents the divide between the lava piles of the North Iceland rift zone and the older lavas of the Snæfellsnes – Húnaflói zone. This will be discussed in more details in the chapter of paleomagnetism and age (chapter 9).

 All these above mentioned members have been found in East-Jökulsá canyon (Björnsson and Friðleifsson unpubl.).

Reference profile for the formation is the Bakkadalur PG-profile of Sæmundsson et al. (1980, fig. 2). The base is R/N magnetic reversal at 460 m a.s.l. and the top is at the upper level of the thick Merkidalur sediment (me3) near 750 m a.s.l. Paleomagnetic sampling and measurements have been performed and published for the lower part of the PG-profile and the upper part is under preparation.

2.2 Tinná Group

The Tinná group is indicated by the acid and intermediate rocks of the Tinná Central Volcano (fig. 14). It has around 800 m thickness in Austurdalur and is divided into 11 formations (table 11). It takes its name from the Tinná river, a tributary to Austari Jökulsá. The group contains three series defined by Hjartarson et al. (1998), the Ábær, Tinná and Nýibær series. This division will be reviewed. The bottom of the group is at the top of the Merkidalur sediment and is exposed in Merkidalur and in the canyon of Austari Jökulsá. The Tinná group covers extensive areas in the mountains at both sides of Austurdalur and Vesturdalur valleys but disappears in to the pile due to the general dip in the inner Austurdalur near the Fossá tributary river and in Vesturdalur near Hraunþúfuá tributary river.

2.2.1   Fjóslækur Formation (áb1 and áb2)

Fjóslækur formation is the lowest formation of the Tinná group. It is divided into two units, a tholeiitic lava pile (áb1) and a porphyritic and olivine basalt formation (áb2). They interfinger in the westernmost slopes of Merkigilsfjall and their boundary is far from clear-cut. Hjartarson et al. (1998) defined these members as the lower part of their Ábær formation that here will be dismissed.

 The tholeiitic lava pile (áb1) is found in Merkidalur valley above the Merkidalur sediment. It is best described in the Bakkadalur profile (PG) of Sæmundsson et al. (1980) where it comprises at least 250 m thick assemblage of lavas with normal polarity (N).

In Austurdalur and Vesturdalur valleys these lavas are replaced by the Fjóslækur porphyritic and olivine basalt formation (áb2). In Austurdalur south of Merkigil farm it is 350 m thick. In Vesturdalur it is 125 m thick. The Fjóslækur formation is dominated by normal paleomagnetic polarity (N) with two short reverse subchrons (R). The lower one, that contains only two lavas, wedges into the pile in outer Goðdaladalur canyon. The other one is found at the top of the formation. There are a few reverse lavas of olivine and porphyritic basalts and a sedimentary horizon of siltstone, sandstone and fine-grained conglomerate.

Reference profile for the Fjóslækur formation is in the ravine of Fjóslækur. The base of the Fjóslækur profile is at a sedimentary layer (me3) in Jökulsá canyon but the top at the upper level of a thick scoria layer in 600 m a.s.l.
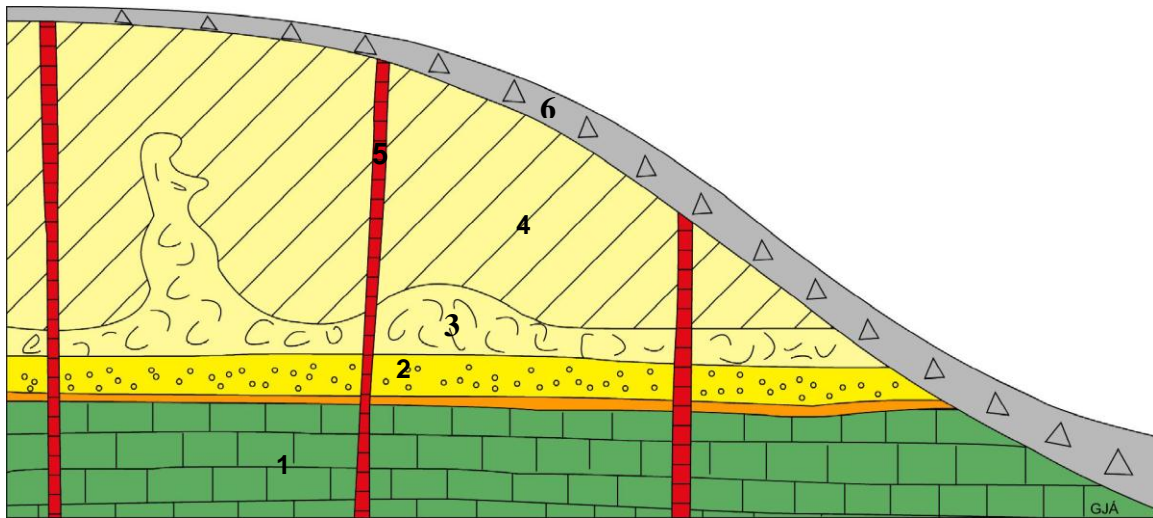
Fig. 12. The Ágúll rhyolite dome, a cross-section along the strike in Ábær river gorge. 1) Olivine tholeiite lavas. 2) Acid tephra. 3) Basal breccia. 4) Rhyolite dome. 5) Dyke. 6) Till cover.

2.2.2   Ágúll Rhyolite Formation (áb3/áb4)

At Ábær farm in Austurdalur a rhyolite lava dome (or a coulee) appear in the excavated walls of the river canyon. It is at least 250 m thick where it is thickest. This rhyolite is the lowest one of several acidic layers in the Skagafjörður valleys and it indicates the formation of a magma chamber underneath the area and the initiation of the Tinná Central Volcano. (See fig. 14 for location).

The Ábær tributary river flows in rapids and cascades in a deep gorge from the hanging Ábær valley to form a broad alluvial fan at the east bank of Jökulsá river in Austurdalur. At its north side the rugged landslide of Brennigilshólar (2,0 km²) covers the bedrock but the walls of the gorge are eroded into the acid rocks of the dome. Sveinn Pálsson was the first scientist to mention this place as early as 1792 when he described the white clayish material in the gorge.

The lowest layers encountered in the gorge are olivine basalt lavas of Fjóslækur formation. Above them are the acid layers of the dome. In Ábær gorge they can be divided into three units. The lowest one is a light coloured, fine bedded, sandy tephra, 15 - 20 m thick with coarser pyroclastic layers of angular stones, most often 2 - 4 cm in diameter but can reach up to 50 cm. The upper part of the tephra layer is coarser than the lower one.

Above the tephra, at 280 m a.s.l. near Ábær farm, brecciated rhyolite and acid scoria is seen. Angular rhyolite blocks, up to 1 m in diameter, are sitting in more fine-grained material. This unit attains a thickness of 10 m and forms the basal breccia of the massive lava dome. Somewhere this breccia forms huge pillows rising many meters up into the lava itself.

Fig. 13. The Ábær gorge. An acid lava dome resting on light grey tephra.

The lowest 60 m of the lava are seen in the gorge. The upper part is more or less capped by screes and tillite but reaches 500 m a.s.l. in the slopes east of the gorge. The texture of the lava is fine grained with small plagioclase and pyroxene (augite) phenochrysts that make up 30% of the rock volume. (See also a description of a thin section in chapter 2.2.14). The rock is dense and comprises irregular flow layering and flow foliations and forms coarse polygonal columns. Cracks and fissures are most often perpendicular to the layering. The acid lava can be traced for 3 km upstream along Ábæjará where it disappears because of the local tectonic tilt (10° SE). At Jökulsá it also can be traced 3 km upstream where it disappears below the bottom of the valley. The Brennigilshólar landslide cover the western part of the lava dome but it terminates most probably below the western part of the slide. East of the landslide the rhyolite lava doesn't appear but the rhyolite tephra is found along with a sandstone layer, (áb4). In the south slopes of Austurdalur valley the dome has not been found either but tephra and sandstone instead. The surface breccia of the dome or the contact between its top and the country rock is nowhere exposed. The acid dome is cut by numerous basaltic dykes related to the overlaying basaltic lava pile.

Polarity measurements in the field give very weak polarity but it seems to be normal (N). Field measurements on the magnetic polarity of acid lavas often give weak and confusing results.

The stratigraphy in Ábær gorge indicates nearby eruptive vents. The activity has started with explosive plinian phase and accumulation of a thick tephra layer. Along with the plinian phase an acid lava was extruded. It was very viscous and piled up around the conduit and flowed slowly out over the light coloured tephra carpet forming a layer of basal breccia and a massive rhyolit dome (or a ridge). The dimensions of the lava dome are not clear but it can be traced 5 km from east to west and 3 km from north to south. Its thickness reaches at least 250 m. The base of the dome might either be semi-circular or elongated, and that is more likely, with the long axis trending north south. If it is however assumed to be circular, 5 km in diameter and with average thickness around 100 m its volume would be 0.5 km^3.

A separate acid or intermediate lava is found close to the rhyolite dome at the base of the Göngufjall mountain in Ábæjardalur indicating prolonged acid volcanism after the eruption of the dome.

Reference profiles for the Ábær dome is in the Ábær gorge.

2.2.3   *Ábær Tholeiite Formation (áb5)*

The Ábær tholeiite formation (áb5) is a thick pile of thin-layered tholeiite lavas that bank up against, and cover, the Ábær dome. The contact has nowhere been found yet. In Merkigilsfjall mountain the formation is 250-300 m thick and in Vesturdalur valley it is 200 m. In Tinná valley are one or two olivine basalt layers at the top of the formation. (They are not shown on the map). All interbeds are thin and seem to witness very active volcanism and high production rate.

Reference profile for the Ábær tholeiite formation is in the ravine of Brennigil at the northern scarp of the Brennigilshólar landslide. The base of the profile is seen on top of the Ábær sediment (áb4), in 480 m a.s.l. Its top is at the base of the light coloured acid tephra of Tinná Central Volcano at 760 m a.s.l.

2.2.4   *Tinná lignite sediment (ti1)*

The Tinná lignite sediment (ti1) is fine-grained, light coloured sand- and siltstone, rather soft and with layers rich in acid volcanic tephra. It is wide spread in the area and can be traced across the mapped area from west to east. Its maximum thickness is over 30 m in Goðdaladalur canyon. The sediment might indicate a hiatus in the pile. Table 3 gives different thickness of the layer at various localities.

Table 3: Thickness of the Tinná lignite sediment

Site	Thickness
	m
Merkidalur, north	3
Göngufjall	3
West of Illagil	16
Illagil, Tinnárdalur	27
East of Illagil	20
Geldingaskarð	9
Sandafjall	18
Goðdaladalur canyon	> 30
Brennigil	11

Thin seams of lignite have been found at several locations in Skagafjörður valleys, as already mentioned in the chapter of Previous work. Excellent descriptions of the lignite sites are found in the reconnaissance articles of Magnússon (1980, 1981). Through the ages people tried to make use of the lignite as fuel but the quantity is limited and the quality is poor so mining never was economical. The best known localities are in Goðdaladalur canyon, Klakkar and Giljadalur in Vesturdalur, Sandafjall in Austurdalur and Illagil in Tinnárdalur. The mapping work has revealed that in most of these places the lignite seams belong to the Tinná lignite sediments where they form thin interbeds. Líndal (1964)had already in 1938 proposed the lignite to belong to a single sedimentary horizon. The thickest lignite layers are at 300 m a.s.l. in Goðdaladalur canyon. Five seams can be seen, made of black, compressed wooden trunks (up to 40 cm across) and branches, sitting in a 4.4 m thick light coloured siltstone. Leaf imprints have been found here and there but no investigation has been made on the flora or a possible fauna in the layers. The Tinná lignite sediments indicate a time of quiescence before increased volcanic activity in the following ages, as will be mentioned in the text a little later.

2.2.5   Tinná olivine basalt (ti2)

The Tinná olivine basalt is the most voluminous basalt lava in the whole area. Its thickness varies from place to place between 20 to 60 m but in some localities it is absent. In most places the layer is formed of one thick lava unit unlike the more common compound olivine basalt lavas. Inside the map its aerial extent is around 25 km x 10 km = 250 km². If the mean thickness is 25 m its volume is 6 km³. In fact not so big compared to the great Holocene lavas but still the most voluminous reported Neogene basalt lava of Iceland. It is noteworthy how small the Icelandic lavas are compared to the huge lavas in the basaltic regions of Greenland. The Milne Land Formation in Scoresby Sund contains the largest flows, the Hjörnedal Marker Lava, 11400 km² in area and 285 km³ in volume and the Lower Ti-tholeiite Horizon that exceeds 300 km³ in volume (Larsen et al. 1989). These lavas are continental flood basalts formed faraway from the mid-ocean ridge and under different geological circumstances than are prevailing in Iceland today.

The deposition of the widespread Tinná lignite sediment was not stopped by the eruption of the Tinná olivine basalt. In some places, as in Giljadalur valley, the accumulation carried on along with formation of more lignite. There these layers are 28 m thick. In places that were not covered by the Tinná olivine basalt the Tinná lignite sediment forms the base for the Skati rhyolite.

2.2.6   Skati Rhyolite Formation (the Skati dome ti3 and the Skati tephra layer ti4)

Skati rhyolite is made of two members, Skati acid lava dome and Skati tephra. Hjartarson et al. 1997 called the formation Tinná rhyolite, but because many rhyolite formations belong to the Tinná Central Volcano the name has been changed. The Skati rhyolite is mostly a thick and extensive monogenetic eruptive formation that has created a mountain, the Skati Dome, rising at least 500 m above its base with the highest peaks in the present Skatastaðafjall area (fig.14).
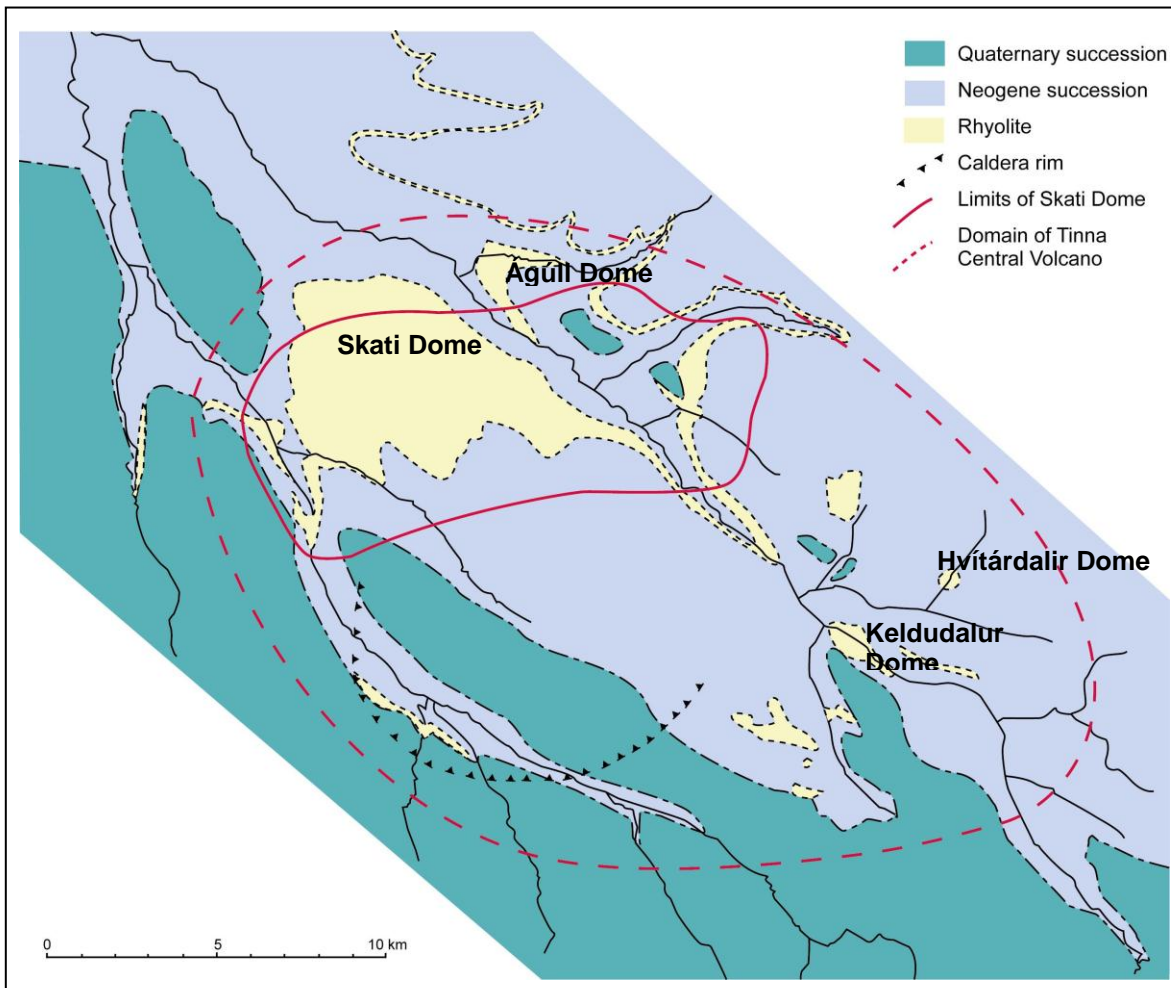
Fig 14. A simple geological map of the Tinná Central Volcano. The domain of the volcano includes all the acid and intermediate lavas related to it. The approximate limits of the Skati dome are indicated and the proposed alignment of the caldera.

The tephra layer is extremely thick or up to 100 m in the closest neighbourhood of the eruptive vent. Along with the Tinná lignite sediment (ti1) they form a thick and easily recognizable marker horizon that can be traced for long distances in the mountains east of Austurdalur valley.

The distribution and thickness of the layers ti3 and ti4 have been measured in 15 locations in the mountain slopes of Skagafjörður Valleys (table 4). On its base the isopatch lines for the tephra can be drawn as shown in fig. 16. It must however be kept in mind that a map based on this material is not as reliable as maps of Holocene tephra sectors that are based on tens or hundredths of soil sections, like the Icelandic Hekla layers (Larsen and Thorarinsson1977).

The tephra layer in the figure is extremely voluminous and indicates greater eruption and more explosivity than known in the Holocene volcanism in Iceland. Welded tuff or ignimbrite has only been found in a close neighbourhood of the eruptive centre. The direction of the main axis of the layer is towards ESE, indicating WNW wind during the main phase of the eruption.

Table 4: Thickness of Skati rhyolite

Place	Top masl	Bottom masl	Thick-ness m	Facies
Merkidalur, north	922	913	9	tephra
Merkidalur, south	890	878	12	tephra
Brennigil	785	755	20	tephra
Göngufjall	712	654	58	tephra
Tinnárdalur mouth	718	546	172	tephra/lava: 94/78
West of Illagil	753	708	41	tephra
Illagil	730	679	51	tephra
Tinnárdalur head	753	703	50	tephra
Tinnárdalur SE side	735	550	185	tephra/lava: 82/103
Geldingaskarð	660	545	115	tephra/lava: 60/55
Sandafjall	610	430	180	tephra /lava: 120/60
Skatastaðafjall	870	400	470	tephra /lava: 100/370
Skriðugil, at Hof	540	496	44	tephra
Vesturdalur	260	380	120	tephra /lava: 10/110
Goðdaladalur canyon			>20	tephra

Viscous rhyolite lava has been extruded along with the tephra outpouring. It covered the tephra from the earliest phase of the eruption but on top of it the ash fall from the later phases is seen. The site of the main crater is not known. The conduit is buried below the main formation in Skatastaðafjall mountain, but the best assumption to pinpoint it would be at the centre of the mountain. From this site the lava flowed out in all directions out from the crater on top of a thin layer of black basal breccia of obsidian aggregate. This is the black jasper mentioned by the pioneers Ólafsson and Pálsson already in the 18th century. Towards east it can be traced for over 8 km into the central Tinnárdalur valley and towards west it disappears into Hofsfjall mountain at 7 km distance from the assumed crater. The extension towards south and north is not known but it can be traced for 6 km (fig. 14). The maximal thickness is found to be 460 m in Skatastaðafjall mountain but it was higher in the beginning, the highest peaks have been worn away.

According to geological mapping, the area of the dome is over 80 km^2 (fig. 14) and if the mean thickness is 100 m then the volume is 8 km^3. Using the thickness of the tephra in table 4 and the distant occurrences of the layer in ODP-holes discussed later, the tephra might correspond to 10 km^3 of dense rock making the total volume equivalent to 18 km^3 of acid magma (table 5). This is the largest monogenetic rhyolite formation still reported in Iceland. For comparison the Þórsmörk ignimbrite has been estimated to be around 8 km^3 in original volume (Jörgensen 1980). It forms the Ash Zone 2 (Z2) in the North

Atlantic and was erupted 54.500 years ago in a major eruption in Tindfjallajökull, South Central Iceland.
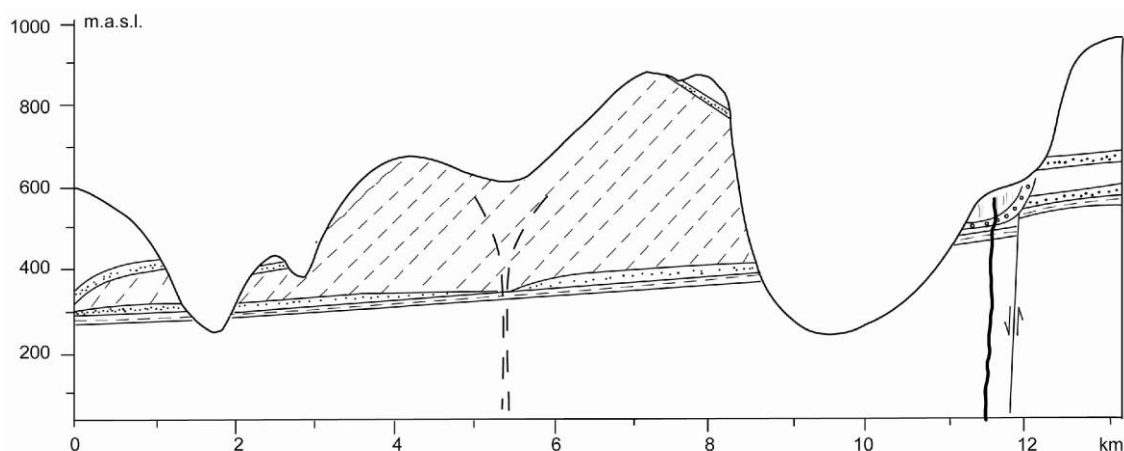
Fig. 15. The Skati rhyolit dome. White = Tholeiite lava piles. Broken line = Sedimentary layers. Dotted = Skati rhyolite tephra. Hatched = Skati dome (rhyolite lava). Austurdalur Pleistocene Volcano and its feeder dyke is shown to the right. Estimated location of the feeder dyke of the Skati dome is also indicated.

Table 5. Volumes of the Skati lava and tephra

	Volume km^3	Dense rock eqv. km^3
Lava	8	8
Tephra > 10m thick	20	5
Additional tephra on land	10	2.5
Tephra off shore	10	2.5
Total volume	47	18

It seems reasonable to expect the formation of a huge collapse caldera after such an immense eruption but no signs of it have been found. Irregularities in the dip and strike and a local anticline in the mouth of Tinnárdalur might indicate some crustal deformation and depression below the main body of the rhyolite formation but no caldera faults have been observed.

Chemical analyses show high silica content in the Skati rhyolite or around 75% on a water-free basis(Table 8).
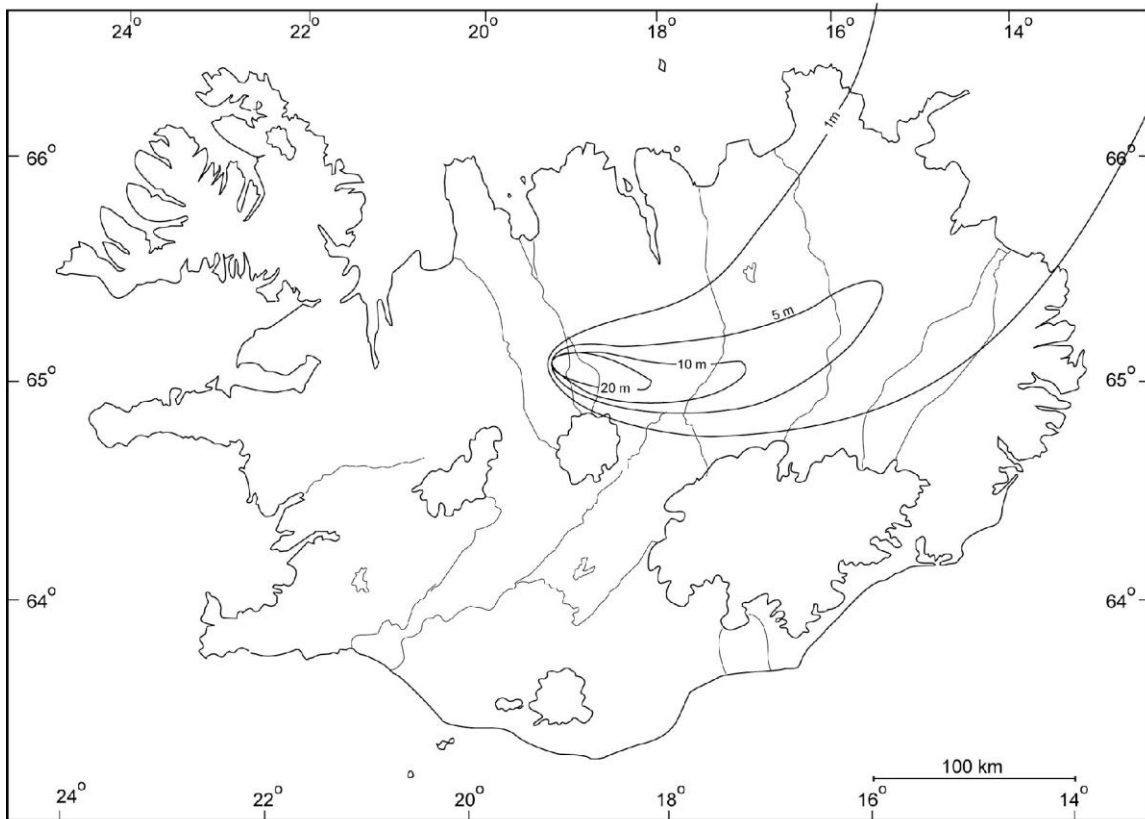
Fig. 16. The tephra sector close to the eruptive site. The innermost ellipses are the measured and estimated isolines for the 10 and 20 m tephra thicknesses. The outer isolines are hypothetical. The tephra sector is projected on the country as it looks like today. Its size and shape 5.5 Ma is unknown. The regional distribution is illustrated on fig. 17.

2.2.7   Correlations

It should be possible to find the Skati tephra interbedded in the lava pile elsewhere in Iceland. As discussed later it is around 5.5 Ma and should be found within the reverse polarity subzone C3r, the lowest subzone of Gilbert. The stratigraphy of this subzone has been observed and described both in East and West Iceland but in the literature no layer is mentioned that seems reasonable to correlate with the Skati tephra.

Sæmundsson and Noll (1974) and McDougall et al. (1977) made a stratigraphic and paleomagnetic survey including K/Ar dates on a composite lava section, including Gilbert polarity zone, in Borgarfjörður 125 km southwest of the Tinná Volcano. The localities are upwind from the Tinná Volcano. No tephra or sedimentary layers are shown in those profiles. The same is for investigation in Bjarnarhafnarfjall and Eyrarfjall West Iceland where lower Gilbert is exposed (Kristjánsson and Jóhannesson 1999). No rhyolite tephra layers appear in the profiles.

In East Iceland, chron C3r has been proposed in profile R at Melgræfur in Fljótsdalur (Watkins and Walker 1977). No tephra layers appear there.
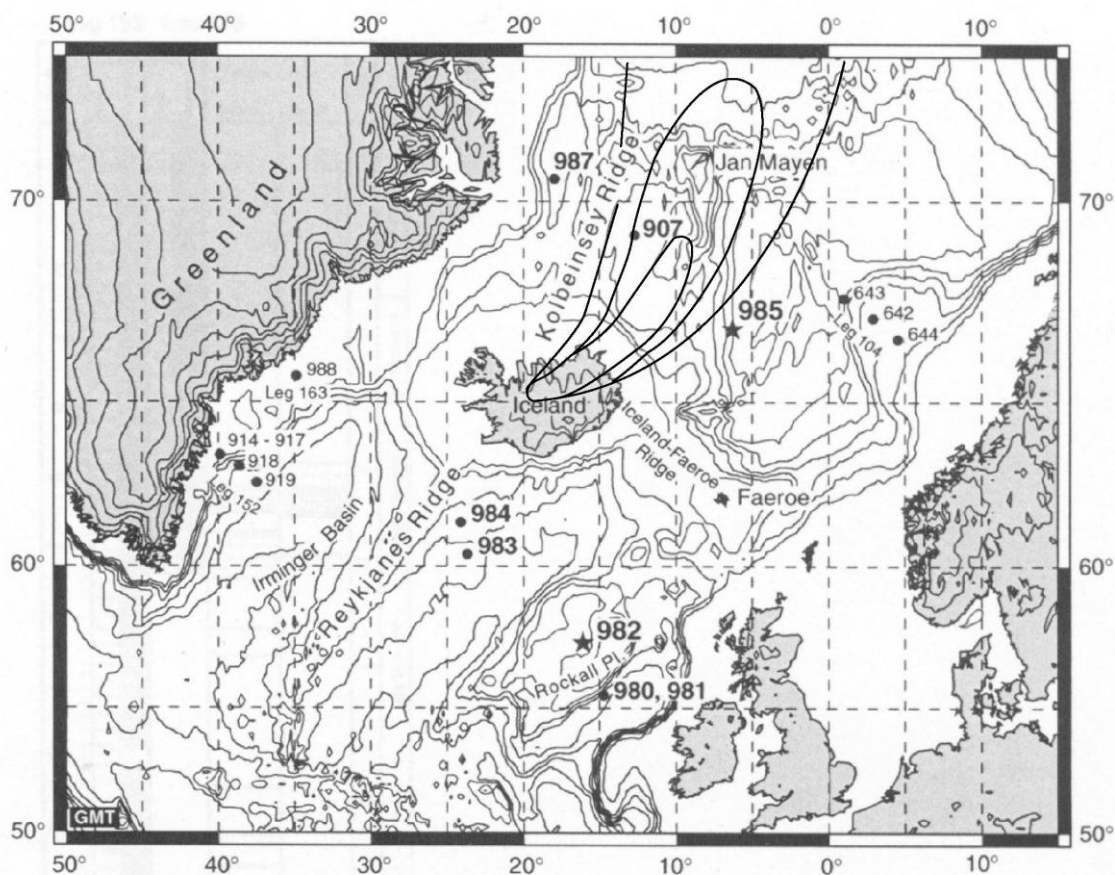
Fig. 17. Map of the central North Atlantic showing locations of ODP-sites and suggested distribution of the Tinná tephra. The alignment of the tephra sector is in fact unknown but here the most probable alignment is shown.

It might be possible to recognize the Skati tephra in ODP cores from the ocean floor around Iceland, especially to the east and northeast of the country. The major semi-axis of the tephra sector has an ESE direction close to the eruptive sites. This indicates the wind direction near the surface. Higher up in the troposphere the situation might have been different, as there the predominant wind direction is from southwest and is supposed to have remained unchanged during the last 10 Ma (fig. 16 and 17). Lacasse and Garbe-Schönberg (2001) studied the explosive volcanism in Iceland and the Jan Mayen area during the last 6 million years, by tephra layers in the deep-sea sediments. They recreated a composite marine tephra record based on data from ODP sites 907, 985, 919, 983, 984 and from the sites SU9029 and SU9032. They found out that 90% of all identified tephra layers were recognized in the ODP sites 907 and 985, downwind from Iceland. The lowest part of their record is entirely based on these two sites. It should be 90% probability to find the Skati tephra there.

Table 6. ODP-drill holes in the Iceland area penetrating late Miocene sediments

Coring site	ODP-leg	Ocean basin	Latitude (N°)	Longitude (W°)	Water depth (m)	Length (m)	Remarks
642B	104-	Vöring Plateau	67°13.5'	- 2°55,7'	1286	221,1	Good core
642C	-	Vöring Plateau	67°13.2'	- 2°55,8'	1277	221.1	Good core
643A	-	Vöring Plateau	67°42.9'	- 1°02.0'	2753	565.2	Hiatus 4 – 6.5 Ma
907A	151	Iceland Plateu	69°14.989'	12°41.894'	1800.8	216.3	Hiatus 7 – 10 Ma
907B	-	Iceland Plateu	69°14.989'	12°41.898'	1801.6	211.7	Hiatus 7 – 10 Ma
907C	-	Iceland Plateu	69°14.998	12°41.900	1801.2	215.1	Hiatus 7 – 10 Ma
914 – 918	-	Irminger Basin					Poor core recovery in late Miocene
982A	162	Hatton – Rockall Basin	57°30.992	15°52.001	1135.3	248.7	Poor time control
982B	-	Hatton – Rockall Basin	57°31.002	15°51.993	1134.0	614.9	Poor time control
982C	-	Hatton – Rockall Basin	57°31.009	15°51.992	1133.7	250.8	Poor time control
985A	-	Iceland Basin	66°56.490	6°27.012	2787.6	587.9	Poor time control below 7 Ma
985B	-	Iceland Basin	66°56.498	6°27.001	2787.8	126.9	Poor time control below 7 Ma
987E	-	Greenland Plateau	70°29.787	17°56.188	1672.7	859.4	Poor time control below 5 Ma

Table 7. Lowest Gilbert tephra horizons from ODP-site 907 (based on Lacasse et al. 1996 and Lacasse and Garbe-Schönberg 2001).

No.	Depth m	Core section	Interval cm	SiO_2 %**	Thickness cm	Age Ma***	Remarks
1	81.93	9H-6	63-68	73.53	5.0	5.183	From Icelandic rift zone (Z)
2	82.30	9H-6	100-104	71.36	4.0	5.215	From Icel. off-rift zone (AA)
3	82.44	9H-6	114-120	76.18	6.0	5.228	From Icelandic rift zone (AB)
4*	84.75	10H-1	84.75				
5*	84.80	10H-2	0-13	75.26	18.0	5.480	From Icelandic rift zone (AC)

* No. 4 and 5 are the same layer. It's in two parts because it was recovered in two core sections during drilling. ** The silica content is given on a water-free basis. *** The age is based on Lacasse and Garbe-Schönberg 2001 in spite of a discrepancy with Channel et al. 1999

Table 8: Chemical composition of Skati dome and the tephra layers in ODP-907A

Element	ODP-907* AC	Skati dome S-23	Skati dome S-31	Skati dome S-16	Skati dome S-17	Hvítárdalir dome S-11
	(wt %)	(wt %)	(wt %)	(wt %)	(wt %)	(wt %)
SiO_2	71.49	74.60	75.95	73.75	74.65	74.56
TiO_2	0.14	0.142	0.10	0.166	0.122	0.146
Al_2O_3	12.18	13.17	12.87	12.86	13.12	13.27
Fe_2O_3	2.23 **	2.13	1.8	2.21	2.11	2.28
MnO	0.09	0.06	0.05	0.05	0.02	0.03
MgO	0.01	0.03	0	0.05	0.02	0.05
CaO	1.11	1.01	0.94	1,20	0.95	1.07
Na_2O	5.36	4.64	4.69	4.54	4.69	4.71
K_2O	2.38	3.40	3.42	3.30	3.26	3.24
P_2O_5		0.031	0.010	0.018	0.017	0.021
LOI		0.84	0.09	1.9	0.74	0.84
Tot.	94.99	100.15	99.98	100.15	99.8	100.32
	ppm	ppm	ppm	ppm	ppm	ppm
BaO	508	713	712	671	685	675
Ce	162	138	103	149	141	130
Co	0.77	26	22	10	20	21
Cr_2O_3	11.6	< d/l	< d/l	18	< d/l	< d/l
Cu	23.5	19	19	32	11	52
Ni	6.90	< d/l	< d/l	3	< d/l	< d/l
Sc	1.55	< d/l	< d/l	< d/l	< d/l	< d/l
V		< d/l	< d/l	< d/l	< d/l	12
Zn	236	128	131	132	119	121
Ga	28.7	26	25.5	24.9	24.7	25.5
Nb	81.4	73.3	71.8	70.7	71.5	69.8
Pb	9.32	3.2	3.1	3	3.2	3.2
Rb	63.3	81.3	83	80.2	79.3	77.2
Sr	76.3	61.9	54.7	65.5	70.3	78.1
Th	8.15	5.5	4.9	5.9	5.6	5.4
U	2.47	< d/l	< d/l	< d/l	< d/l	< d/l
Y	126	113	132.8	128.5	131.2	123.1
Zr	403	253.2	232.8	268	288.6	324.1

*From Lacasse and Garbe-Schönberg 2001. **Total iron as FeO.
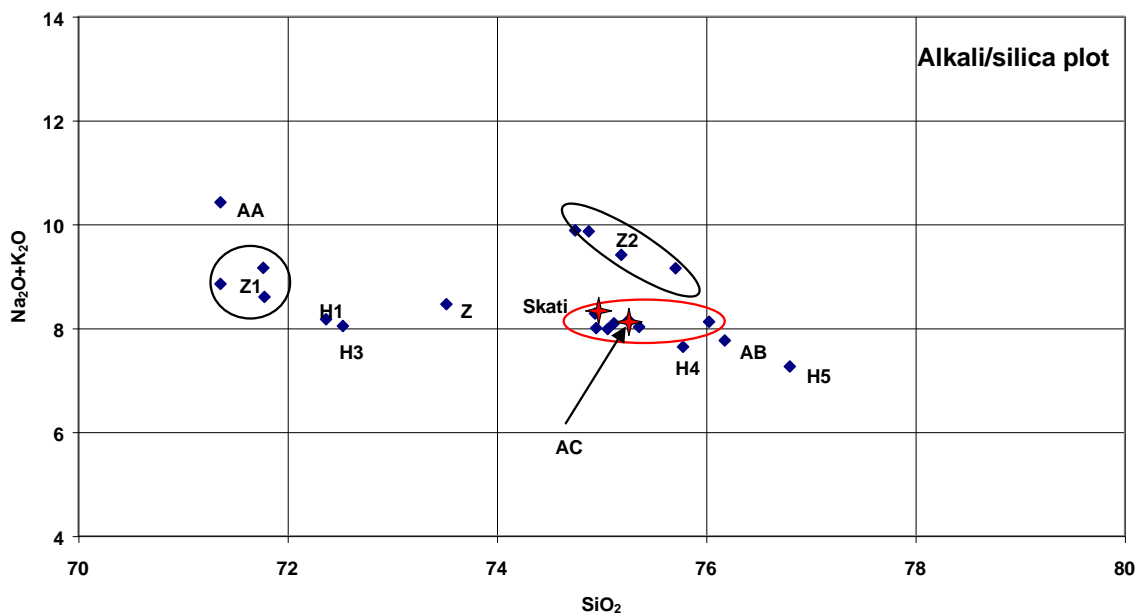
Fig. 18. Alkali/silica plot for Skati Dome and several tephra layers in the deep-sea sediments. The Skati samples are the blue diamonds inside the red ellipse. The AC-layer from ODP-907 is indicated by red stars. It plots close to the Skati samples. The layers AB, AA and Z are also found in ODP-907 next to the AC-layer. Z1 and Z2 are well known tephra layers in the deep sea sediments in the North Atlantic. H1, H3, H4 and H5 are Hekla tephra samples from deep-sea cores. (The deep-sea data is from Lacasse et al. 1996a and b).

The main attributes of this ash layer should be the following:

> Discrete layer
>
> Age 5.3-5.9 Ma (Gilbert polarity subchron C3r)
>
> Colourless glass shards
>
> Silica content around 75% SiO_2 (on a water-free bases)
>
> Polarity R. (Located in the central or lower part of a reverse polarity subzone)

Four tephra layers from ODP-core 907A, fulfil these requirements. They are listed in table 7 (see also fig. 20).

Chemical analysis indicates that ash layer no. 2 in table 7 originates in an off-rift volcanic zone in Iceland and can therefore be eliminated. The others originate in a rift zone (Lacasse and Garbe-Schönberg 2001). Layers 1 and 3 are very close to the upper boundary of the C3r chronozone which makes them unlikely as the missing one. (Lacasse and Garbe-Schönberg (2001) even place them inside Þverá subchron but that seems to be in discrepancy with the paleomagnetic stratigraphy of Channel (1999) and the locations of ash layers as reported by the Shipboard Scientific Party (1995)). Altogether, layer no. 4-5 would be the best candidate for the Skati tephra and if it is hidden in the data

collection at all, it is actually the one. The layer is the oldest tephra in the composite core of Lacasse and Garbe-Schönberg (2001) and they point it out and say: "Dispersal, thickness and grain size of the tephra layer clearly indicate that it was likely derived from one of the largest Icelandic explosive eruptions that ever occurred in the late Tertiary".
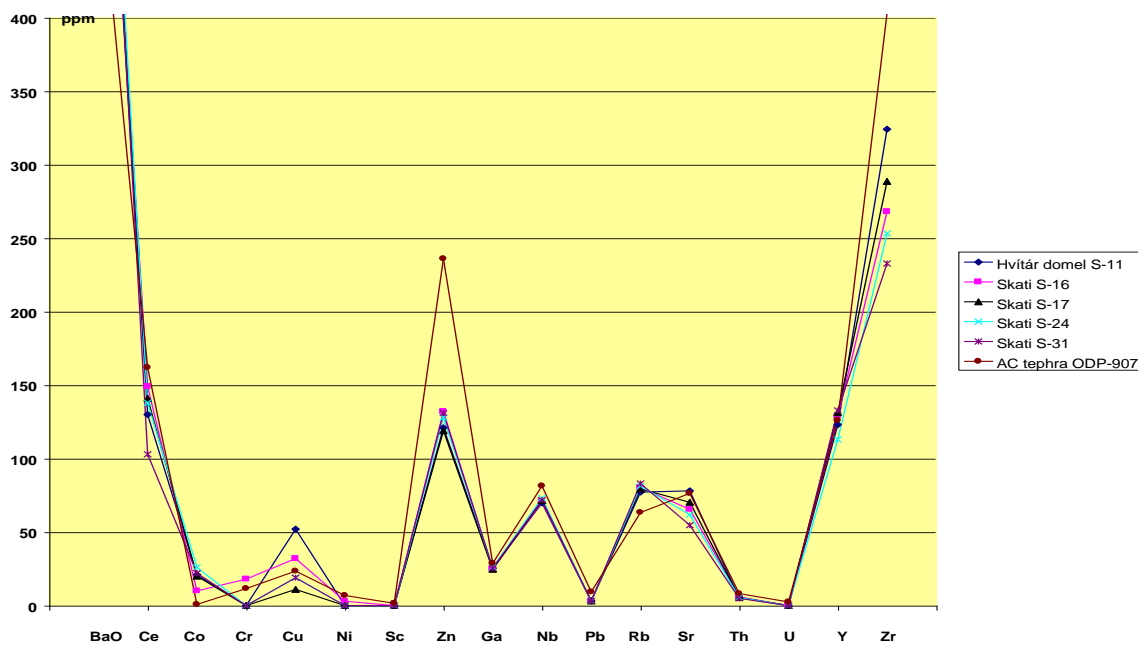
Fig. 19. Comparision on trace elements in Skati rhyolite dome and the AC-tephra in ODP site 907.

Comparing the chemistry it must be kept in mind that the ODP-sample is a glass fragment from a tephra layer that is analysed using a microprobe but the Skagafjörður samples are of lava that were analysed in a conventional way an a commercial laboratory. The comparison shows however similar composition in most of the major elements as well as in the trace elements and does not indicate any difference that is not possible to explain by the different analytical methods and sedimentary environment. If calculated on a water-free bases the average silica content in the four samples from Skati Dome is 75.39% versus 75.26% in the ODP-sample. The Na_2O+K_2O/SiO_2 plot also gives just about the same values for all of the samples (fig. 18 and 19).

In the Scientific Results of the Ocean Drilling Program, Vol. 151, the ash layers from site 907 are described (Lacasse et al. 1996a). There this layer has got the label AC. It is bimodal and the top differs from the bottom. The bottom is crystal-poor tephra of 100% colourless glass shards and can be considered as exclusively silicic. The top contains 9% feldspar, 3% clinopyroxene and up to 2% olivine. This sorting might have taken place as the ash particles settled through the 1800 m deep water column. Normal grading is also observed between the bottom and top of the layer. It can also be interpreted as the result of extensive size fractionation in the water column.

Microscopic investigation on thin-sections from Skati lava and tephra has not revealed phenochrysts or other easily recognized characteristics for identifying the tephra layer in distant sites. Description on the thin-sections is given in chapter 2.2.14.
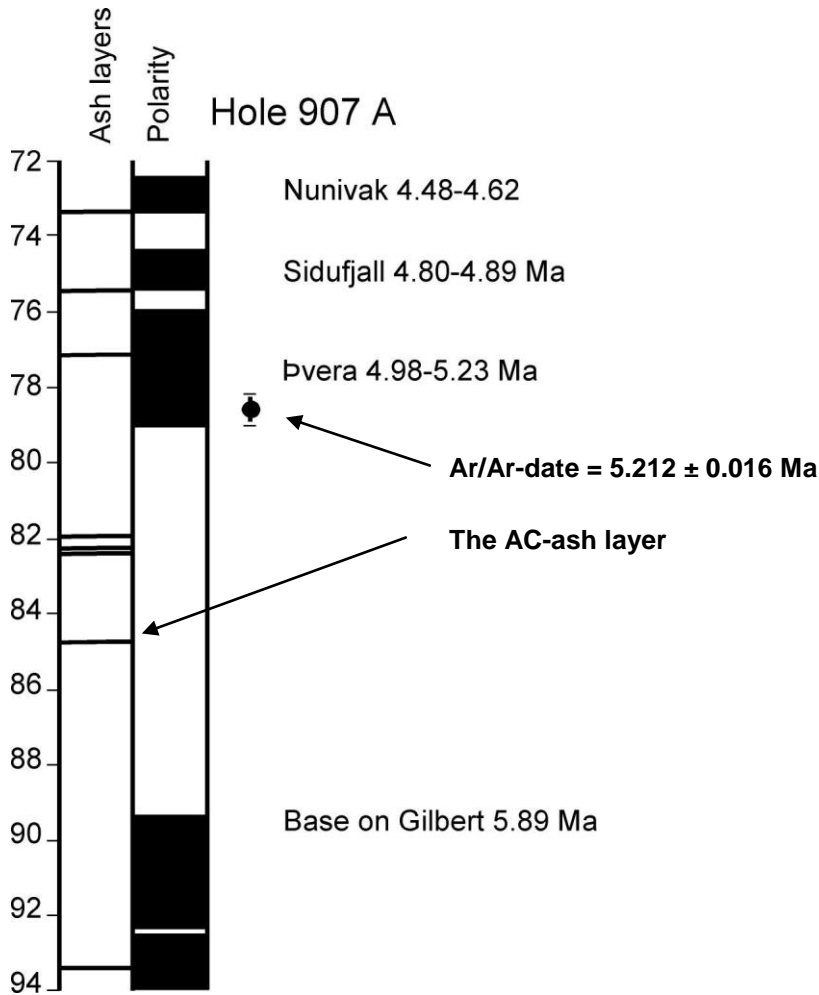
Fig 20. Core segment from drillhole A at ODP-site 907 displaying ash layers, polarity and ages. The AC-ash layer, the best candidate for the Skati tephra is indicated. The Ar/Ar determination on Skati Dome is also shown.

ODP-site 907 is in the Artic Ocean NNE off Iceland, 550 km away from the Tinná Volcano. The holes (A, B and C) are at 1800 m below sea level and the tephra layer is close to 85 m depth in the sediments (table 7).

At ODP-site 985 NE of Iceland several layers of lower Gilbert age have been identified and the Skati tephra could hide among them. All are rather thin (1-4 cm) and they have not been correlated with the layers at site 907 (Wallrabe-Adams and Werner 1999). Chemical analyses are lacking. At ODP site 642 on Vöring Plateau the cores from holes B and C contain sediments from 5 – 6 Ma. No promising ash layer is found there.
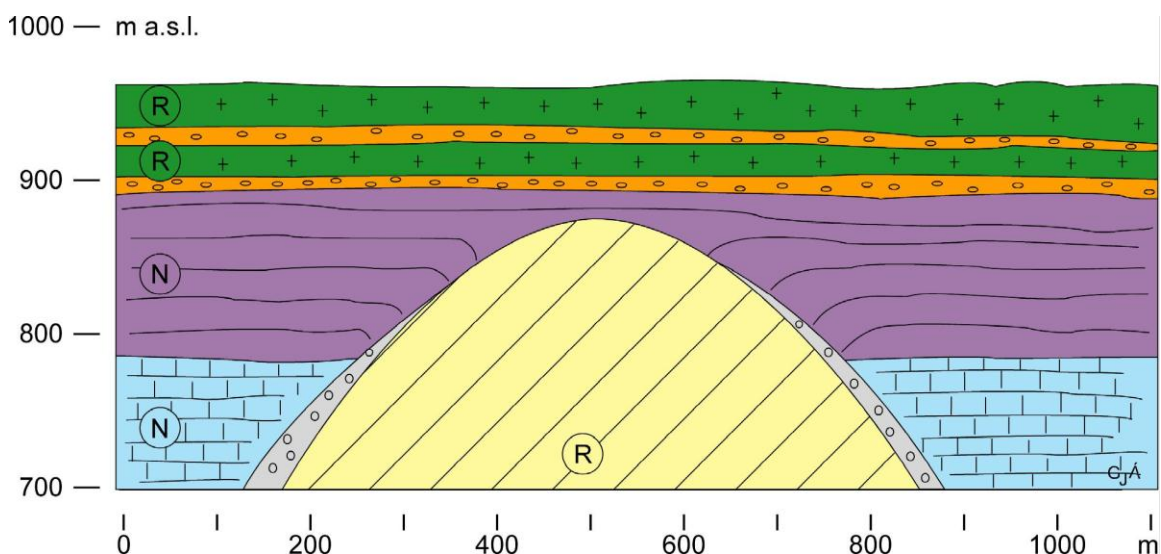
Fig. 21. The Hvítárdalir dome with fossil screes at its sides (grey) buried in younger lavas. Green = Fossá porphyritic basalt. Orange = Fossá sediments. Yellow = Nýibær andesites. Blue = Nýibær tholeiites. N and R show the polarity (field measurements).

2.2.8 Hvítárdalir dome

In one of the narrow tributary valleys of Hvítár there is a rhyolite dome, or a ridge, buried in the lava pile. Its cross-section can be seen in the steep slopes at both sides of the valley (fig. 21). It is at least 100 m thick but the base of the dome is below the bottom of the valley and can't be seen. The diameter of the dome along the river Hvítá is 750 m. Chemical analysis shows 75% SiO_2 (table 8) and close petrological relations with the Skati rhyolite. A prominent sedimentary layer, fossil scree, is situated between the dome and the lava pile banking up against it. The dome has reverse polarity and might be of similar age as the Skati Rhyolite and a member of that formation. (See fig. 14 for location).

2.2.9 Tinná Andesite Formation (ti5)

After the great eruption of the Skati rhyolite a new eruptive phase began with production of intermediate lavas. Several andesite lavas, 20 - 70 m thick, have been mapped along with a few acid lavas that pile up against the southern slopes of the Skati dome. They can be seen in Austurdalur near Hvítá tributary and in Vesturdalur near Giljar farm. Their origins are somewhere south and southeast of the dome. The centre of the volcanic activity seems to have migrated in that direction immediately after the Skati eruption.

2.2.10 Tinná Tholeiite Formation (ti6)

The Tinná tholeiite is a pile of thin lavas with thin interbeds covering the andesite lavas banking up against the rhyolite dome and encircles it. The thickness varies from the

maximal 160 m to zero. The upper limit of this formation is marked by a shift of the magnetic field from reverse (R) to normal (N). The production rate of the tholeiites seems to have been high because of its thin interbeds and at the magnetic reversal field measurements show how the strength of the field declines and increases again after the shift recording a polarity transition zone.

Reference profiles: For the lower part of the group, i.e. ti1-ti4, the reference profile is located in outer Tinnárdalur, between 520 –775 m a.s.l. For the upper part of the group, ti5-ti6, the reference profile is in Sandafjall, 1.5 km upstream of Hildarsel hut between 520-775 m a.s.l.
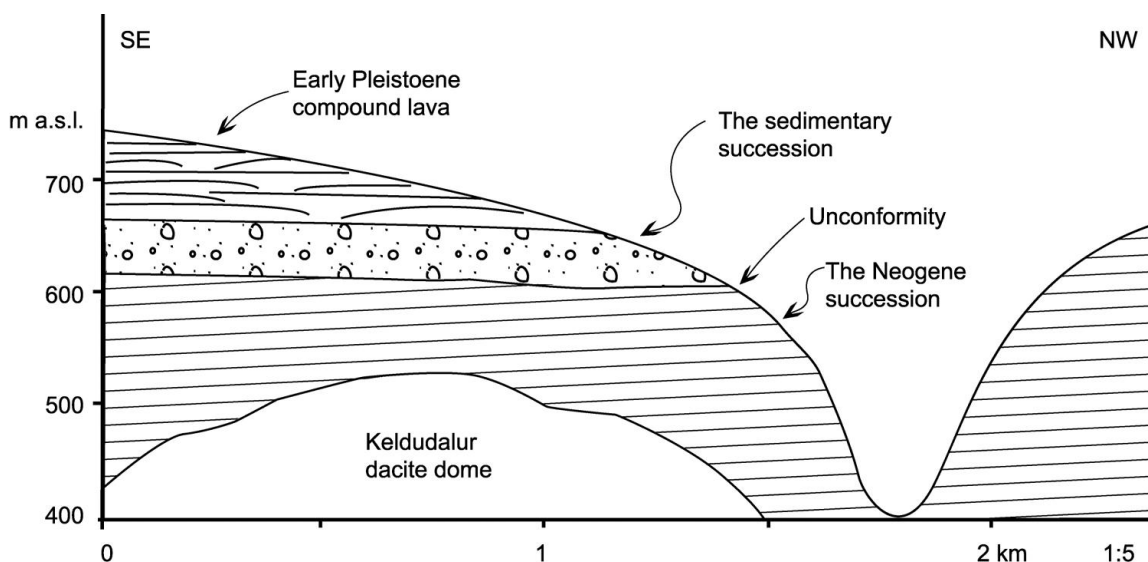
Fig. 22. Keldudalur dacite dome. A cross-section seen from south. The dome is buried below a Neogene lava succession (Nýjibær formation, nb1). The lavas are dipping 6° S. Above is the unconformity and its associated sedimentary layer (conglomerate). On top is olivine tholeiite compound lava of early Pleistocene age (Giljamuli formation, ep3).

2.2.11 Keldudalur Dacite Formation

Keldudalur is a tributary valley in inner Austurdalur. Near its mouth and in the valley itself acid rocks are found in several places all of which are thought to be exposures of a single formation. The rock type is dacite (68% SiO_2). It forms a ridge or a row of two or more volcanic domes trending north-south. Keldudalur dome is partly covered by loose screes and tillite so its base and contacts with the adjacent rocks have not been investigated. Its highest part reaches 600 m a.s.l. (fig. 22). The visible size is 3 x 2 km and its maximal thickness is at least 200 m. Assuming elliptical shape and 100 m mean thickness the volume of the ridge is 0.5 km^3.

The polarity is very uncertain. The field measurements show anomalous or a weak normal direction. The location of this unit inside the geological column is not clear and the Keldudalur dacite formation might even be an intrusion. Further investigations are needed to judge on that. (See fig. 14 for location).

2.2.12 Nýibær Olivine Basalt (nb0) and Nýibær Tholeiite Formation (nb1)

The Nýibær Olivine Basalt (nb0) is found at the northern limits of the map, on top of Merkigilsfjall mountain. It is made of thick lavas of olivine basalts, sparsely porphyritic with greyish colour. These layers seem to be regional basalt floods from a remote volcanic system, or systems, and therefore, in fact, do not belong to the rocks of the central volcano. They have limited distribution inside the map and farther south and east they are replaced by the Nýibær tholeiite (nb1).

Nýibær Tholeiite formation (nb1) is very similar to the Tinná tholeiite (ti6) in all respects except polarity. From petrological point of view they would be regarded as the same formation. It carried on the work of the Tinná tholeiites and immersed the Skati dome totally. Remnants of the Nýjibær tholeiite lavas can be still seen near the summit of the former acid dome in Skatastaðafjall mountain. The polarity is normal (N) and according to paleomagnetic time scales and the interpretation in table 23 this is the Þverá subchron that started close to the Miocene/Pliocene boundary 5.2 Ma (see fig. 20).

2.2.13 Nýibær Andesite Formation (nb2) and the Nýibær Rhyolite Formation (nb3)

At the end of the Nýibær period an intermediate and acidic volcanism started up again in the area forming the Nýibær andesite formation (nb2) and the Nýibær rhyolite formation (nb3). In Vesturdalur valley there is a hill of andesitic hyaloclastite and fragmental basalt lavas with remnants of pseudo craters. These formations might indicate an eruption inside a lake and high groundwater level and wetlands at this place in the time of the Nýibær volcanism. The acidic rock is mostly in the form of lavas, but tephra layers have not been reported. Chemical analysis show that at least some of the lavas are in fact dacite with 65% SiO_2. Tholeiite lavas are also found in this pile. It reaches 180 m thickness. Several cone sheets (cone dykes) indicating caldera formation, have been mapped near Fossá river in Vesturdalur but no distinct caldera fault has been found. An assumed caldera rim is shown on the map. Its location is mainly based on cone dykes near Fossá.

The upper limit of the Nýibær formation is in many places marked by the Skagafjörður unconformity but in other places it disappears underneath older formations.

The Nýibær acid and intermediate lavas are the youngest known formations of the Tinná Central but it cannot be stated that they mark the extinction of the volcano. Still younger formations might have been wiped away by the erosion as seems to be indicated by dykes and sheets cutting the topmost layers of the Nýibær formation.

The centre of the activity in the Tinná Central Volcano seems to have migrated to the southwest by the time. It was ignited near Ábær river in Austurdalur, then it was moved to Skatastaðafjall between Austur- and Vesturdalur valleys, but ends near Hraunþúfuklaustur in Vesturdalur, 15 km southwest of Ábær river.

Reference profile for the lower part of the formation (nb1) is in Hjálmarselslækur in Austurdalur but for the upper part (nb2 and nb3) it is in Afréttarfjall in Austurdalur.

2.2.14 Thin sections of rock samples from the Tinná Volcano

Four thin sections were made from rock samples collected in some Tinná acid formations. The primary aim of this investigation was to find out the crystallization, texture and main characteristics of the Skati lava and tephra in found in Skagafjörður Valleys in order to correlate the Skati eruption with tephra layers recovered from deep-sea cores of the ODP. Thin sections were made from the tephra, from the obsidian sole of the lava and from the dense interior of the lava itself. A sample from the Ágúll Dome also was observed. The microscopic investigation did not revealed phenochrysts or other easily recognized characteristics for identifying the tephra layer in distant sites.

AH-1 Skati Dome, obsidian

The obsidian sample is taken from the base of the Skati Rhyolite lava at 660 m a.s.l. in Tinnárdalur mouth. The obsidian layer is 10 m thick but grades into more crystalline rock higher up. The lava is altogether 80 m thick, resting on the Skati tephra. Light coloured phenocrysts are seen in the black surface of the hand specimen.

In the microscope the glass is most prominent as colourless groundmass, black in polarized light.

Microcrystalline plagioclase is abubdant. Most of thcrystals are elongated and some are needle like. The orientation of the crystals is irregular. No lamination detected.

Twins are rare or absent.

The sides of the plag. crystals are rather diffuse (subhedral).

Occasional (clino)pyroxene crystals seen with diffuse subhedral sides.

Iron-titanium oxides, titanomagnetit or ilmenit, are rare but not absent.

No olivine seen.

No vesicles.

The phenocrysts seen in the hand specimen do not show up under the microscope.

AH – 2 Skati rhyolite tephra

The sample is collected in the upper part of the 75 m thick Skati tephra layer below the Skati lava in Tinnárdalur mouth at 630 m a.s.l. This is in the same section as sample AH-1. The rock type is pyrioclastic rock composed of primary Skati rhyolote and lithic fragments of various rocks from the conduit. No plutonic fragments seen.

The thin section is of rather poor quality.

The rock seems composed of heterogeneous fragments, a primary acidic pumice made of nearly pure glass. Bubbles are abundant.

In between there are fragments of secondary crystalline rocks. In the microscope fragments with small crystals of plagioclase are the most prominent. The crystals are elongated. Their sides are irregular and broken but not diffuse. Twins abundant. Irregular (clino)pyroxene crystals are in between the plagioclase crystals. Suspected small orthopyroxene crystals, gray in polarized light. Iron-titanium oxides, black grains, irregular in shape. The rock seems fresh and unaltered and may be of basic lavas.

More altered fragments are also seen, perhaps from sedimentary layers.

The secondary rocks are most likely from layers below the volcano but brought up by the explosive eruption.

AH – 3 Skati Dome laminated rhyolite lava

The sample is from Skati lava at one of the innermost rhyolite exposures in Vesturdalur valley, 300 m a.s.l. The rock is fine grained and densely laminated. Dark gray bands are in between light gray bands with pink hue.

Under the microscope the lamination appears in highly orientated, small, elongated plagioclase microcrystals. The crystals seem often about 5 times longer than their width. Their outlines are subhedral and diffuse.

No phenocrysts.

Iron-titanium oxides, small black crystals abundant, titanomagnetite or ilmenite, euhedral and subhedral.

Groundmass mostly glass, light brown in colour but dark brown or black in polarized light.

Small crystals of tridymit cristobalite seen (high temperature polymorphs of quarts)

Pyroxene and olivine not seen.

Small vesicles line up in between the laminas here and there.

Unidentified rock fragments with vesicles seen.

AH – 4 Ágúll Dome, acid lava

The sample is from the Ágúll lava dome, the lowest acid formation of the Tinná Central Volcano. The sample is collected in Ábær gorge near the base of the dome. The rock is coarse grained, having dark brown colour. Small phenocrysts of plagioclase and pyroxene are prominent making up 20-30% of the rock volume.

Under the microscope large, well-shaped, euhedral phenocrysts of plagioclase appear. Twins abundant, sometimes zoned. The volume is around 20% of the rock mass. Outlines are sharp, no diffusion. Metal inclusions seen.

Clinopyroxene 3 – 5 % by volume, smaller than the plagioclase. Outlines irregular. Beautiful elongated augite crystal seen with metal inclusions, 10 times longer than its width light brown but blue-green in polarized light.

Orthopyroxene with high relief and small vesicles also seen.

Small black metal crystals in between the plagioclase, square to oval, never with sharp corners.

Groundmass dark brown but black in polarized light. No vesicles in the groundmass, small crack filled with brown amorph material seen.

2.3 The History of Tinná Central Volcano

The Tinná Central Volcano is one of ca. 50 known central volcanoes in the Tertiary regions of Iceland. In the Quaternary regions they are around 40 according to the Geological Map of Iceland by Jóhannesson and Sæmundsson (1998b). The Tinná Central Volcano derives its name from Tinná river and Tinná valley which are situated near the centre of the volcano. The name was given by Hjartarson et al. (1998). In Icelandic the word tinna (or hrafntinna) means obsidian and the name Tinná means the Obsidian River, or a river where obsidian rocks are common. Obsidian is the black form of rhyolitic glass that often is found at the top or the bottom of acid lavas and thus indicates a central volcano.

Jóhannesson (1991) mentions two central volcanoes in Skagafjörður valleys, the Ábær Central and the Keldudalur Central. But based on petrological and stratigraphical observations Hjartarson et al. (1998) suggested these two centrals to be parts of the same volcano and in the last edition of the Geological Map of Iceland this opinion has been adapted (Jóhannesson and Sæmundsson 1998b).

The acidic rocks in the Skagafjörður valleys are first shown on Kjartansson's Geological Map of Iceland (1965) but there they cover far too limited area. On later maps these areas have been enlarged but on the most recent ones (Jóhannesson and Sæmundsson 1998a) the volcano is however still too small. There its domain is shown to cover the mountains at Vestur- and Austurdalur valleys. A domain is defined as the area that covers the acid (and intermediate) lavas of a central volcano. According to that the domain of the Tinná Central Volcano should be shown at least twice as big, reaching farther to the southeast, covering 200 km². It might even be five times as big if the rhyolites of Eyjafjörður Valley (the Torfufell Central Volcano) belongs to it as suggested by Hjartarson et al. (1998). If that is correct the size of the Tinná Central Volcano is 40 km x 12.5 km = ~500 km². Then it would be among the greatest ones known in Iceland. It was not a stratovolcano as Snæfellsjökull but more like Hengill or Krafla, an irregular massif of heaps and domes without any major summit crater. At times it rose above the environment but while it was being piled up flood-basalts were issuing from fissures covering the surrounding country. The products of the volcano interdigitate with them. The volcano rests on flood-basalts and it was finally buried, or nearly buried, by such lavas.

Intrusions and dykes connected to the Tinná central are not conspicuous. The reason is that the volcano is so moderately eroded. It is only exposed to the depth of 600 – 800 m beneath the initial surface. Its roots and magma chamber are deep below the valleys. A few cone dykes from the last episode of the volcano have been found near river Fossá in Vesturdalur as already mentioned. No signs of geothermal activity or high temperature areas have been found in the vicinity of the volcano.

The age of the Tinná central Volcano is based on a single Ar/Ar- dating from on the Skati rhyolite dome (table 21). The date is discussed in chapter 9.1 and there the dome is

suggested to be around 5.5 Ma, which means that it was erupted during the lowest Gilbert polarity subchron (C3r). The lifetime of a central volcano is in the order of 0.5 – 1 Ma (Guðmundsson 2000). After that it cools down and is buried by younger lavas while it drifts out of the active volcanic zone. The Tinná volcano stayed active for at least three polarity subchrons (C3An.1n – C3r – C3n.4n), from ca. 6 Ma to 5 Ma. According to plate tectonics, and the history of ridge jumping in Iceland, the volcano was originated in the North Iceland volcanic zone 10 – 15 km west of Askja.

The development of the Tinná Central Volcano can be divided into three main phases, all starting with rhyolitic volcanism (fig. 23).

1. phase: The initial stage started ca. 6 Ma. Then an acidic eruption started with the formation of the Ágúll rhyolite dome in Austurdalur valley. It was at least 250 m high and 0.6 km^3. The acidic eruption was followed by an intense basaltic volcanism producing Ábær tholeiites that finally covered the dome. The stage ended in a long period of quiescence while the Tinná lignite sediment was formed. The lignite seams contain trunks and branches reflecting Tertiary woodlands.

2. phase: The central stage might have started by the eruption of the large Tinná olivine basalt lava. The relationship of the lava and the central volcano is however unclear. Shortly after the lava eruption violent explosive volcanism took place with colossal ash fall and the extrusion of a new and large rhyolite dome, building up at least 500 m high mountain in the area. A thick and wide spread tephra sector was accumulated with the axis of maximal thickness extending to ESE from the crater area, reflecting the wind direction during the explosive initial plinian stage of the eruption. Farther away, and higher up in the atmosphere, more southerly winds might have diverted the tephra sector more towards north. A thick tephra layer in a sediment core from ODP-site 907, 500 km NNE off Iceland, is suggested to be from this eruption. The total volume of ash and lava emitted could have been 18 km^3 of solid rock equivalent. The eruption was followed up by intermediate volcanism and later on by the formation of the thin-layered pile of the Tinná and Nýibær tholeiites.

3. phase: The final stage appears as a mixture of acidic and intermediate volcanism piling up the Nýibær formations nb2 and nb3. The centre of activity was farther south than in the beginning. A collapse caldera was formed during this phase and a caldera lake was developed. Eruptions in the lake produced small andesitic hyaloclastite ridges and hills. This might have happened around 5 Ma. After that the volcano stayed dormant and drifted to the west, away from the active North Iceland volcanic zone while younger lavas piled up against it and covered it partly, but its highest summits always seem to have extended above the environment, as they still do.
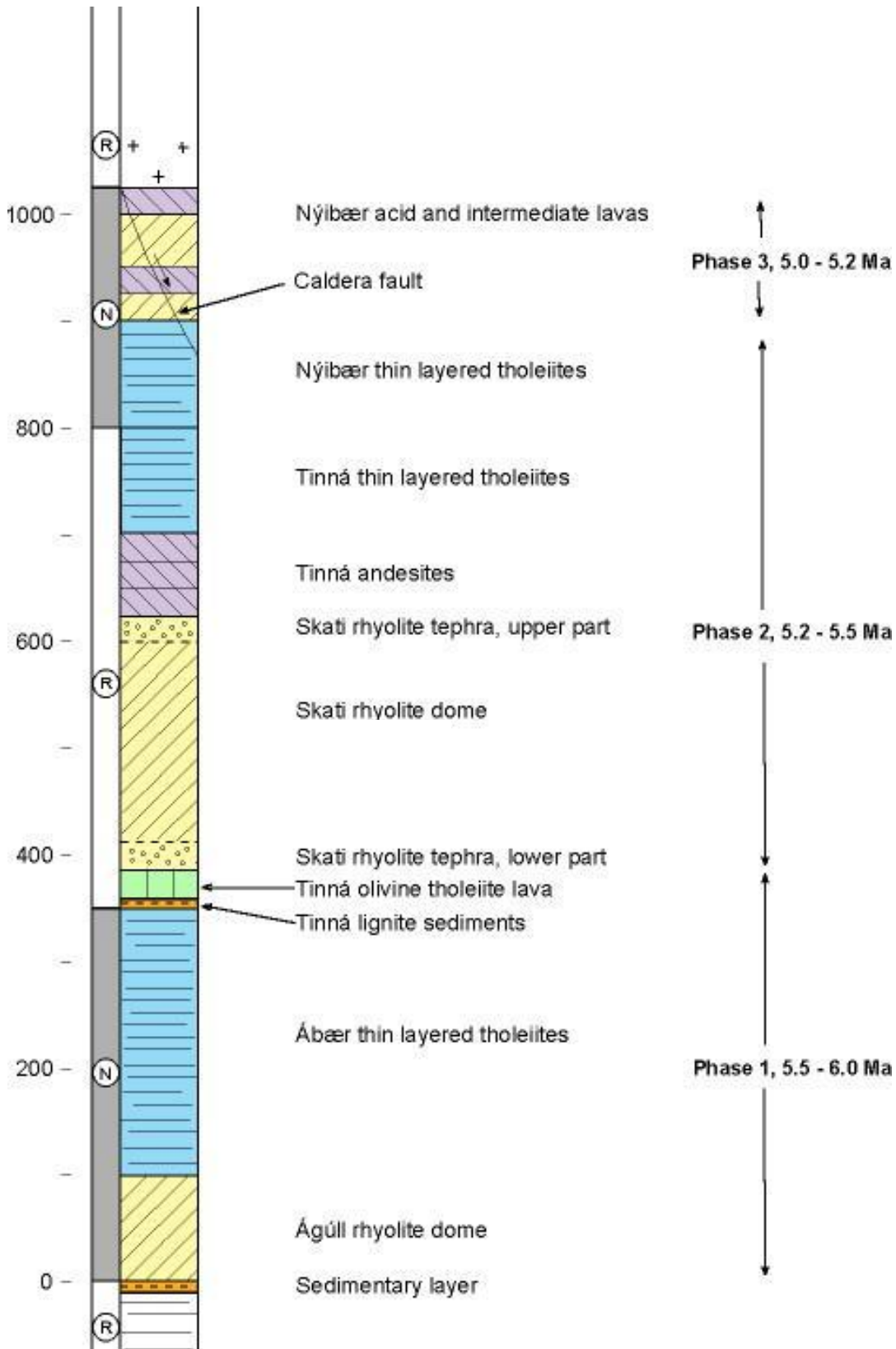
Fig. 23. Stratigraphy and history of the Tinná Central Volcano. A simplified geological column.

Table 9: The acid domes of the Tinná Volcano

Name	SiO$_2$ %	Lava km^3	Tephra (DRE) km^3
Ágúll dome		0.5	
Skati dome	75	8	10
Hvítárdalir dome	75	0.1	
Keldudalur dome	68	0.5	

The total volume of the central volcano is mostly based on estimations. The Ágúll sediment and the bottom of Ágúll dome are defined as its base. The top is at the lower level of Fossárdalur formation. The stratigraphical column on the enclosed Bedrock Map shows a 600 m thick rock pile between these levels. The average thickness is much less or around 300 m. The base of the volcano might be around 30 km wide and if it is assumed to be roughly circular its area is 700 km^2. Hence the volume would be 0.3 km^2 x 700 km^2 = 210 km^3. For comparison the Breiðdalur Central Volcano, East Iceland has been estimated to be 400 km^3 (Walker 1963).

The Tinná Central Volcano is unusual in many respects. The volume of the acid and intermediate eruptives is high, especially the monogenetic acid lava and tephra of the Skati dome, ti3 and ti4. Rhyolite domes, such as described here (table 9), are rare in Iceland. Only a few domes are known in the Tertiary areas. G.P.L. Walker (1963) describes such domes in the Breiðdalur Central Volcano which he calls the Röndólfur group of parasitic rhyolites. In the Quaternary areas the Hágöngur mountains in Central Iceland are the most prominent examples and from Holocene time no such dome is known to have been erupted. The lack of indications of a high temperature area is remarkable. Geothermal activity and high temperature areas are common for most of the rift-related central volcanoes of Iceland.

2.4 Torfufell Central Volcano

 It should be kept in mind that if the acid region in Eyjafjörður valley (the Torfufell area) belongs to the Tinná Central Volcano the above description only covers the western part of it. The Torfufell Central Volcano is located in the mountains of Eyjafjarðardalur valley. It has never been mapped or investigated with any accuracy. No caldera is found. It is in the strike east of Tinná Central Volcano and of the same age. They seem to be separated by the high mountains of Nýjabæjarfjall, but in fact this is only an apparent separation because the young strata pile in the uppermost part of the mountains hide their interconnection. Chemical analyses (tables 19 and 20) and petrogenic comparison have led to the assumption that the Tinná and Torfufell Central Volcanoes belong to the same volcanic system and are in fact one large central volcano (Hjartarson et al. 1997).

2.5 Fossárdalur Group

Fossárdalur group is composed of three formations; Fossárdalur porphyritic formation, Hölkná formation and Geldingsá formation. Here, sedimentary layers are coarser and make up a higher fraction of the pile than below. This is thought to be an indication of cooling climate and is also perhaps due to slower accumulation of eruptive layers.

2.5.1 Fossárdalur Porphyritic Formation

Fossárdalur porphyritic formation is dominated by porphyritic lavas and relatively thick sedimentary layers. It is about 100 m thick and can be divided in to three members.

1. *Fossárdalur lower porphyritic basalt (fo1)* is the lowest one, consisting of 2 - 4 basalt layers, 30 m altogether. The phenochrysts are of plagioclase with 15 - 20% of the rock volume.

2. *Fossárdalur sediments (fo2)* are made up of 10 - 20 m thick dark brown sandstone.

3. *Fossárdalur upper porphyritic basalt (fo3)*, the uppermost member, 50 m thick, consists of porphyritic layers similar to the ones in the lowest member.

More scattered porphyritic layers and tholeiite lavas are also found inside the formation. It has a reverse polarity (R). As a whole it forms a good marker horizon and differs from the adjacent formations, both in rock type and polarity.

The formation is found in the mountains at the inner part of Austurdalur valley but disappears under younger layers to the west. At the inner Vesturdalur valley one or two layers of fo1 appear again just below the unconformity opposite to the junction of Hraunþúfuá river. At the time of the Fossárdalur formation the Tinná Volcano's activity had ceased. Lavas from distant eruptive sites flowed along the slopes of the extinct and eroding mountain where sedimentary beds were washed away from the slopes on to the lavas all around. The lavas of Fossárdalur formation are the youngest Neogene lavas found in Vesturdalur.

The Fossárdalur sandstone (fo2) seems to indicate a hydrogeological boundary in the pile. Above it prominent fresh water springs are common but below it springs are small and rare.

Reference profile is in the northern slope of Fossárdalur at 600 – 725 m a.s.l.

2.5.2 Hölkná Formation

The Hölkná formation is characterized by tholeiite and porphyritic basalt with intermediate sedimentary layers. Its thickness is 150 - 200 m and it is only found in the mountains of upper Austurdalur and its tributary valleys in the southeast corner of the Bedrock Map. Its lower part has normal polarity (N) but at its top a few lavas display

reverse polarity (R). Great springs are to be found inside this formation in Fossárdalur valley. The spring area in Fossárdalur is the most productive one in the Tertiary areas of Iceland, issuing several cubic meters per second of cold (3,6°C) and clear water (Hjartarson et al. 1997).

Reference profile is found in Langahlíð at Austari Jökulsá at 580 – 720 m a.s.l.

2.5.3 *Geldingsá Formation*

A coarse diamictite layer and magnetic reversal indicate the divide between Hölkná formation and the above Geldingsá formation. This is the lowermost sedimentary layer with glaciogenic character that appears in the Skagafjörður pile. It is 8 - 18 m thick, made of clastic sediment (ge1) and is found in the upper Austurdalur and Hölknárdalur valleys. It rests on an eroded surface of adjacent lava. No glacial striation has been found and its age is very uncertain. Hjartarson and Friðleifsson (1997) have suggested that this layer might be correlated to the so-called Bót sediments in East Iceland, around 7 Ma. The Bót sediments are suggested to be a part of the sedimentary horizon located at the East Iceland Unconformity. Among the sediments is a coarse and widespread clastic layer. It is of Late Miocene age and is assumed to be the oldest tillite layer in the Icelandic strata pile (Hjartarson and Hafstað 1998). Ar/Ar dates published in this work (table 21) are not in favour of these ideas. According to them this layer is only about 3 Ma.

Geldingsá porphyritic member (ge2) covers the diamictite layer. It is at least 150 m thick composed of porphyritic lavas with plagioclase phenochrysts. Sedimentary layers are found in between, some of them with glaciogenic appearance.

Geldingsá formation is the youngest Neogene formation investigated and mapped so far below the Skagafjörður unconformity, and is only found in the mountains of upper Austurdalur valley. Two of its uppermost layers have been dated. (no. 17175 and 17173 in table 21). They give 2.49 Ma and 2.84 Ma respectively or upper Pliocene age (fig. 33). They are normally polarized but seem to be from different magnetic subchrons. The upper one fits to the short Reunion II cryptochron (C2r.2r-1) and the lower one are from the end of the Gauss polarity subchron (C2An.1n) (Cande and Kent 1995).

Only normal polarity (N) has been found inside the formation although it spans more than one subchron.

Reference profile is in the eastern wall of the canyon of Austari Jökulsá river, downstream of Geldingsá tributary.

2.6   Summary on the Neogene succession

The total thickness from the top of Sólheimar formation to the top of Geldingsá formation is around 1600 m. Proportional thickness of each rock type is given in table 10. The percentage of acid and intermediate rocks is much higher than in adjacent areas, such as Tröllaskagi (Sæmundsson et al. 1980) and Langidalur (Kristjánsson and Jóhannesson 1992). The reason is that here the section cuts the massive formations of the Tinná Volcano whereas the sections in Tröllaskagi and Langidalur avoid central volcanos.

Table 10. Thickness of different rock types in the Neogene succession of Sakagafjörður Valleys

Type	Thickness (m)	%
Tholeiite	850	53
Olivine tholeiite	100	6
Porphyritic basalt	190	12
Intermediate lavas	90	6
Acidic lavas	310	19
Sediments	70	4

Table 11. Groups, formations and members of the Neogene succession

Group	Formation	Member
		Geldingsá porphyritic member
Fossárdalur	Geldingsá formation	Geldingsá clastic member
	Hölkná formation	
	Fossárdalur formation	Fossárdalur upper phorph. basalt
		Fossárdalur sandstone
		Fossárdalur lower phorph. basalt
Tinná	Nýjibær rhyolite formation	
	Nýjibær andesite formation	
	Nýjibær tholeiite formation	
	Tinná tholeiite formation	
	Tinná andesite formation	
	Skati rhyolite formation	Skati Dome
		Skati tephra
		Hvítárdalir Dome
	Tinná olivine tholeiite formation	
	Tinná lignite sediment	
	Ábær tholeiite formation	
	Ábær rhyolite formation	Ágúll Dome
		Ábær sediments
	Fjóslækur formation	Fjóslækur member
		me4 member
Sólheimar	Merkigil formation	me3
		me2
		me1
	Sólheimar formation	

3   THE SKAGAFJÖRÐUR UNCONFORMITY

Paleo-landscape and erosion surfaces inside the Icelandic strata pile have been very sparsely investigated. Trausti Einarsson is about the only one among Icelandic geoscientists that has dealt with the subject (Einarsson 1958, 1958b, 1959, 1959b, 1961, 1972). Extensive erosion surfaces are known in several places, such as at the lignite bearing sedimentary horizons in W- and NW-Iceland, at the unconformities below the N-Iceland spreading zone at each side of it (in Fnjóskadalur and between Fljótsdalur and Vopnafjörður) and at the unconformity below the Snæfellsnes Pleistocene/Holocene formations. Some of these horizons have been investigated but mainly from the aspect of sedimentology and palaeontology. A mature paleo-landscape is only seen at two of these horizons, i.e. in Skagafjörður and Snæfellsnes. An imperfect study of paleo-landscape and a peneplain in the Tröllaskagi massif was carried out by Hjartarson (1973) and in later chapters his study will be summarized.

Líndal (1964) made some comments on the landscape below the Skagafjörður unconformity in his note-book that later was published, but he never made a special report on the matter.

Einarsson studied the landforms of N-Iceland, especially in the Skagi peninsula, and established a general hypothesis on the geomorphogeny of Iceland. Although a pioneer in paleomagnetic investigations and magnetostratigraphy, Einarsson was in other respects a loner in the geological community of Iceland. He was a dissident of the hypothesis on sub-glacial origin of hyaloclastite formations and also a stable and uncompromising opponent of the plate tectonics (Einarsson 1965, 1968). His ideas about the evolution of the topography and tectonics have never got much adhesion. The reason is mainly because he didn't involve plate tectonics in his analyse. According to his major morphological analysis in the book "Upper Tertiary and Pleistocene Rocks in Iceland" he divides the history of the landscape into ten steps (Einarsson 1961, see also Einarsson 1972):

1. Formation of the Tertiary lava pile (Miocene)
2. Volcanic intermission and formation of an unconformity
3. Outpouring of the Young Plateau Basalt
4. Tilting and faulting of the basaltic plateau
5. Peneplanation (Late Miocene)
6. Block faulting, uplift and formation of the high terrains
7. The 300 and 250 m valley erosion phases (Pliocene)
8. Eustatic drop to 100 m a.s.l., formation of the 100 m strand-plain (Early Pleistocene)
9. Migration towards current sea level. Formation of Skagi and Reykjavík dolerites, Pleistocene glacial erosion and complex history of tectonics in SW-Iceland
10. Accumulation of recent lavas and loose overburden (Holocene)

In the following sections of this thesis the development of the geomorphology in Skagafjörður valleys will be discussed. The discussion will not be based on the analyses of Einarsson although some analogy may be found. His first four steps are for example identical to stages no. 1 – 4 in table 12 below but the course of events is different. Step seven is also in a way identical to stages no. 3 – 4. No position will be taken as to his statements about the valley erosion phases, the formation of the 100 m strand-plain nor to the high eustatic sea level. It will however be pointed out that faint indications of very high Pliocene sea-level are found in Skagafjörður valleys.
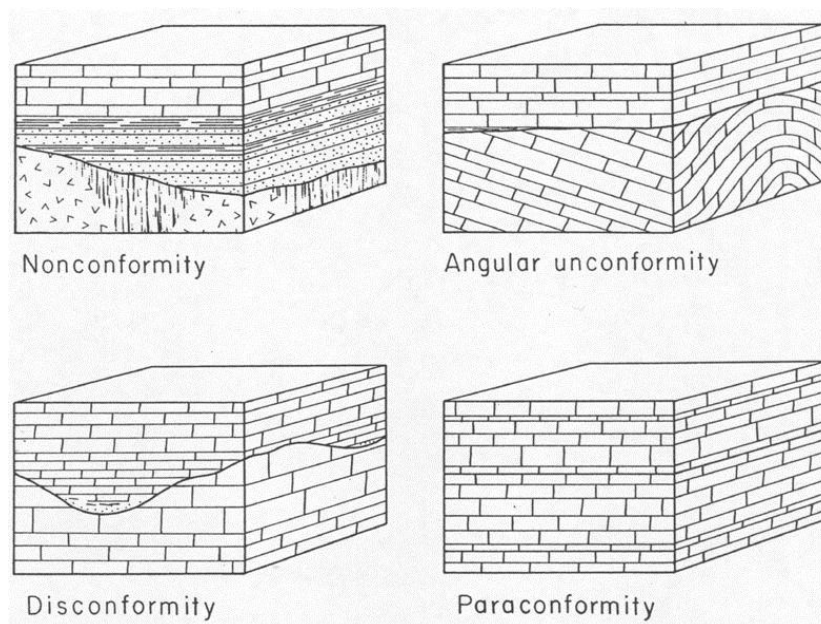
Fig. 24. The four types of unconformity (from Dunbar and Rodgers 1957).

3.1   Unconformity and hiatus – Definition of the terms

The main topic of this thesis is an unconformity in the bedrock as indicated in its title. Therefore this term – unconformity – should be defined precisely. It is, according to textbooks, a break in the sequence of strata in a certain area that represents a certain period of time during which no deposition took place. It may be a result of uplift and erosion or an interruption in the sedimentation. The absence of rocks, normally present in a sequence, indicates a hiatus in the geological record. There are four basic types of unconformity; disconformity, angular unconformity, non-conformity and paraconformity (fig. 24). In a *disconformity* the buried erosion surface, the hiatus, lies between two series of strata that are parallel on a large scale. *Angular unconformity* exists where the beds beneath and above are not parallel. This indicates tilting or faulting of the lower layers before being levelled by erosion and covered by younger strata. *Nonconformity* is an unconformity between overlaying stratified sediments and underlying unbedded metamorphic or plutonic rocks. P*araconformity* is similar to disconformity in that the strata are parallel. However there is a little discernible evidence of erosion or prolonged non-deposition (Dunbar and Rodgers1957).

As unconformity refers to the structural relations above and below a break in the geological record, a hiatus represents the break, or erosion surface, itself.

The Skagafjörður unconformity is in most places an angular unconformity but the southern most parts of it seem more like a disconformity.

3.2 The main stages in the morphological development

The development of the landscape in Skagafjörður Valleys can be divided into six main stages according to table 12.

Table 12: Main stages in the development of the Skagafjörður Valleys

No.	Stage	Time Ma	
1	Formation of the Tertiary volcanic succession	9 – 5	Accumulation
2	Peneplanation	~5	Erosion
3	Formation of the Pliocene valleys	5 – 3	Erosion
4	Formation of the sedimentary succession and the unconformity	3 – 2	Accumulation
5	Formation of the Pleistocene volcanic succession	2 – 1	Accumulation
6	Formation of the current landscape	1 – 0	Erosion

Jakob H. Líndal was the first to recognize the angular unconformity and break in the strata pile of Skagafjörður valleys as mentioned earlier (Líndal, 1964). Sedimentary layers separate in most places the older and younger volcanic successions. Here they will be called the sedimentary succession of the unconformity. The unconformity does not only mark an abrupt change in the strike and dip of the strata sequence. The sedimentary layers inside the successions below and above it are different. Below, most of them are rather fine-grained, with regular layering and seem to have been formed under moderate climatic conditions. Above, the sedimentary layers are much coarser and more irregular in structure and layering. Tillite layers, glaciofluvial and fluvial sediments and lacustrine and marine layers have been identified. These sediments seem to have been formed under energetic environmental conditions. The eruptives also change both petrologically and structurally. Compound olivine basalt lavas forming shield volcanoes and hyaloclastite formation, both very rare below the unconformity, become dominant in the upper pile. The chances in the structure of the stratigraphy will be described in later chapters but in this chapter the hiatus itself, and associated period of non-deposition, will be dealt with.

3.3 The genesis

The first stage, the genesis of the Neogene succession, has already been described. It lasted from 9 to 5 Ma. A major unconformity hides in this succession, the Hidden unconformity. The geological evolution of the region becomes rather complicated

because the Neogene succession seems to originate in two different axial rift zones. Its lower part seems to be erupted in the extinct Snæfellsnes – Húnaflói rift zone. The succession drifted towards east, away from the volcanic belt and slowly became a part of the rigid Eurasian crustal plate. But it didn't remain there. Because of a rift jump 7 Ma and the ignition of the North Iceland rift zone the region was suddenly located inside a volcanic belt again. There it was buried below the upper part of the Neogene succession. But now it drifted towards west to become part of the North-American crustal plate. Thus this region has experienced spreading towards both east and west and location inside two different crustal plates. The matter becomes even more confusing as volcanism started for the third time during the Skagafjörður spreading event, but that is another story.

From the Bedrock Map of Skagafjörður valleys it is evident that the hiatus and the age gap represented by the unconformity is widest in the northernmost part of the area where it spans several millions of years but becomes narrower towards south. (This will be discussed in the chapter 10, handling the age of the strata pile).

3.4   Peneplanation and erosion of the primordial lava plain

The period of the peneplanation started when the area drifted gradually away from the active volcanic zone. Accumulation of volcanic extrusives stopped and erosion took over, shaping the primordial Neogene lava plain and carving out new landscape. This might have happened near the end of Miocene and in the beginning of Pliocene, 5 Ma. A long-lasting period of continuous erosion started. In the beginning, when the area was still inside the volcanic zone, and during the first ages after it had drifted out of it, the landscape was supposedly relatively flat with widespread lava plains between crater rows and low hills. The most mountainous areas were the central volcanoes but in this case the major structures of the Tinná Volcano were mostly buried under younger formations. Hyaloclastite ridges and tufas, scenically one of the most conspicuous features of the current volcanic landscape, didn't exist at this time and shield volcanoes were rare. The hydrology was on the other hand probably similar to what it is inside today's volcanic belts. Most of the precipitation penetrated into the ground forming groundwater streams, issuing in great spring areas at the borders of the lava plains. Glaciers didn't play a big role if they existed at all. The erosion was of low degree, perhaps mainly in form of wind and chemical weathering. The remnants of this relatively flat and uniform landscape can still be seen in the even elevation of the flat mountaintops in the Tröllaskagi massif. This stage in the development of the landscape is the peneplanation in table 12 and it might have terminated around 5 Ma. It might be questioned wheter the peneplanation should be defined as an independent morphological state as it is so closely related to the primordial lava plain.

As already mentioned, Hjartarson (1973) studied the peneplanation of Tröllaskagi massif and attemted to reconstruct the apparent ancient plateau. The main conclusions are that it is possible to map and reconstruct a continuous plateau across the southern part of Tröllaskagi, from Eyjafjörður to Skagafjörður. Both the fjords seem to appear in the shape of the peneplain, but in a different way. The peneplain is sloping gently towards Skagafjörður but Eyjafjörður is more like a rift zone extending into it. In the northern part of Tröllaskagi and especially in an area on both sides of Eyjafjörður, the plateau is totally destroyed and eroded away. The mountaintops are all lower than the peneplain. The

remnants of the plateau lie between 1000 – 1500 m a.s.l. It has two summits, at 1450 m near Skíðadalur valley and 1500 m in Mt. Kerling near Akureyri. The original elevation of the plateau must have been much lower but along with the erosion of the landscape, and removal of the material, appreciable isostatic uplift is supposed to have taken place. The ablation of the primordial lava plain during the peneplanation is hard to estimate. The uppermost part of the lava pile is unaltered and fresh and in many cases the upper level of the zeolite zones is 400 – 500 m below the peneplain. According to that the general ablation might be in the order of 100 – 200 m in most places (Walker 1960). In some places the erosion seems to have been more effective. Hjartarson and Jónsdóttir (1999) however have described and mapped a large, layered intrusion exposed at 1000 – 1350 m a.s.l. near Mt.Kerling. Such intrusion could hardly have formed close to the surface and should therefore indicate a high-laying primordial surface level at this place and considerable ablation during the peneplanation in the area.

As time vent by, the lava planes became less permeable, the surface waters grew more efficient and the weathering rate increased. The period of the peneplanation was over. The plateau began to erode and a valley system started to form. In the beginning it might have been dominated by shallow V-shaped valleys excavated by fluvial erosion. The first full scale glaciations with glaciers occupying the main valleys and fjords are supposed to have occurred 2.4 Ma (Geirsdóttir 1990). Before that time smaller, local glaciations had overridden the central highlands of Iceland several times. Glacial erosion might have resulted in a system of U-shaped valleys already in late Pliocene. At the end of the Pliocene a fjord and long, deep valleys had developed with steep slopes carved out into the plateau. This system can be followed for at least 125 km from the northernmost shores of Skagafjörður towards south, along the fjord, through the major valley of the district and along the main tributary valleys, Austur- and Vesturdalur, to their heads in the Central Highland north of Hofsjökull.

3.5   Size and shape of the Plio-Pleistocene valleys

3.5.1   The Principal Valley

The Tungusveit parish, in the central part of the main Skagafjörður Valley, there is a hillocky spur between two of the main rivers of the district, Héraðsvötn and Svartá. There the sedimentary succession of the unconformity forms a thick pile resting on the floor of the old Plio-Pleistocene valley. This site, called Eggjar and Stapar, is around 40 km inland from the present shore. The elevation of the floor is at about 70 m a.s.l. and below the northernmost part of the sedimentary pile the ancient valley floor disappears down into the current glaciofluvial plain of Héraðsvötn river. There the erosion level had already, before the formation of the sediments, become lower than it is today, but the valley floor also seems to have been steeper. The size and shape of this primeval valley cannot be reconstructed in any details but the mountains at each side were about 1000 m high. Mælifellshnjúkur, the proud landmark of the district, now towering as high as 1138 m, didn't exist at this time. It is a hyaloclastite peak, belonging to the Pleistocene formation, formed in a subglacial eruption. It rests unconformably upon the old Tertiary plateau at around 900 m a.s.l. at the west side of the valley (fig. 25).
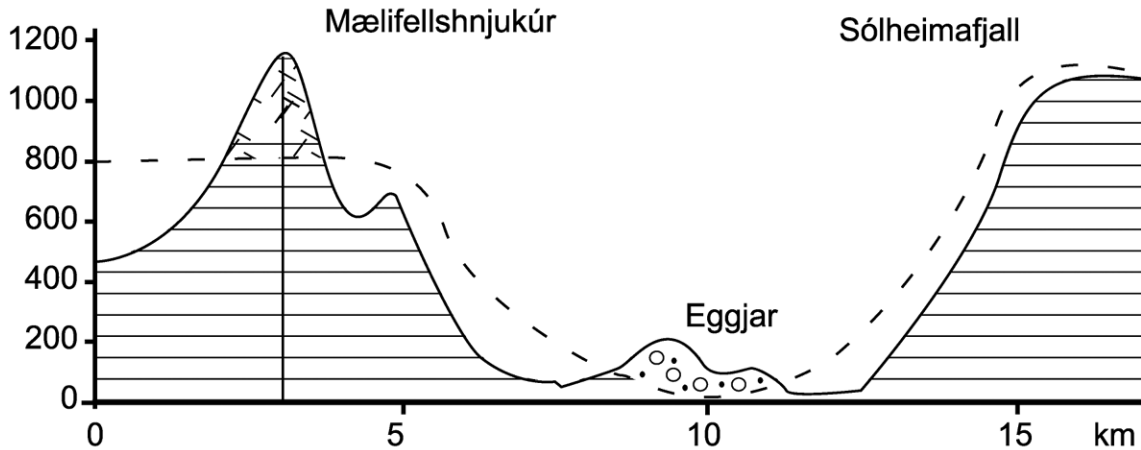
Fig. 25. Skagafjörður principal valley a cross-section along the strike. Horizontal lines = The Neogene lava pile. Circles and dots = The sedimentary succession of the unconformity. The Pleistocene hyaloclastite in Mælifellshnjúkur is resting directly on the Neogene pile.

At the east side, in Tröllaskagi highlands, the mountains were even higher and the old plateau is seen at 1100 m. The ancient valley seems to have been an U-shaped valley, as deep or even deeper than the current valley, but narrower.

3.5.2   Vesturdalur valley

Vesturdalur seems, at this time, to have been more or less a direct continuation of the principal valley but not a tributary valley as it is today. The ancient form of the outer Vesturdalur is considerably well preserved below the Pleistocene pile. It was a U-shaped valley, eroded into the old plateau along the western side of the rhyolite body of the Skati dome. The ancient valley floor can be seen at around 230 m a.s.l. or about 30 m higher than today. The width of the flat floor seems to have been near 1000 m or similar to what it is now but the slopes may have been a little more leaning and the mountains at each side lower than today. Later on, sometime in the Mid-Pleistocene, this deep primeval valley was completely filled up by sediments and lavas as can be shown on fig. 26. The unconformity is found at both sides of the valley and can be followed from its mouth near Goðdalir and 30 km inland. Its continuity is only broken in the western slopes of the Skati dome. There all traces of the unconformity are worn out. The ancient valley might have been considerably larger than today, i. e. having larger catchment area. Later on Austurdalur occupied part of it.
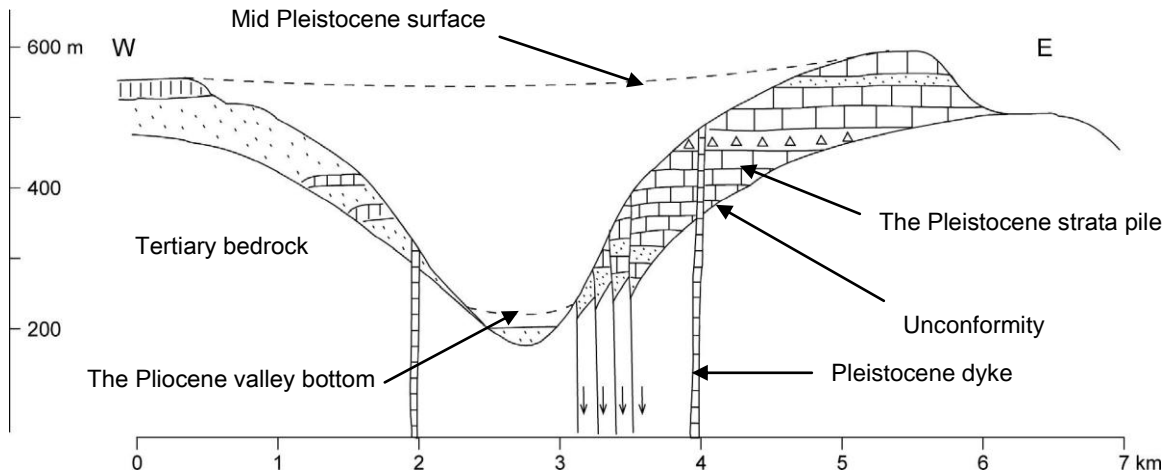
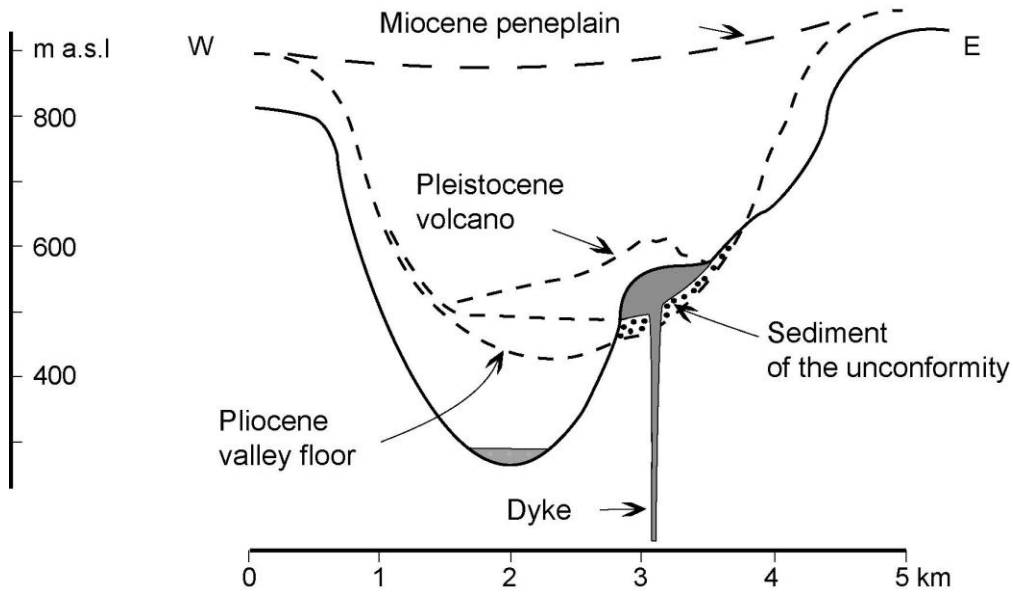*Fig. 26. Vesturdalur cross-section. The picture shows the Pliocene valley, the unconformity, the Pleistocene lavas and sedimentary layers that filled up the valley, Pleistocene dykes and faults and the shape of the valley today.*

The situation in Vesturdalur today is unusual in the sense that the main river, Vestari-Jökulsá, that is thought to have carved out the valley, doesn't stream along it. Instead it flows along a shallower tributary valley, Goðdaladalur, west of it in a deep canyon and doesn't reach Vesturdalur until near its north end. This seems to have been the case for the whole Holocene time. This must be due to some topographic changes under the ice cap of Hofsjökull that have diverted the river from its traditional course and towards Goðdaladalur. Geological evidences show that Goðdaladalur existed already in early Pleistocene time as a tributary valley to Vesturdalur.

3.5.3   Austurdalur valley

Austurdalur is eroded into the old plateau along the eastern side of Skati dome. The ancient erosion surface is not as continuous there as in Vesturdalur but it can be investigated here and there below the early Pleistocene formations. The most instructive place is below the Austurdalur Pleistocene Volcano at the junction of Austurdalur and Tinnárdalur. An eruption in the valley built up the volcano, as will be described in details later (chapter 5.1.1). It rests on thick fluvial sediments that had accumulated in the valley itself before the eruption. Fig. 27 shows a cross-section in Austurdalur and imaginary outlines of the ancient valley. It might have been a little narrower but 100 - 200 m shallower than the present valley. Further inland, at the tributary valleys of Hvítá and Keldudalur, the erosion surface appears again below younger sediments and lavas, reflecting a valley floor at 600-650 m a.s.l. The floor seems to have been a little more sloping upstream, along the river, than today.

Fig. 27. Austurdalur cross-section. Traces of the Pleistocene volcano can be seen in its eastern slopes along with the feeder dyke resting on the remains of old sediments. The dashed lines indicate early stages of the development; The Miocene peneplain (~5 Ma), the Pliocene valley and its sedimentary layer (~2 Ma). The Pleistocene volcano(~1.5 Ma) is also outlined. To day the valley is 600 m deep but in Pliocene it was 400-500 m deep.

According to this, Austurdalur was not as big and evolved in the early Pleistocene time as Vesturdalur but now the valleys are both in a similar state of erosion. The reason is, as will be described in the following chapters, that Vesturdalur was completely filled up by lavas in early Pleistocene while Austurdalur was only partly filled.

Ancient sediments in the slopes of the tributary valleys Tinnárdalur and Merkidalur prove their existence already in early Pleistocene.
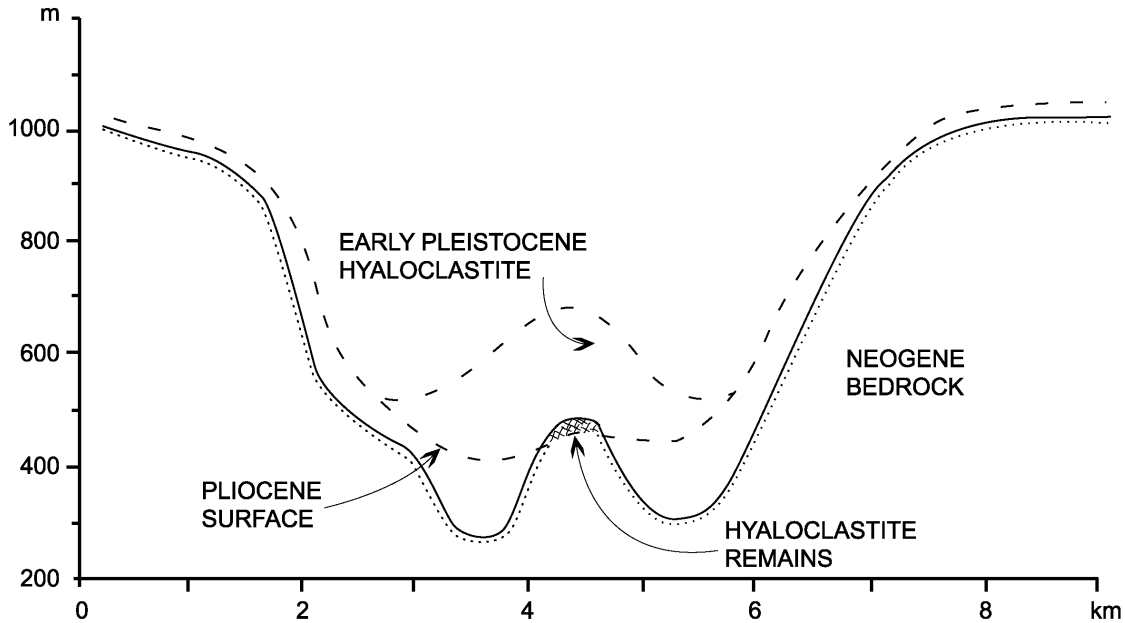
Fig. 28. Heiðarsporður hyaloclastite. Cross-section in Norðurárdalur valley.

3.5.4   Norðurárdalur Valley

Norðurárdalur is one of the main valleys of Skagafjörður District and inhabited as Austur- and Vesturdalur. Indications of its Plio-Pleistocene existence and its ancient shape are preserved below early Pleistocene volcanic formation at two sites inside the valley. These formations are the Kotagil lava and Heiðarsporður hyaloclastite. Their stratigraphy will be described in chapters 4.4, 5.1.3 and 5.1.4. The Heiðarsporður hyaloclastite is most likely formed subglacially by an eruption in the upper part of the valley. Its base is resting on a thin sedimentary layer at between 410 and 510 m a.s.l. Here the bottom of the valley is near 280 m a.s.l. At both sides the mountain summits reach 1100 m (fig. 28). According to this Norðurárdalur was at least 700 m deep in early Pleistocene and has only been deepened by 120 m during the last one million years of full-scale glaciation. The Kotá lava is a remnant of early Pleistocene interglacial lava erupted near 610 m a.s.l. in a little tributary valley in Norðurárdalur. It doesn't only indicate the existence of the main valley but it reflects as well the early development of its tributaries.

3.6   Conclusions

The conclusion of this chapter on the unconformity is that already at the end of Pliocene, before the onset of the Pleistocene glaciations, a mature valley system had evolved and been carved out in the primordial lava plain of the Skagafjörður District. In the principal valley the erosion level was lower than at present and the mountains at each side as high or even higher. Vesturdalur was a direct continuation of the principal valley, considerably larger than today, taking over part of the current catchment area of Austurdalur. The main tributary valley, Austurdalur, Svartárdalur and Norðurárdalur, were deeply eroded and many of the small tributaries of the second order also existed.

Excavation of the 1000 m deep and several km wide Pliocene valleys in Skagafjörður in only 3-4 million years indicates a very high erosion rate and might require a special explanation. Subsidence of the area because of spreading and normal faulting prior to the Pleistocene volcanism could be part of the explanation but only part of it because the valley system seems to be older than any possible effects of the Skafagjörður spreading event.

By analogy it can be predicted that all the main fjords and valleys in the Tertiary regions in Iceland are old in Icelandic geological context and already existed in Pliocene time.

This conclusion is in contradiction with the statement of Einarsson (1968, 1978,1998). He describes the topography as it had been a rather even and continuous volcanic plateau, without deep valleys and high mountains. According to his textbooks the present topography of Iceland is mainly due to the glacial erosion during Pleistocene.

The conclusion is more in line with Trausti Einarssons (1961) interpretation of the age of the landscape. There he states that the main valleys and valley systems of Iceland were deeply eroded already in Pliocene.

4 THE SEDIMENTARY SUCCESSION OF THE UNCONFORMITY

After the formation of the early landscape the valleys started to fill up again, for some unknown reason. Thick layers of sediments were accumulated leading to the formation of the sedimentary succession of the unconformity (table 12). The unconformity itself is the exact contact between the eroded top of the Tertiary volcanic pile and the base of over-laying beds. The sedimentary succession is defined as the strata found between the surface of the unconformity and the earliest Pleistocene lavas. The sediments found in between the lavas are defined as a part of the Pleistocene volcanic group. The sedimentary succession of the unconformity can be traced for long distances inside the research area in between the Neogene (or Tertiary) and Pleistocene volcanic successions. Its internal structure is variable from place to place as well as its thickness. It is most obvious in the slopes of Vesturdalur valley and its tributaries but instructive sections are also found in Austurdalur and Norðurárdalur valleys, in the principal valley of Skagafjörður and in Skagi Peninsula as already mentioned. In this chapter several sections through the sedimentary succession will be described and at the end the environmental conditions during the sedimentation will be discussed.

4.1 The Principal Valley

As mentioned in the last chapter, Plio-Pleistocene sedimentary layers are to be found in the Tungusveit parish in the principal Skagafjörður Valley. There they form an up to 200 m thick, uncapped sedimentary strata pile. The Pleistocene lavas that probably capped these sediments are entirely eroded away if they ever existed at all. No dykes or veins have been found in the sediment.

The sedimentary sequences can be divided into 5 units. Each unit is of variable thickness from place to place:

1. Sandstone resting on the Neogene bedrock
2. Fine-grained conglomerate, pebbles up to 10 cm, with lenses of sandstone
3. Conglomerate, medium grained
4. Coarse, bedded conglomerate with boulders up to 100 cm in diameter
5. Diamictite, erratics up to 220 cm in diameter along the major axes

The layering is in general horizontal and has a coarsening upwards trend. All the formation is semi-consolidated. No fossils have been recognized. The internal relations of the units have not been studied and the whole formation remains to be carefully investigated.

The diamictite unit is in the upper part of the sedimentary pile. It is 5 – 10 m thick. Layering is irregular and absent in many places. It is poorly sorted with wide range of grain size from silt to boulders. Clasts are well rounded to angular. The unit is gently sloping towards north but that seems to be a landscape slope but not a tectonic. The best exposure of the diamictite unit is near a top of a hill north of the farm Stapi. The

diamictite is interpreted as tillite and thus it indicates a glaciation of the Skagafjörður lowlands in Late Pliocene – Early Pleistocene times.

4.2 Vesturdalur valley

The sedimentary succession of the unconformity is found in the mountain slopes at both sides of the Vesturdalur valley, i.e. in Goðdalafjall mountain at the west side (fig. 30) and Hlíðarfjall mountain at the east side (fig. 32). It can be traced for 30 km upstream the valley in various thickness but is absent in a few places in between.

4.2.1 Goðdalafjall

Very thick sediments are found at Djúpugil gullies in Goðdalafjall mountain, in the north west corner of the Bedrock Map. There they are at least 325 m thick with one basalt layer dividing them into two parts of nearly equal thickness. On top it is capped by a single basalt flow. By definition only the lower part of the sediments belong to the sedimentary succession of the unconformity. The upper part belongs to the Pleistocene group. Farther south in Goðdalafjall, and in Hlíðarfjall at the other side of the valley, the upper part of the sediments contain several interbedded lava flows.

The sediment in Djúpugil is mostly hidden below Weichselian till coat so its layering is unclear. The base is somewhere near 250 m a.s.l. The lowest 100 m are totally covered but at 350 m a.s.l. coarse-grained conglomerate has been excavated with irregular layering. Boulders are up to 80 cm across. At 390 m a.s.l. the sediment is capped by 10 m thick tholeiite lava. The Ar/Ar-age of this lava is 1.48 Ma (see table 21). In the summer of 2000 fragments of balanus were found near 330 m a.s.l. in the Weischelian till coat that covers the early Pleistocene sediments in Djúpugil, where the fragments most likely originate. They were observed by Leifur Simonarson, professor of palaentology at the University of Iceland. The shell fragments are slightly altered and deformed but most likely they belong to Balanus crenatus (Simonarson, pers.comm.). This species has been found in Pleistocene fossil sites in Iceland and still exists along the shore all around the country. In spite of later search more fossils have not been found.

Although the upper part of the sediments at Djúpugil does not belong to the sedimentary succession of the unconformity it will be described here. On top of the lava in the middle of the section there rest 30-40 m of brecciated hyaloclastite but higher up is a 150 m sedimentary sequence (440-590 m a.s.l.) with layers of sandstone, conglomerate and diamictite. The conglomerate makes up the topmost part of the sediment. It is 20-25 m thick, horizontally stratified, coarse gravel. Inside it is a layer of extraordinary coarse material, packed with rounded boulders, 1-2 m across (fig. 29). The conglomerate is interpreted as fluvial or glaciofluvial material and indicates the flow of a river with high transport ability. The coarsest layer must reflect huge river flooding. The sediment is capped by a nicely columnar jointed tholeiitic lava flow from early Pleistocene times with reverse polarity, comprising the summit of Goðdalakista. The Ar/Ar-age of this lava is 1.48 Ma (see table 21). No signs of soil or vegetation have been found at the base of the lava so it seems to have covered the desolate flood plain of the river. Thus this 150 m thick sedimentary sequence is formed in a period of 150,000 years.

Fig. 29. Exceptionally coarse fluvial sediments in Goðdalafjall. The largest boulders are 200 cm in diameter.

The best section in Goðdalafjall is found in the Jórgil gorge, 2.5 km south of Djúpugil. There the section starts at 250 m a.s.l. (fig. 30). The lowermost part of the sediment is covered by late Weischelian and Holocene deposits. In the exposure 115 m thick sedimentary pile can be investigated. The layering is sometimes irregular but more often roughly horizontal. The grain size is remarkable. Rounded boulders up to 1 m across are sitting in a sandy and silty ground mass. The grain size has upward grading, i.e. with fine-grained lower part but coarser upper part. Most of the beds seem to be of fluvial and glaciofluvial origin reflecting glacial and periglacial environment and existence of rivers with high transport ability. A dyke, cutting the layers in the section, shows that volcanic activity took place in the valley itself after the formation of the sediments, as will be discussed in later chapters.

The sediment is covered by thin olivine-tholeiite lava. It seems to have been very fluid when it covered the area because it is very thin (1-2 m) and gravel and boulders from the underlying sediment are like inclusions in the bottom layer of the lava. No signs of soil or vegetation are found at the contact.
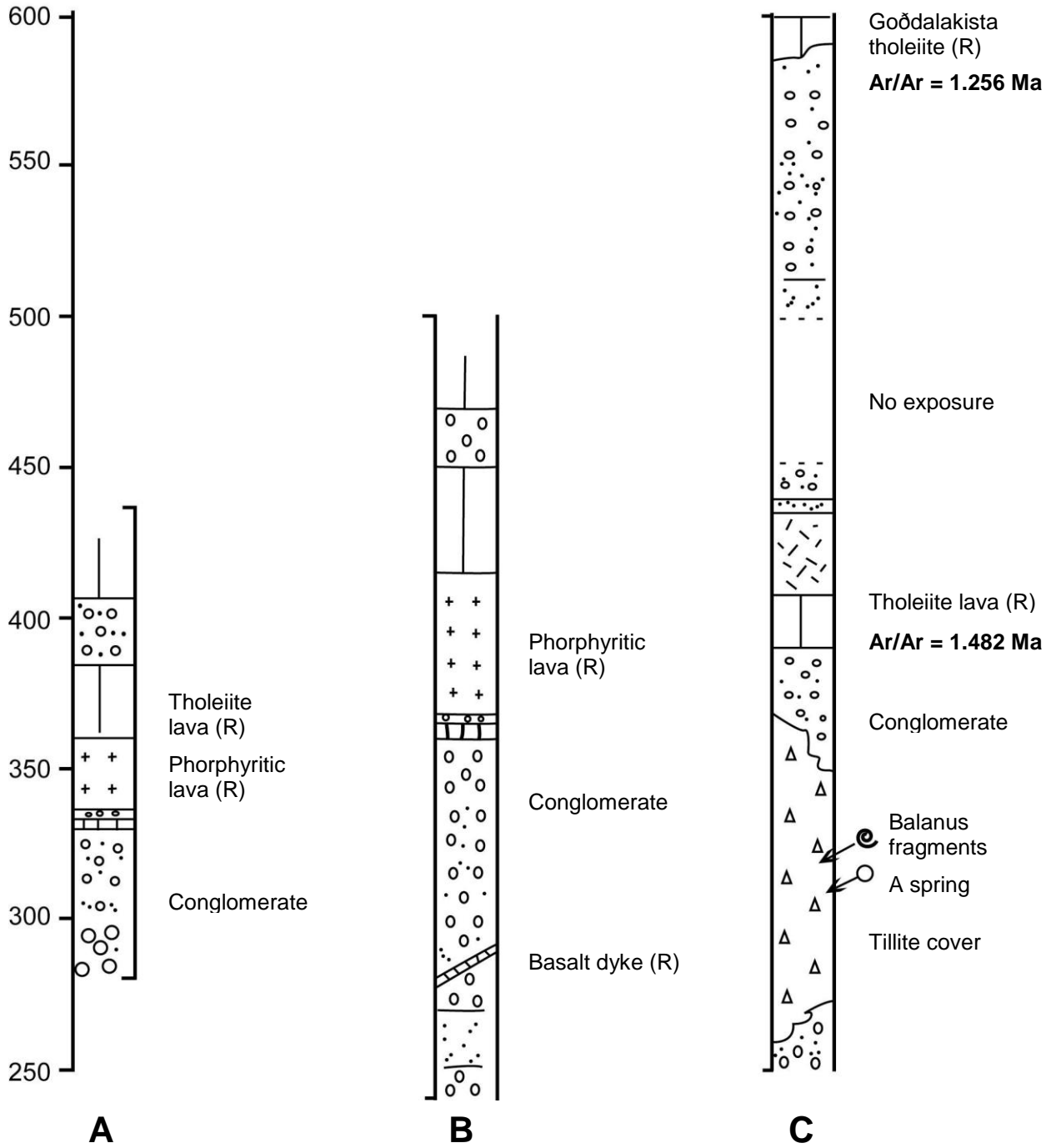
Fig. 30. Sections from Goðdalafjall. A) Dalakot B) Jórgil B) Djúpugil

Fig. 31. Rounded boulders at the bottom of the sedimentary layer of the unconformity in Hlíðarfjall in Vesturdalur valley at 290 m a.s.l.

4.2.2 Hlíðarfjall

Informative sections are located in Hlíðarfjall mountain, shown in fig. 32. In the first two the contact between Neogene and Pleistocene rocks is exposed. There, at 290 m a.s.l., there is a 8-10 m thick layer of surprisingly big, rounded and polished boulders, up to 2 meters in diameter with fine gravel and sand packed in between. The boulders are of olivine tholeiite and originate from the Neogene base rock just below the sedimentary layer and look like erratics on a sea-shore. Together with the balanus fragments from the opposite slopes of the valley indications of exceptionally high shoreline are found here. The thickness of the Plio-Pleistocene sediment layers is 50-60 m.

Near the farm Bjarnastaðahlíð two parallel gorges are cut into the mountain slope where the bedrock is well excavated. But the stratigraphy is rather confusing because at least four normal faults, all with 10-20 m throw towards east, hide parts of the sedimentary pile (see fig. 26). These faults are important because they give information about the early Pleistocene tectonics, which will be discussed later.

Ar/Ar-date on one of the capping lava flows gives 1.6 Ma (table 21). Below the sediment there is 7-8 Ma Neogene bedrock so here the unconformity spans 6-7 Ma.

Fig. 32. Sections from Hlíðarfjall. A) Hlíðarfjall north B) Hlíðarfjall C) Bjarnastaðahlíð.

4.2.3   Upper Vesturdalur (Hraunþúfuá – Runa)

The sediment of the unconformity can be traced to the upper Vesturdalur, upstream from the confluence with Hraunþúfuá river. There it d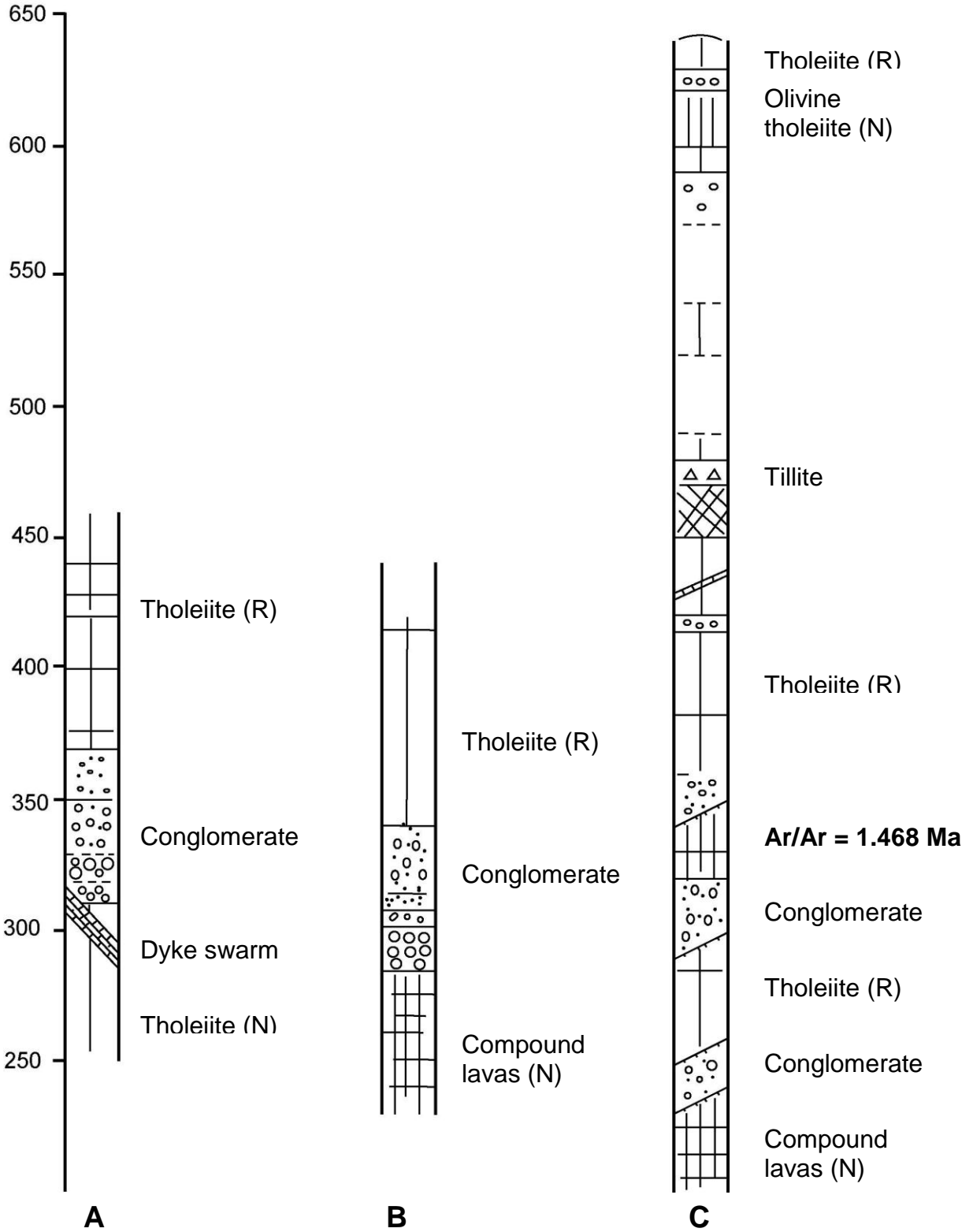isappears down below the valley floor due to increasing elevation of the floor and gentle southerly dip of the strata pile. At this site the sediment is 20 m thick. The lower 16 m are made of fluvial and glaciofluvial sandstone and conglomerate but on top of it are 4 m of diamictite with clasts up to 60 cm in diameter. This is interpreted as tillite.

The tributary river Hraunþúfuá in the upper Vesturdalur valley has cut an excellent section through the sediment. There it is capped by the thick Hraunþúfa compound lava (hþd) of mid Pleistocene age. At the outer part of the tributary valley the sediment is divided by a thick columnar-jointed lava into a sandstone layer (8 m) at the bottom and conglomerate (4 m) on the top. At the inner part of the tributary valley the lava has disappeared but instead of it there are over 40 m of conglomerate made of coarse grained fluvial gravel and boulders with horizontal layering.

4.3   Austurdalur valley

In Austurdalur valley the sedimentary pile of the unconformity is discontinuous. It is found in several places along the slopes of the valley as erosion leftovers, in some cases capped by younger lavas but in others not. The most prominent place is below the lava cap of Austurdalur Pleistocene Volcano where the sediment has been preserved by the lava.

4.3.1   Sediment below Austurdalur Pleistocene Volcano

As will be described in later chapters an eruption took place in Austurdalur in early Pleistocene and a lava flow covered the flood plane of the valley floor. Later on most of the lava and underlying sediments have been weathered away, but small patches of the lava are still left resting on the ancient sediment. The thickness of the sediment below the Austurdalur Pleistocene Volcano varies from place to place, reaching at least 90 m where thickest. The lowest 2-3 m are made of coarse gravel and rounded boulders weighing up to one ton or so. Rhyolite rock fragments are abundant, derived from the nearby Skati acid dome. It is covered by 30 – 40 m of layered conglomerate of fluvial or glaciofluvial origin. The ground mass is made of sand and gravel but boulders, 20 – 30 cm across, are common. Then come 25 m of slightly finer conglomerate. On top of it are 30 m of fine gravel and stratified sand. The top of the layer is at ca. 500 m a.s.l. or 200 m above the present flood plain of the valley. The topmost 5 m are of silty sand. The layers can be traced for 4 km along the slopes of Austurdalur at each side of Tinnárdalur mouth. No diamictite has been found here. The sedimentary beds in general become gradually finer upwards in the sequence. In some other sites the beds display a coarsening upwards sequence.

The lava at the top will be described in the chapter of the Pleistocene volcanic succession (see chapter 5.1.1).

4.3.2 Keldudalsmúli

Continuation of these fluvial layers can be observed 5 km farther inland, below Pleistocene lava caps at each side of Austurdalur, i.e. at Hvítá tributary river and in Keldudalsmúli. In the latter locality the sediment appears as 50 m thick conglomerate (at 615-665 m a.s.l.), made of coarse glaciofluvial material such as below the Austurdalur Pleistocene Volcano but more robust (fig. 22). Rounded boulders are up to 55 cm in diameter. Farther upstream, between Keldudalsmúli and Geldingsá river, the layers become more like diamictite.

4.3.3 Geldingsá tributary river

In the farthest inland site, at the Geldingsá river in 630 m a.s.l., the sedimentary layer is entirely made of poorly sorted diamictite with crude layering, rich in the silt fraction, dark grey in colour. Erratics are up to 100 cm in diameter. The layer seems to be of glacial origin. There it is up to 55 m thick (fig. 33). In one place near the junction with Jökulsá a basaltic dyke is seen cutting the layer. The sediment can be traced for long distances at both sides of Austari Jökulsá. The lava layer below has a weathered and polished surface. Glacial striation has not been found.
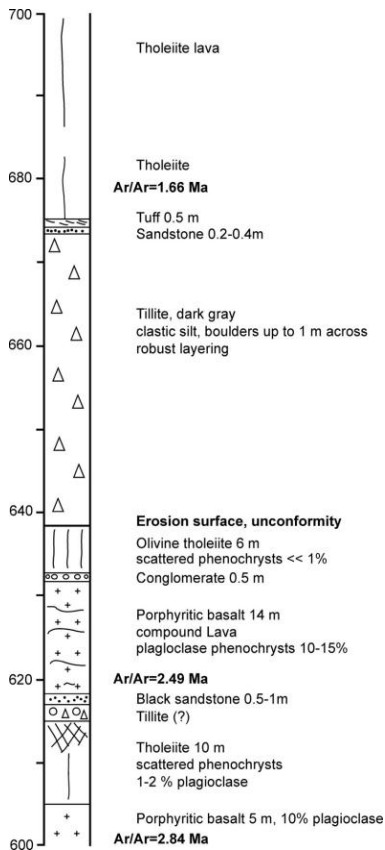
Fig. 33. A section across the unconformity at Austari Jökulsá river near Geldingsá north of Hofsjökull. Samples for dating were collected above and below the thick diamictite layer.

Ar/Ar-dates were made on the lavas at each side of the sediment (fig. 33, see also table 21). They gave 2.49 Ma and 1.66 Ma respectively. The older age might represent the Reunion II cryptochron. According to that the sediment is sitting inside a gap in the strata pile that spans 0.8 million years. The sediment itself seems however accumulated in a short period of time.

4.3.4 Merkidalur tributary valley

Merkidalur tributary valley is cut towards east through the eastern slopes of lower Austurdalur into the Nýjabæjarfjall mountain. The majestic Merkigil Gorge is cut in the Austurdalur slopes below the mouth of the Merkidalur. At 500 – 600 m a.s.l. in the southern slopes of Merkidalur there is a 50 – 60 m thick layer of early Pleistocene sediment (fig. 34). Although no lava cap is seen on top of the sediment such rock has most likely covered it in earlier times but has been completely eroded away. Otherwise the sediments would hardly have been protected. In Ytrigjá ravine the sediment is composed of stratified conglomerate with layers of sandstone but on top of it there are 24 m of diamictite in a vertical wall. Rounded and striated boulders, weighing 4 – 5 tons, are sitting in a sand – silt ground mass. Farther south, in Víðagjá ravine, a 20 m thick lens of disturbed and broken bedrock – where the layering is very irregular, steeply dipping and even vertical – is sandwiched in between a thick diamictite layer at the top and under laying conglomerate. The lens is interpreted as remnants of a rockslide. A rockslide like this one can only occur in steep slopes and and it most likely indicates a slope failure in the ancient Merkidalur valley. The diamictite is interpreted as early Pleistocene tillite.
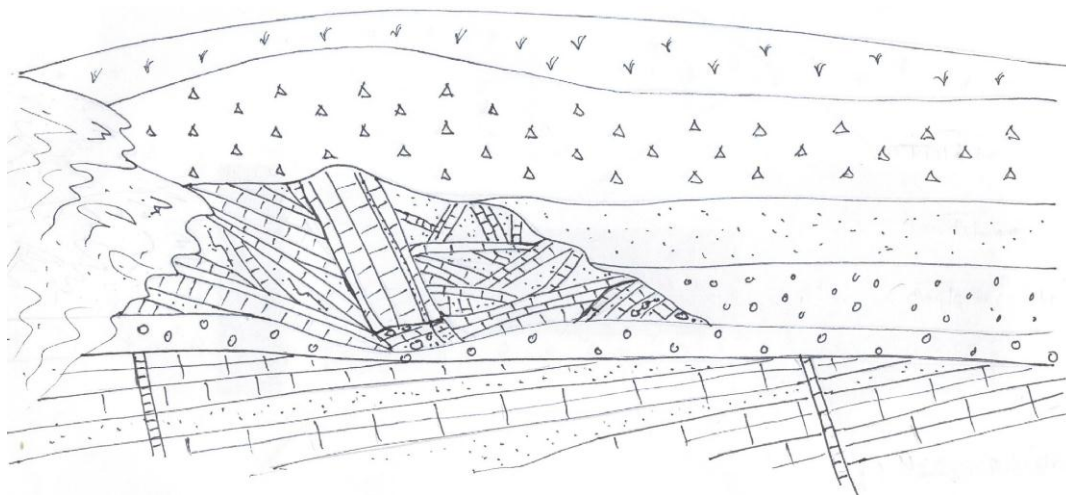
Fig. 34. The sediment in Merkidalur (Ytrigjá and Víðagjá). Remnants of a rockslide sandwiched in between a tillite layer on the top and under lying conglomerate.

4.4 Norðurárdalur valley

As already mentioned, Norðurárdalur is one of the main valleys of Skagafjörður District. Highway no. 1, the Ring road, passes through it and from there it crosses the mountain pass Öxnadalsheiði, towards Eyjafjörður. The sediment of the unconformity is found below early Pleistocene volcanic formation at two sites inside the valley, at Kotagil ravine and in Heiðarsporður spur.

At 610 m a.s.l., on the eastern brink of the Kotagil ravine, a remnant of Pleistocene lava is found resting on 90 m of sediment. The sediment is made of stratified conglomerate with variable grain size sitting between 520 and 610 m a.s.l. (fig. 35). No diamictite is found. The layers are semi-consolidated and near their bottom small coldwater springs seep out. The sediment, which is only partly capped by the lava, extends further upstream along the brink of the ravine. The ravine has been deepened for ca. 100 m since the accumulation of the sediment. According to that the Kotagil valley and the deep ravine cut into its bottom already existed in early Pleistocene. (A map of the site is in fig. 43).

At Heiðarsporður spur a thin layer of sediment is found below Pleistocene hyaloclastites resting unconformably on the Tertiary lava pile. In most places it is only 10-20 cm and sometimes absent at all. (A map of the site is in fig. 44).

The sites at Kotagil and Heiðarsporður will be described in more details in the chapter on the Pleistocene volcanic succession (chapter 5).
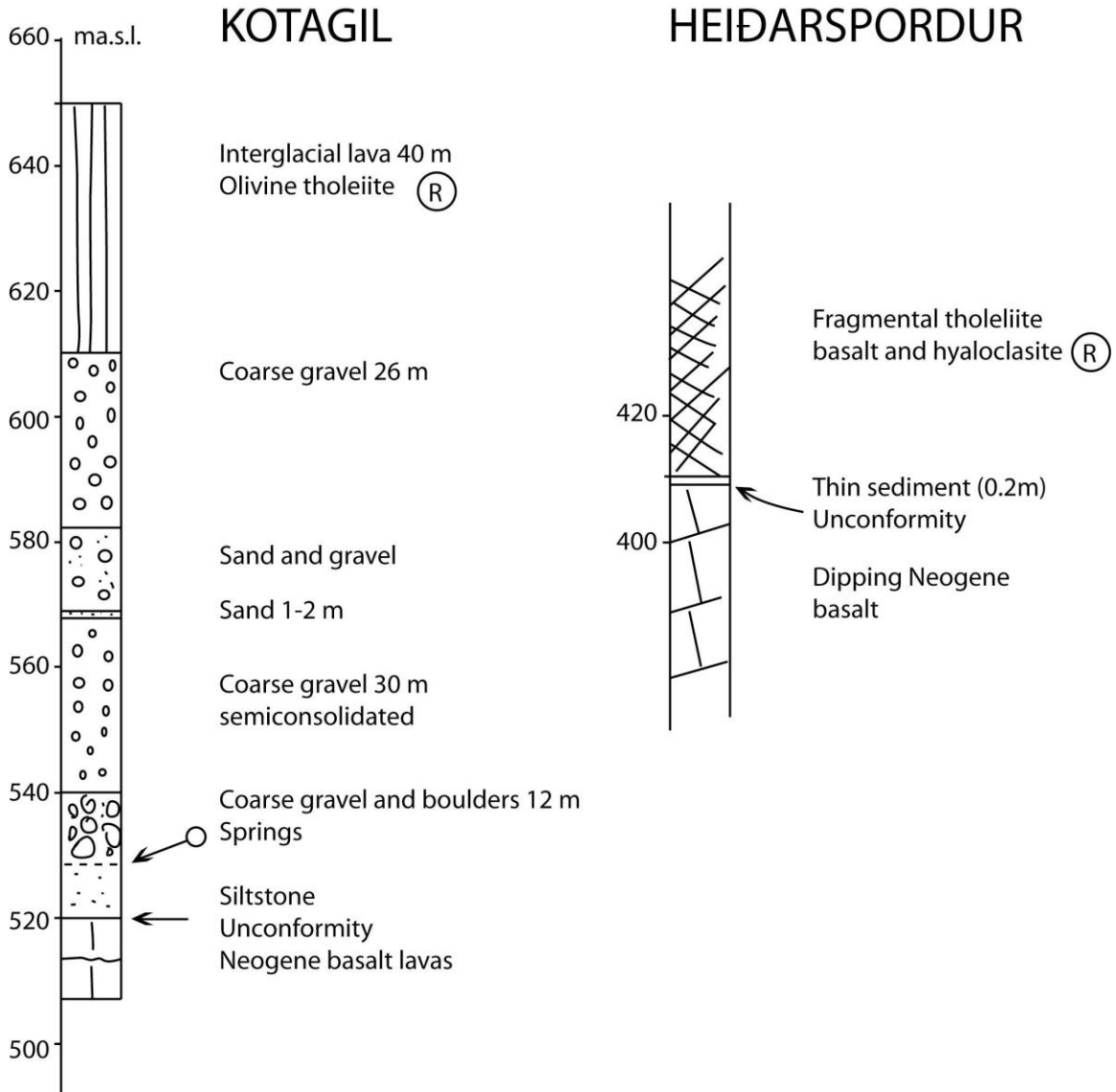
Fig. 35. The early Pleistocene formations of Norðurárdalur; Kotagil and Heiðarsporður

4.5   Skagi Peninsula

In Skagi Peninsula the sediments separate the Tertiary and Quaternary successions as they do in Skagafjörður valleys. They were first described by Péturss (1905) and later by Einarsson (1959) and Everts (1975). Everts calls them marine tuff sandstone and divides them into several series (fig. 36). At the west and north coast of Skagi he found marine molluscs in the sediments *(arctica islandica, macoma sp.).* The thickness of these layers is at least 20 m.
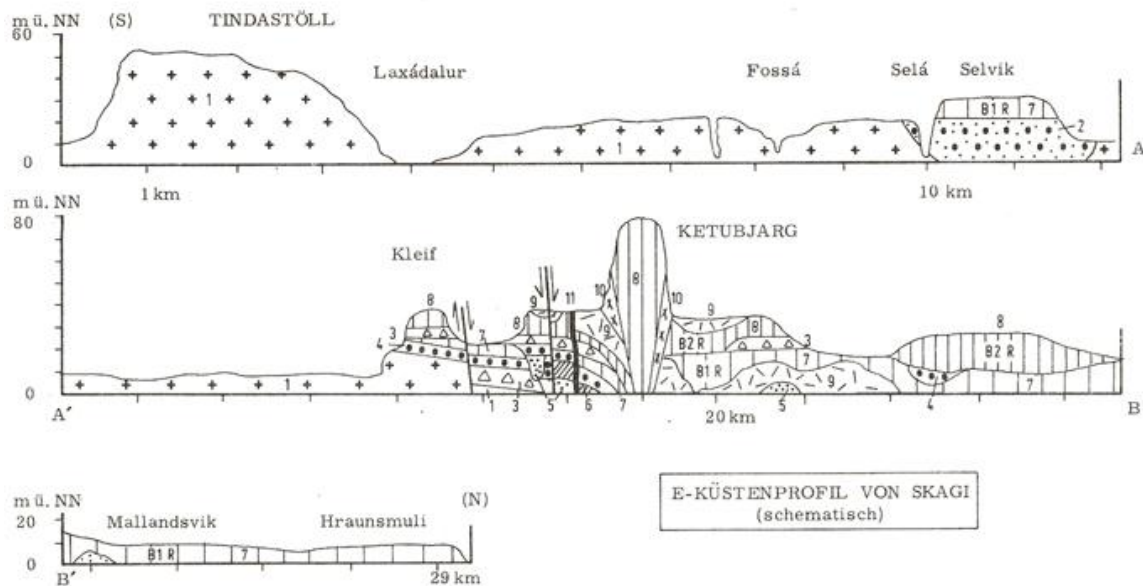
Fig. 36. Cross section along the east coast of Skagi. 1) The Tertiary volcanic succession. 2) Conglomerate. 3) Tillite 4) Conglomerate. 5) Fluvioglacial sandstone. 6-8) Quartz-tholeiite (R). 9) Hyaloclastite. 10) Breccia. 11) Basalt dyke (N). (N) = normal polarity. (R) = reverse polarity. (From Everts 1975).

4.6   General remarks

At the end of Pliocene a mature landscape had evolved in Skagafjörður district featuring the fjord itself, the principal valley, all the main tributary valleys, many of the smaller valleys and even ravines as Kotagil. Then, for an unknown reason, the valley system started to fill up again and the sedimentary succession of the unconformity was formed. This change, from digging out to filling up, doesn't seem to have been due to a great shift in the erosive energy. Rather the erosion level seems to have changed drastically. Volcanic activity near the shore or in the fjord itself or even in Skagi Bank on the4 sea bottom might be the reason, but other more general circumstances might have caused this such as a major transgression of the sea in North Iceland or in all of Iceland. The sedimentary succession of the unconformity is thickest in the valleys, tens or even hundreds of meters, but becomes thinner in the highlands and in some places it is totally

absent. The grain size is diverse but in general rather coarse. Tillite layers, glaciofluvial and fluvial sediments, lacustrine and marine layers with fossils, have been indentified.

The environment reflected by the sediment of the unconformity seems diverse. It is a time of sedimentary accumulation. In most places the sedimentary beds are a coarsening upwards sequence, indicating increasing transport ability of the Skagafjörður rivers.

Table 13. Marine fossil sites in the volcanic pile of Iceland

Place	Age Ma.	Elevation m a.s.l.
Elliðavogur series (Reykjavík – Brimnes)	0.2	0-3
Goðdalafjall	1.7-2	330-360
Skagi	1-2	0-3
Snæfellsnes series, Kirkjufell	2	180
Snæfellsnes series, Búlandshöfði – Stöð		
Snæfellsnes series, Ólafsvík – Rif		
Tjörnes series (Breiðavik Group)	1.2-2.5	
Tjörnes series	3	

The supposed marine and costal layers in Vesturárdalur are noteworthy. Sediment containing marine fossils is rarely found (table 13) and this is far the most inland and elevated fossil site reported in Iceland, located at 350 m a.s.l., 50 km from the present shoreline. It reflects a major subsidence and sea level fluctuation during Pleistocene time.

The diamictite in some of the sections has been interpreted as tillite, indicating glaciations. The first full scale glaciations with glaciers occupying the main valleys and fjords are supposed to have occurred 2.4 Ma (Geirsdóttir 1990). The central highlands of Iceland may have experienced extensive glaciations before that time. According to the Ar/Ar dates the tillites might be around 2 Ma.

The age or the time span of the sedimentary layer of the unconformity is somewhat unclear but its accumulation ceased when the lava flows of the Pleistocene volcanic succession covered the floors of the Skagafjörður valleys 1.6-1.7 Ma.

5   THE QUATERNARY SUCCESSION

Volcanism and rifting started again in the Skagafjörður Valleys in early Pleistocene after quiescence for million years. According to new Ar/Ar-dates, supported by paleomagnetic surveys, the volcanic activity initiated around 1.7 Ma or in the upper part of the Matuyama magnetic chron, between the subchrons Olduvai and Jaramillo. The Olduvai subchrone has not been recognized. The volcanism never became intensive and in the beginning the accumulation of lavas was slow. The lavas flowed along the Plio-Pleistocene valleys that at this time had been excavated at the east and west sides of the extinct Tinná Volcano. The western valley seems to have been the bigger one and there most of the lavas flowed. In between the lavas thick sedimentary layers were accumulated, including tillite layers, and hyaloclastite formations are also found. The pile indicates the cooling climate of early Pleistocene.

The succession is thickest in the highlands north of Hofsjökull where it forms a continuous cover. Towards north it becomes thinner and more scarce. In the central part of Skagafjörður District it disappears. There it has been worn out by erosion, if it existed at all. Farther north, in Skagi peninsula, the Quaternary succession appears again as a continuous coverage of interglacial lavas.

Table 24. Skagafjörður group and its formations

Formation	Age, Ma	Polarity
Hofsjökull formation	< 0.4	N
Hofsafrétt volcanic formation	0.4 – 0.8	N
Giljamúli volcanic formation	0.8 – 1.8	R/N/R
The outliers	0.8 – 1.8	R

Pleistocene lavas and eruptive sites form an elongated zone, the Skagafjörður volcanic zone. Its length from Hofsjökull to the northern tip of the Skagi Peninsula is 150 km. Its width is 50 – 60 km. The area on land is therefore near to 8250 km^2 but there is reason to belief that the Skagagrunn bank north of Skagi is a continuation of the Pleistocene subaereal basalt area (Everts, 1975). In table 15 some Pleistocene eruptive sites are listed. The thickness of the Pleistocene eruptives is very variable from one place to another, about 100 m in Skagi Peninsula in the north, thin or absent in the central part of the district and 200 – 300 m in the southernmost part, the highlands north of Hofsjökull. The geological evidences indicate low productivity of volcanic material. The maximum thickness is found where the lavas filled up the Vesturdalur valley. In the section at Bjarnastaðahlíð 12 lavas are recorded separated by sedimentary layers. There the whole section is 300 m, whereof the lavas make up 230 m.

Fig. 38. The Austurdalur Pleistocene Volcano

The Quaternary volcanic succession is regarded as one group, the Skagafjörður group, divided into four formations (table 14). These are the outliers the Giljamúli formation, the Hofsafrétt formation and the Hofsjökull formation. The stratigraphy of the group is given in more details with its formations and members in table 17.

5.1   The Outliers

The outliers are single patches of volcanic rocks, scattered over the central part of the Skagafjörður District, of uncertain age and stratigrapic order. Four eruptive units belonging to this category will be described. They are the Austurdalur Pleistocene Volcano, Mælifellshnjúkur hyaloclastite mountain, Kotagil lava patch and the Heiðarsporður hyaloclastite. All of them are remains of early Pleistocene volcanism inside the Skagafjörður valley system. The Austurdalur Pleistocene Volcano is the only member of this formation that is inside the Bedrock Map.

5.1.1   *Austurdalur Pleistocene Volcano (ep5)*

The remnants of the Austurdalur Pleistocene Volcano can be seen in rocky promontories in the east slopes of Austurdalur, on each side of Tinnárdalur mouth, at 500-600 m a.s.l. resting on thick and coarse sediments of the unconformity (fig. 38 and 39). The bottom of the lava is about 200 m over the present lava floor. The lava is porphyric with plagioclase phenocrysts in a fresh, gray groundmass. The cliffs are of regular polygonal columns, 10 – 20 m high but higher up they become less regular. The total thickness of the lava is around 50 m. It has suffered heavy erosion for over million years and experienced several glaciations so originally it must have been considerably thicker. The lava was erupted on a short fissure in the valley itself and the feeder dyke can be seen where it cuts the sediments in the Tinnárdalur mouth. It is 2 m wide with horizontal columns but increases to 10 m width at the bottom of the lava where the columns become irregular as they incorporate the surface basalt. The type of the craters was most likely a short row of cinder cones.
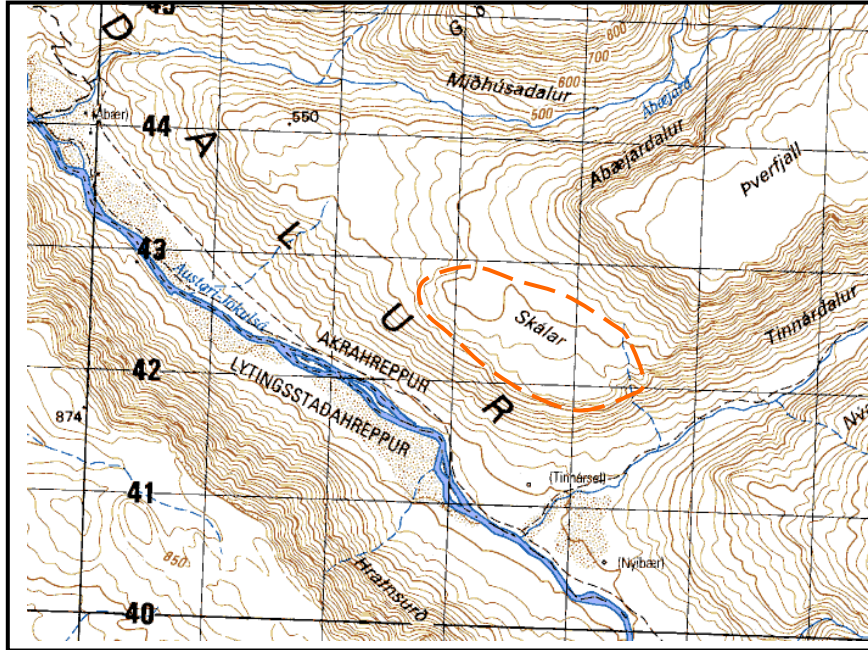
Fig. 39. Austurdalur Pleistocene Volcano is encircled with a dashed orange line. The squares are 1 km across. (With permission from NLSI)

A feeder dyke and corresponding lava is hardly anywhere as clearly exposed as here. Guðmundsson (1995), who has studied dykes and their mechanics all over Iceland, states that in the Neogene and early Pleistocene regions not a single dyke has been found that is, beyond reasonable doubt, a feeder to a lava flow. The best known example of this phenomenon in Iceland is the feeder to the crater row Sveinar – Randhólar. It is a young formation, 6 – 8 thousand years old, exposed in the deep Holocene canyon of the river Jökulsá á Fjöllum.

Fig. 40. Mælifellshnjúkur hyaloclastite peak

5.1.2 Mælifellshnjúkur hyaloclastite mountain

Mælifellshnjúkur peak is on of the most prominent landmark of Skagafjörður District and proudly decorates its blazon. It is 1138 m high, towering over the province and surrounding mountains (fig. 40 and 42). The geology of Mælifellshnjúkur was first described by Líndal (1940). He climbed the mountain and found out that the lower part of it belongs to the Neogene succession and is mainly built up of gently dipping basaltic lavas with thin interbeds. An unconformity divides the strata near the 900 m elevation and on top of it a Pleistocene formation is resting.
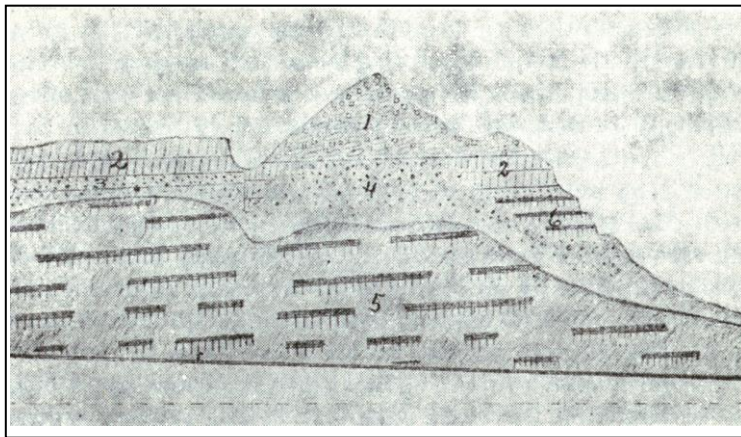
Figure 41. Mælifellshnjúkur peak. (The drawing is from Líndal 1940).

In his article from 1940 Líndal describes the topmost 200-300 m of the peak as a subglacial volcanic formation of hyaloclastite, pillow lavas and breccias cut by basaltic feeder dykes and veins. His interpretation is still valid and will be adapted here although his cross-section of the stratigraphy of Mælifellshnjúkur will be revised (fig. 41).

The rock type of Mælifellshnjúkur is olivine tholeiite with scattered small olivine phenocrysts. The rocks in the breccia, pillows and veins are rather porous with small vesicles. Its lower part seems slightly altered, opal being a prominent secondary mineral but has not been found in the upper part.
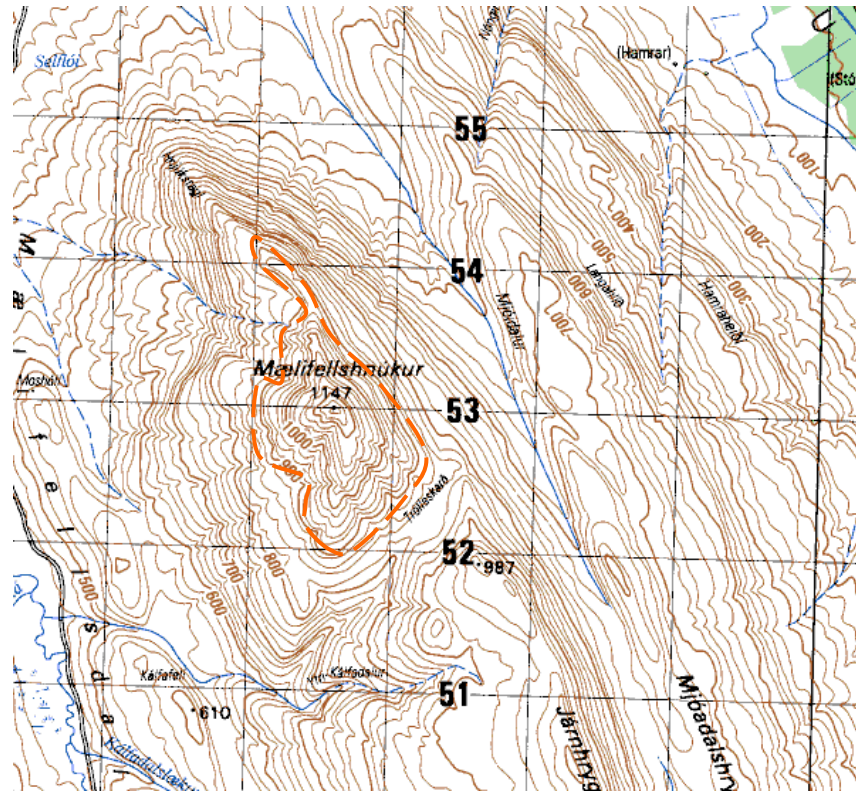
Fig. 42. Mælifellshnjúkur. The hyaloclastite is encircled with an orange line. (With permission from NLSI)

Mælifellshnjúkur is a monogenetic formation, 240 m in thickness, elongated north-south and around 2.5 km in length. This elongation might as well result from erosion as from tectonics. Subglacial volcanic origin seems very likely but is debatable. The vesicles in the rock might indicate low pressure and a rather thin ice cover but no signs of a lava cap or subaerial volcanism are found indicating that the eruption took place under severe glacial conditions when at least 1100 – 1200 m thick ice filled the Skagafjörður valleys. The state of erosion seems moderate although the peak has suffered million years of full-scale glaciation. It seems likely that during most of that time it extended as a nunatak above the glacier limits and thus escaped heavy glacial erosion, but during glacial maximums it might have been totally covered by ice.

Contrary to the cross-section of Líndal (1940) no Pleistocene lavas were found at the base of Mælifellshnjúkur peak and sediment corresponding to the sedimentary layer elsewhere at the unconformity has not been found. The contact between the Pleistocene and Neogene rocks is in most places hidden by scree.

The polarity is reverse (R) and most likely reflecting the Matuyama paleomagnetic field. No direct dates are available but here the age is proposed to be around 1 Ma.
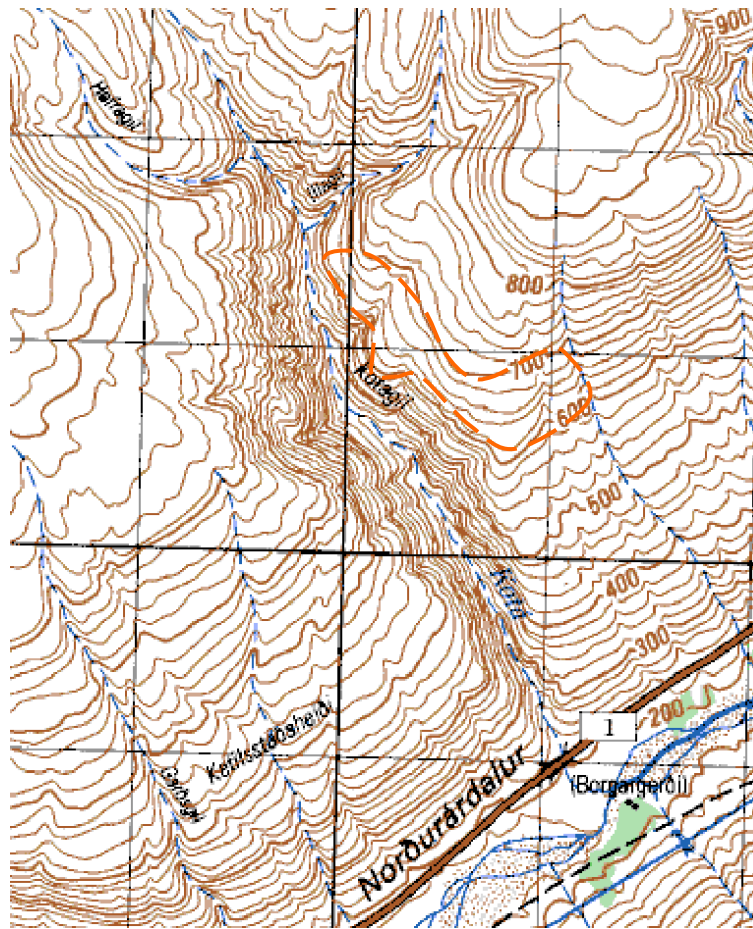
Fig. 43. Kotagil ravine. The Kotagil lava is encircled with an dashed orange line. The squares are 1 km across. (With permission from NLSI)

5.1.3 Kotagil

Kotagil ravine in Norðurárdalur valley is well known in the geological literature of Iceland since Thorarinsson (1973) described the numerous imprints of wooden trunks and branches that are found in the lavas cut by the ravine. The Pleistocene lava and underlying sediment at the ravine is much less known and has in fact never been described (fig. 43). It is shown on the geological map of Iceland (Jóhannesson and Sæmundsson 1998) but its location is rather inexact. There it is marked at the ravine Garðsgil 5 km west of Kotagil so it took a whole day to find it.

Kotagil is a large V-shaped ravine that has been cut towards NW into the 1100-1200 m high Tröllaskagi massive, perpendicular to the Norðurárdalur valley. A little rivulet, Kotá, flows along it. The ravine is narrowest near the conjunction with Norðurárdalur but becomes wider and deeper further upstream where it reaches 200-300 m depth. The

steep walls are made of Neogene basalt flows and thin read-coloured interbeds with regular layering. Dykes and faults are rare.

At 610 m, on the eastern brink of the ravine, remnant of a Pleistocene lava is found resting on 90 m of sediment. The sediment is made of stratified conglomerate with variable grain size sitting between 520 and 610 m a.s.l. (fig. 44 and 35). No diamictite is found. The layers are semi-consolidated and near their bottom small coldwater springs issue out. The sediment is partly capped by the early Pleistocene lava but extends further upstream along the ravin brink.

Figure 44. Kotagil interglacial lava resting on light-coloured Pleistocene conglomerate. Springs issue from the bottom of the sediment. The lovermost layers seen on the picture are Neogene basalt lavas with thin red-coloured interbeds.

The lava is 40 m thick, made of grey olivine tholeiite with small olivine phenochrysts (< 1%). Small plagioclase phenochrysts are also found (< 1%). This is only a patch of originally larger lava flow, the last remnant of a little volcano. It is now 800 m in length and around 100 m where it is widest but its original thickness and area remains unknown. The polarity is reverse (R). Although no feeder dyke is seen the eruptive site cannot be far away. It must have been erupted inside the primeval Kotagil.

The ravine has been deepened by some 100 m since the accumulation of the sediment. According to that the Kotagil valley and the deep ravine cut into its bottom already existed in early Pleistocene.
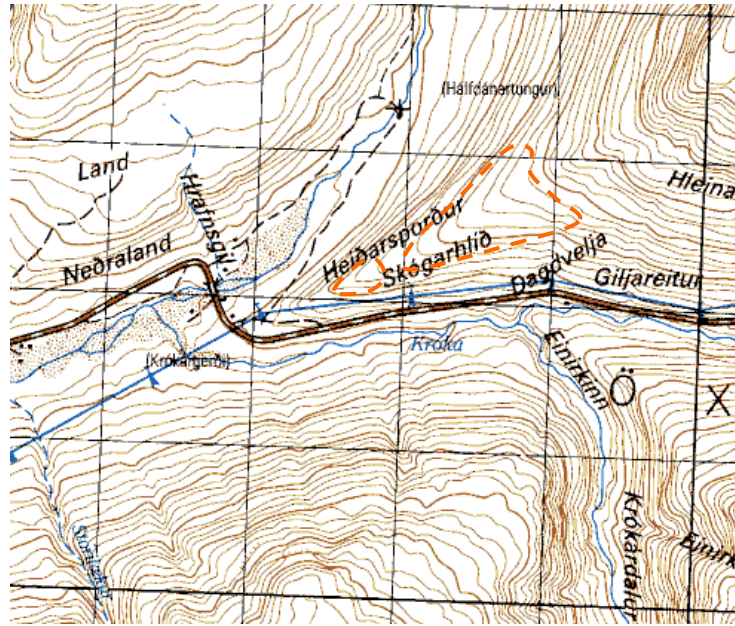
*Fig. 45. Heiðarsporður hyaloclastite. The subglacial formation is encircled with an dashed orange line. The squares are 1 km across. (With permission from NLSI)*

5.1.4 Heiðarsporður hyaloclastite

In the upper part of Norðurárdalur valley, near the junction of Norðurá and Króká rivers, a hyaloclastite unit is resting unconformably on the Neogene lava pile (fig. 45 and 35). The unit is 1500 m in length, 400-500 m where it is widest and 50-100 m thick. The rock type is fine-grained, dark grey tholeiite, without phenochrysts. In the lower part hyaloclastite tuff, pillows and pillow breccia is abundant, but higher up the layer is made of fragmentary basalt. It is fresh and no secondary minerals are found. No feeder dyke is seen but the rock is most likely erupted locally. The polarity is reverse (R).

The sediment below is very thin sand- and siltstone. In most places it is only 10-20 cm and some places it is absent at all.

5.2 Giljamúli Volcanic Formation

Giljamúli volcanic formation is resting conformably on the sedimentary succession of the unconformity (kvs). Its upper boundary marks the early/late Pleistocene divide and Brunhes/Matuyama reversal as well as the classic divide of the Younger and Older Grey Basalt Successions (or Dolerite Susseccion). This nomenclature has been used in Icelandic geology for a century. The divide can be seen on all geological maps of Iceland. In fact it is a magnetostratigraphic marker but not a conventional lithostratigraphic boundary and thus, in a strict sense, not according to the rules of correct stratigraphy but is used here because of its long tradition.

5.2.1 Jorgil lavas (ib1)

The oldest known members of the Giljamúli volcanic formation are lavas interbedded in the sediment of the unconformity. They are best studied in Goðdalafjall and Hlíðarfjall mountains. These lavas flowed along Proto-Vesturdalur, the Plio-Pleistocene valley of the district, at the western side of Tinná Volcano, covering fluvial deposits, the desolate flood plain of the river. These layers were later covered by thick beds of more fluvial sediments. One of the first eruptions issued feldspar porphyritic, compound lava (ib1) found at both sides in the lower slopes of Vesturdalur. Its thickness is 20 – 40 m. Its age is 1.6 Ma (table 21).

5.2.2 Eyfirdinga tholeiite (ep1) and Giljamúli tholeiite (ep2)

These lavas form widespread flows, extending over 40 km from the Geldingsá area, in the southeast corner of the Bedrock Map, to Goðdalafjall and Hlíðarfjall in the northwest part of the field. The lavas are of intermediate thickness with sedimentary layers in between. The sedimentary succession of the unconformity is right below them in most places. The division into two members (ep1 and ep2) might be unnecessary but is due to lack of connection between the main tholeiitic areas. These lavas flowed along Vesturdalur and added considerably to its filling. Even the lavas, that were erupted in the area now inside the catchment area of upper Austurdalur, seem to have reached Vesturdalur. That improves the statement that Vesturdalur was longer and further developed in early Pleistocene than Austurdalur. A thick tillite layer is found in between the tholeiite lavas near Fossá in Vesturdalur and in Hlíðarfjall.

5.2.3 Giljamúli olivine tholeiite (ep3)

Giljamúli olivine basalt (ep3) is a compound lava expected to be formed in a shield volcano. The eruptive site is unknown but could be near Reyðarfell, a prominent hill near the southern border of the Bedrock Map. From there the main lava lobes flowed 30 km towards north, along Vesturdalur valley and completely filled it up. After that it covered wide areas in the highland plateau on both sides of the valley. A branch from the main flow found its way to Austurdalur, south of the Tinná Volcano and flowed along it. The lava can be seen resting conformably on the coarse fluvial sediments of the unconformity in Keldudalsmúli, 50-100 m in thickness (fig. 22), remnants of which still exist as erosional leftovers on the eastern slopes of the valley at Hvítár rivers. It also has been encountered in drill hole cores.

This is the most widespread lava on the Bedrock Map but in spite of that it is sparsely investigated partly because it is to a large extent concealed by glacial ground-moraine.

Giljamúli olivine basalt is one of the oldest compound lavas in Skagafjörður Valleys. In the Neogene pile they are rare but above the unconformity such lavas become abundant. The unconformity therefore marks a change in the behavior of the volcanism. (This will be discussed in chapter 5.7).

Reference profile for the lower part of Giljamúli formation is in the Fossá canyon in Vesturdalur between 540 – 640 m a.s.l. All the lava flows of the Giljamúli Formation are of reverse polarity (R).

5.2.4   Lambá Member (lá1 – lá4)

The lavas of the Lambá member have normal polarity. They are supposed to be from the Jaramillo polarity subchron. Lavas from this subchron have been identified in a few places, 1 – 3 layers in each locality. They form a discontinuous horizon inside the Giljamúli formation and are mostly made of porphyritic and olivine tholeiite lavas.

The Jaramillo lavas at Lambá river, lá1 and lá3, are porphyritic but in between them is a compound olivine basalt layer (lá2). This is the Lambá Shield Volcano of light gray, medium-grained olivine tholeiite. The thickness of the lava is 20 m at Lambá and Fossá rivers. Towards east the normal lavas have been traced to Hraunþúfa gorge. In the deep canyon of Vestari Jökulsá, 20 km farther west, these layers have disappeared but there is one tholeiite lava instead, with normal polarity. Trausti Einarsson (1959, 1961) found by field measurements normal polarity inside the Pleistocene lava pile in Húnavatnssýsla farther west and north (in Stóradalsháls, Sólheimaháls, Höskuldastaðanúpur) that might belong to the Jaramillo lavas as well.

Lavas from Jaramillo have been mapped in several places in Iceland: In all these sites only a few lavas (1-3) comprise normal polarity. According to Shackelton et al. (1984) the lowest part of Jaramillo in the North Atlantic area coincides with rising carbonate content and increasingly negative $\delta^{18}O$ values indicating an interglacial phase but the situation soon changed and the upper part of Jaramillo is characterized by glacial conditions. In Iceland the situation seems to have been similar to this.

5.2.5   Lambárfell Members (lf1 – lf4)

In the gorge of Fossá, where it flows in cascades and rapids northwards along the slopes of Lambárfell, the river has exposed bedrock composed of lavas and sedimentary beds. The lavas have reverse polarity and are thought to date from the last reverse subchron of Matuyama, separated by coarse grained diamictite (tillite) layers. The age of these layers is therefore suggested to be 800.000 – 1.000.000 years (Hjartarson and Hafstað, 1999). Four layers are exposed:

> *Lower Lambárfell porphyritic basalt (lf1)*, 18 – 20% plagioclase phenocrysts, 10 – 20 m thick and solid.
>
> *Lower Lambárfell diamictite (lf2)*, 5 – 6 m thick with angular boulders at the base.
>
> *Upper Lambárfell porphyritic basalt (lf3),* 4% plagioclase phenocrysts, columnar jointed, 4 – 6 m thick with eroded surface. In the southwest corner of the Bedrock Map, near Skiptabakki resthouse, widespread porphyritic lavas are found, most likely a continuation of either one or both of the lavas lf1 and lf3. (Temporarily they have been marked as dd on the Bedrock Map).
>
> *Upper Lambárfell diamictite (lf4)*, 6 m thick with big boulders at the base, more fine-grained in the upper part.

This section was first described by Einarsson (1962, p. 25).

5.2.6 *Hyaloclastite hillocks*

In the highlands south of Vesturdalur three convex hillocks extend from east to west. These are Sandfell, Lambárfell and an unnamed hill near Miðhlutardrög. All of them are covered by a thick till coat so the interiors are unexposed. On geological maps they have been classified as early Pleistocene hyaloclastic hills (Kjartansson, 1965, Guðmundsson 1991 unpubl. map). If this is correct they might have been erupted subglacially during a glaciation shortly before the Brunhes/Matuyama reversal, perhaps contemporarily with the accumulation of the diamictite layer (lf4). On the Bedrock Map the hillocks do not have a special identification but are only shown as covers of tillite.

5.2.7 *Bleikáluháls (báh)*

Bleikáluháls is a major hyaloclastite unit below the lower margin of the Bedrock Map but its northernmost tail crosses its border. It seems to be of similar age as the Lambárfell Members and could have been formed subglacially contemporarily with Sandfell and Lambárfell and the diamictite layer (lf4). It is highly eroded and covered with a thick till coat but the underlying rocks are exposed here and there. It is pillow lava and breccia in brown tuff with plagioclase phenocrysts 7% by volume and small sporadic olivine crystals (<<1%). The polarity is reverse. On all geologic maps it is expected to be of late Pleistocene age but Hjartarson and Hafstað (1999) discovered its Matuyama age.

5.2.8 *Hraunþúfa compound lava (hþd)*

An excellent exposure in this member is to be found where the rivulet Ystakvísl cascades down into the Hraunþúfugil canyon. There are 4 – 5 lava bands in the vertical wall. Their thickness are in most places 2 – 3 m and up to 5 m. The total thickness in this site is 20 m. The rocks are columnar jointed. In a hand specimen the grain size is medium, the matrix is rather porous and very fresh and unaltered. The Hraunþúfugil canyon displays a superb section through the strata pile where the Hraunþúfa compound lava can be seen resting on the thick and coarse sedimentary succession of the unconformity.

5.2.9 *Hraunþúfa tholeiite (hrd)*

North of Sandafell the bedrock is covered by loose material, till and fluvial accumulation. In the till, erratics of tholeiite with scattered plagioclase phenocrysts (2 – 3%) are found and in one place the bedrock is exposed. The polarity of the tholeiite is reverse. Above it is a thin layer of sandstone and agglomerate and on top of that a normal lava from the Brunhes polarity chron.

5.2.10 Ravine tholeiite (far)

The youngest known rock with reverse polarity in the area is tholeiite lava that has filled up small ravines that have been cut into the Hraunþúfa compound lava. The Icelandic name of this member is Farvegabasalt = Ravine basalt.

5.3 Hofsafrétt Formation

Hofsafrétt formation is from Late Pleistocene and the Brunhes polarity chron. Then the volcanism was decreasing and receding to the southern part of the Skagafjörður zone. There is a continuous cover of young eruptives, shield volcanoes and lava fields from the interglacials and pillow lavas and hyaloclastite ridges from the glacial periods. In the northern part they are scattered over the district. Most of these eruptives in north and south are outside the Bedrock Map. The best known member of this formation is Drangey island. Lavas from this period have also been recognized in Skagi peninsula (Ketubruni, Tjarnarfjall) (Everts 1975). Today the active volcanism is entirely restricted to the Hofsjökull central volcano. The northern limit of the Hofsafrétt formation marks the early/late Pleistocene divide and Brunhes/Matuyama reversal as described earlier. The Hofsafrétt formation was first mapped and described, but without a name, by Hjartarson and Hafstað (1999, 2001). In the following description the strata pile is divided into monogenetic units rather than assemblages of layers as has been done in earlier chapters.

5.3.1 Lambárfell tholeiite (lþf)

The oldest lava from the Brunhes chron can be seen in one limited exposure in the Fossá canyon west of Lambárfell. This is dense, columnar tholeiite resting on the upper Lambárfell diamictite (lf4).

5.3.2 Three shield volcanoes

Lavas from three shield volcanoes are seen at the southern border of the Bedrock Map. Their summit craters were south of the map but have been eroded away or buried under younger layers. These three lava shields seem to be of similar age and from the same interglacial stage.

The Sáta compound basalt (sád) is a wide-spread olivine tholeiite lava, part of an interglacial shield volcano. The eruptive site is unknown but might have been near Ásbjarnarvötn lakes, 7 km south of the map. From there the lava has flowed down by the western slopes of Bleikáluháls hyaloclastite formation (báh) and over the northern part of Hraunþúfa compound lava (hþd). A thin layer of diamicton is found between these two lava shields north of Bleikáluháls. The minimum size of the Sáta compound lava is 75 km^2. The lava seems dense, without any phenocrysts, dark gray in hand specimen and is made of several $1 - 2$ m thick lava bands. The total thickness can nowhere be seen.

The Austurkvísl olivine tholeiite (aó) is also compound lava from a shield volcano with unknown crater site. Its southernmost tongue appears in the southeast corner of the Bedrock Map. It is composed of rather thin lava bands but differs from Sáta compound basalt in its coarse grain size and scattered plagioclase and olivine phenocrysts.

The Orravatn compound basalt (ov2) is the third lava shield at the southern border of the Bedrock Map. The eruptive craters are unknown but might have been near Laugarfellshnjúkur, 10 km south of the southeast corner of the Bedrock Map(Hjartarson and Hafstað (2001).

5.3.3   *Orravatn basalt (ov3)*

The Orravatn basalt is suggested to be the youngest lava within the Bedrock Map limits. A major part of it is however located south of its border. Its southern limits are not known but the crater, or crater row, might have been near Illviðrahnjúkar at the northeast corner of Hofsjökull. It has flowed northwards along the eastern slopes of the Bleikáluháls hyaloclastite formation. The size of the lava is at least 50 km^2. The rock type is intermediate basalt, between tholeiite and olivine tholeiite, with scattered plagioclase phenocrysts. It was first mapped and described by Hjartarson and Hafstað (2001). No sedimentary layer has been detected between the Orravatn basalt and the underlying lava shield (ov1) so it is considered to be from the same interglacial as the three compound lava shields (ov1, sád and aó). The possibility remains open that the lava belongs to the Hofsjökull volcanic system and originated somewhere below the current ice cap. More throughout mapping on the southern extension of the lava together with chemical analyses, is needed to verify that.

5.4   Till coat

The till coat (jbk) is a prominent formation in the southern part of the Bedrock Map. It is semi-consolidated sediment at the surface that hides the underlying strata. Its thickness differs from place to place and can be tens of meters where thickest. It is made of heterogeneous material. The ground mass is of sand and silt mixed with gravel, boulders and erratics. The layering is rugged and often faint. This is supposed to be mainly Weischelian till but some of it might be older. It is problematic how this unit should be mapped because it is often an intermediate stage between unconsolidated superficial deposit and semi-consolidated material. On the map it is only shown when it has a considerable thickness (> 2 m) or extension so the substratum is totally covered.

Thick layers of semi-consolidated Weischelian till manifestat fast reactions in the sediment. The reason is likely high content of volcanic glass in the material as can be expected in the neighborhood of the active Hofsjökull Central Volcano. Volcanic glass is an unstable form under atmospheric conditions and tends to convert to a crystalline state by means of hydration and devitrification resulting in consolidation of the till.

5.5   Hofsjökull

It is noteworthy that that neither tephra layers nor other eruptives from Hofsjökull Central Volcano have been found in the strata pile north of the glacier. Geological investigations in the areas south and east of the glacier have not revealed any such formations either. The eruptive history of Hofsjökull is literally unknown. The size of the volcano, a huge caldera, and prominent fissure swarms, all witness great Late Pleistocene activity. The Holocene has however been a rather tranquil period with a few flank eruptions issuing

small basaltic lavas. Three such lavas can be the seen at the northern border of the glacier, East and West Lambahraun and Illviðrahraun. Inside the caldera no eruptions have been recorded although a constant seismic activity is detectable (The Icelandic Meteorological Office, homepage). No glacier bursts or jökulhlaups have been reported. Hofsjökull Central Volcano seems to be younger than most or all volcanic formations described here above.

The classification of Hofsjökull inside the volcanic belts of Iceland has always been problematic where it towers above the surrounding area in between the Reykjanes – Langjökull volcanic zone and North Iceland zone. In many recent papers it is interpreted as an independent spreading zone, the Mid Iceland zone (i.e. Hardarson et al., 1997). Here this opinion is adapted and the Hofsjökull fissure swarm proposed to be the waning activity of the deceasing Skagafjörður rift zone.

5.6   Trausti lava – Does it exist or not

In his book, Upper Tertiary and Pleistocene Rocks in Iceland, Trausti Einarson mentions a small Holocene lava in Skagafjörður valleys where he describes relatively late and prominent fault lines. He says:

> "To these is possibly connected a faint postglacial eruption in the Svartárdalur, evidenced by a very small lava flow. The spot is on the east bank of the Svartá river 350 m above sea-level, 1.8 km south from the deserted farm Ölduhryggur. Above this place we can for a long distance follow a 8 – 10 m thick gravel terrace along the river. The topmost part of this terrace is formed of very coarse gravel, with boulders up to 1 m in diameter. The lava, hardly more than 20 – 30 m across, rests on this coarse gravel on the bank of the river. It is partly covered with river gravel. The lava is probably early postglacial in age. It appears not to have been noticed earlier" (p. 22).

This is a remote place and has rarely been observed by geologists, but in 1999 and 2001 Svartárdalur was visited with the intention to find this lava. Although far away from the active volcanic zones of Iceland and unlikely position its existence was in good agreement with ideas of the Pleistocene spreading of the Skagafjörður zone and here the last sigh of the volcanism in the area could possibly be investigated. The lava had been given the name Trausti lava in honor of the late geologist. But in spite of a thorough search the lava could not be traced. The faults that Einarsson described were in place as well as the gravel terrace and the boulders, but no lava and nothing that possibly could be mistaken for a lava. Einarsson was a trained field geologist and had studied eruptive processes and products and he had observed and written about the Hekla eruption of 1947-48. It seems out of question that he could misinterpret a Holocene lava for something else. It might have been covered by the grassland of the valley during the 40 years since Einarsson made his comments, or his location is inaccurate in some way. But until someone finds the Trausti lava again its existence remains a mystery.

Table 15. Pleistocene volcanoes and vents in the Skagafjörður Volcanic Zone

Name or place	Type	Polarity	Estimated age Ma
Orravatn basalt	Crater row?	N	0.5
Orravatn	Shield volcano	N	0.5
Austurkvísl	Shield volcano	N	0.5
Sáta	Shield volcano	N	0.5
Hraunþúfa	Shield volcano	R	1
Bleikáluháls	Hyaloclastite mountain	R	1
Geldingsá	Dyke	R	0.8 – 1.8
Austurdur Pleist. volcano	Interglacial lava and a dyke	R	0.8 – 1.8
Goðdalafjall	Dyke	R	0.8 – 1.8
Hlíðarfjall	Dyke	R	0.8 – 1.8
Kotagil	Lava	R	0.8 – 1.8
Heiðarsporður	Hyaloclastite	R	0.8 – 1.8
Mælifellshnjúkur	Hyaloclastite mountain	R	0.8 – 1.8
Gönguskörð at Tindastóll	Subglacial pillow lava	N	< 0.8
Drangey	Submarine hyaloclastite ridge	N	< 0.8
Ketubjarg	Hyaloclastite	R	1 – 2
Ketubjarg	Interglacial volcanic centre	R	1
Ketubjarg	Dyke	N	< 0.8
Þórðarhöfði	Interglacial lava and a dyke	N	< 0.8
Selfjall	Craters	R	1 – 2

Table 15 indicates, in geographical order from north to south, known Pleistocene volcanoes and volcanic vents in the Skagafjörður zone. The central part of the area, from Mælifellshnjukur peak to Gönguskörd near Tindastóll, is without any known Pleistocene eruptive centers.

Table 17. Groups, formations and members of the Quaternary succession

| Group	Formation	Member
Skagafjörður group		
	The outliers	Austurdalur Pleist. Volcano
		Mælifellshnjukur
		Kotagil lava
		Heiðarsporður hyaloclastite
	Giljamúli volcanic formation	Jorgil Lavas
		Giljamúli tholeiiter
		Giljamúli olivine tholeiite
		Lambá members
		Lambárfell members
		Bleikáluháls member
		Hraunþúfa compound lava
		Hraunþúfa Tholeiite
		Ravine tholeiite
	Hofsafrétt formation	Lambárfell tholeiite
		Sáta compound basalt
		Austurkvísl olivine tholeiite
		Orravatn compound basalt
		Orravatn basalt
	Hofsjökull formation	
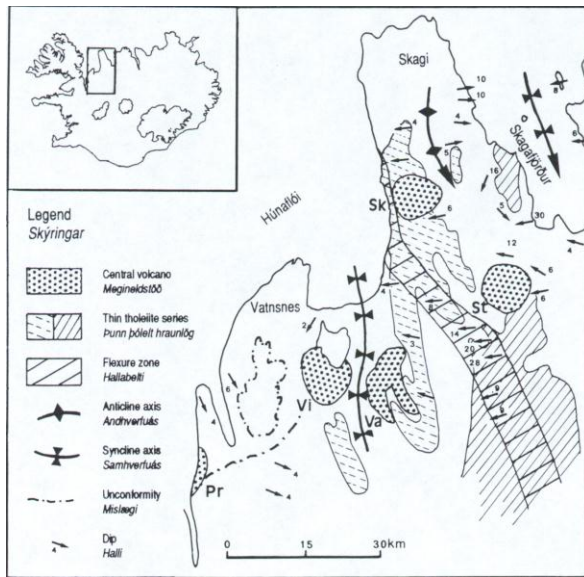
Fig. 49. Map of the main geological features and tectonics of Skagi and surroundings. (From Kristjánsson et al. 1992).

6 TECTONICS

6.1 Dip and strike

The strata of the Icelandic volcanic pile dip on a regional scale towards the central part of the country. The dips increase gradually from near zero at the highest exposed levels of the pile to about 5 - 10° at sea level. The increase in dip is matched by individual lava groups thickening down the direction of the dip. The regional tilt is thought to have been imparted to the pile during its growth, which takes place within axial rift zones that are stationary for long periods of time. Dips over 10° are exceptional and local and most often restricted to monoclinic flexure zones (Sæmundsson 1979). Shallow synclines and low anticlines have been described in the Neogene lava pile. The most prominent one is the syncline in which lies the present day axial rift zone crossing Iceland. It is formed by crustal extension and down sagging due to the load of eruptive materials. Other synclines are thought to be formed in a similar way thus indicating the location of extinct rift zones. In west Iceland the shift from the Snæfellsnes – Húnaflói rift zone created the Borgarnes anticline. The age of its flanks are strikingly different. The western flank was formed prior to 7 million years within the Snæfellsnes –Húnaflói zone but the eastern flank is younger than 7 million years and spans up to recent times inside the active Reykjanes – Langjökull rift zone. In North Iceland unconformities and monoclinic flexure zones are present at both sides of the active rift zone. To the east is the East Iceland flexure that has been traced more or less unbroken for over 200 km from Bakkaflói bay in the northeast to Suðursveit District south of Vatnajökull. In west is the Fnjóskadalur flexure reaching from Skjálfandi Bay towards south, 60 km inland, where it disappears in upper Fnjóskadalur valley (Jóhannesson and Kristjánsson 1998b). But the Borgarnes anticline might correlate to it in southwestern Iceland. At these flexures the older successions of the lava pile are down-warped and can dip 20-30° below the edge of the younger part.

The configuration of dips in Skagafjörður District is rather complicated, with the North Iceland rift zone in the east, the extinct Snæfellsnes – Húnaflói rift zone in the west, the hot spot track in the south, the Trjörnes transverse faults in the north and the temporary Skagafjörður rift zone right across. Consequently dips pointing in all cardinal directions occur.

Líndal (1964) was the first to recognize and estimate the dip and strike in the Skagafjörður valleys in the summer of 1938. His estimates are in good accordance with later measurements. The lava pile dips on a regional scale towards south with variations between SE to SW. In Norðurárdalur the dip is 5-6° S (at low altitudes) but increases southward and is between 10-15° SE at Austari Jökulsá river all the way from the Héraðsvörn confluence to Ábær. The maximal dip, 15°SE, is found in the canyon of Austari Jökulsá near Merkigil gorge. In Ábær gorge the dip is 10° SE. Further upstream it decreases. At Hvítá tributary river it is 7.5° SW and in the southernmost part of the valley the dip is 4° towards S and SE.

In Vesturdalur valley the dip of the Neogene pile is similar or slightly lower than in Austurdalur. In the canyon of Vestari-Jökulsá between the Héraðsvötn confluence and Goðdaladalur it is 12° SE. More upstream, near Hraunþúfuá tributary, it is 4° S – SE.

The maximal dip 12-15° SE in the outer Austurdalur and Vesturdalur valleys on a 5 km broad zone is here interpreted as a flexure, perhaps a continuation of the Fnjóskadalur flexure zone or a segment in the missing part between the Fnjóskadalur and Borgarnes flexures. This flexure seems related to an unconformity that represents the divide between the lava piles of the North Iceland rift zone and the older lavas of the Snæfellsnes – Húnaflói zone. This will be discussed in more details in the chapter of paleomagnetism and age (chapter 9).

In Tinnárdalur a complex pattern has been recognized. At the mouth of the valley the dip is 9° SW but a little further inland it is 10° in the opposite direction. (Local dip reaching 20° SE is found). The stratigraphy is complicated in this locality with thick layers of pyroclastics, acid lavas and fossil screes. These dips might be related to local subsidence around the voluminous rhyolite formation of the Skati dome.

The dip of the Quaternary pile above the unconformity is lower than in the Neogene pile and more regular. It is 2 – 4° S in lower Vesturdalur but in the neighborhood of Hofsjökull it approaches zero. In the central part of the Skagafjörður District the dips are around 5° W and NW, towards the Snæfellsnes – Húnaflói rift zone.

In the northernmost part of District, at the west and east shores of the fjord, a weak indication of a syncline trending N-S can be seen in the opposite dips at each side. They are 5-10° E on the western shore but 6-8° W on the eastern shore. An anticline with similar trend is found in the Neogene pile of Skagi peninsula (fig. 49, Kristjánsson et al. 1992). These tectonics might be a result of the Skagafjörður rift zone.

Surprisingly no eastward dip, resulting from the active North Iceland rift zone, is recognized but the S and SE dips in the Skagafjörður valleys point towards the volcanically active areas in the Central Highlands, where anomalous production of eruptive material has occurred in the track of the Iceland hot spotwhere a row of active central volcanoes lurk below the ice caps of Langjökull, Hofsjökull Tungnafellsjökull and Bárðarbunga.

6.2   Tectonic lineaments, faults and joints

In the Skagafjörður Valleys a number of lineaments can be clearly seen on areal photos. These lineaments are supposedly of tectonic origin. Some of them have been found and evaluated in the field but most of them are only seen on the photos. All visible faults still observed in Skagafjörður valleys are normal faults, reflecting tension in the bedrock. Most of the faults indicate downthrow less than 15 m, but bigger faults are not uncommon. In Tinnárdalur and the upper Austurdalur valleys faults with downthrow up to 120 m are found. Fault breccias often accompany the fault planes. The thickness of the fault breccia is usually less than 1 m. The fault breccia is made of crushed rock, often cemented by altered rock and secondary minerals.

Faults and fissures in the Skagafjörður Valleys can be divided into three main systems; a Neogene one, an active Pleistocene - Holocene one and a Hofsjökull fissure swarm. The main trend of all of them is N-S normal faulting. North of Skagi the transverse faults of the Tjörnes fracture zone dominate the tectonics.

Fig. 50. Geothermal springs at Vestari-Jökulsá river near Bakkakot. The hot springs are issuing out at a young fissure trending NNV. Kristján Sæmunds-son is making description in his note book.

6.2.1   Neogene faulting

The Neogene faulting is rather hard to distinguish in the landscape. The erosion has wiped out the scarps of faults and the fissures are filled with secondary minerals. It is bound to the Neogene areas and in some places it is covered by younger formations. The most prominent tectonic feature on the Bedrock Map, the Tinná caldera rim is very indistinct in the field and it does not show up on aerial photos or satellite images. No distinct caldera faults are found and their outlines are still more theoretical than spatially accurate. The Neogene faulting is supposed to have been active along with volcanism inside the ancient Snæfellsnes – Húnaflói volcanic belt, and to have faded out soon after the area drifted out of the volcanic zone.

6.2.2   Pleistocene – Holocene tectonics

As described in chapter 5 about the Pleistocene succession, spreading and volcanism took place in Skagafjörður in early Pleistocene but although it declined and died out in late Pleistocene the fracture systems, initiated by this crustal activity, are still active. The Pleistocene – Holocene tectonics is easily recognized in the area as it has often left scarps and clear-cut lineaments in the landscape. It also cuts through the whole pile, the Neogene and Quaternary successions and somewhere the loose overburden is disturbed as well. A Pleistocene fault system in the eastern slopes of Vesturdalur valley has been studied. It is composed of four normal faults trending NW - NNW. The throw is towards east, 50 –100 m altogether. Lava has flowed into one of the faults indicating that its has the same age as the lava pile and the other faults are most likely of similar age. This faulting is therefore suggested to be related to rift tectonics that occurred simultaneously with the accumulation of the Pleistocene lava pile. A pronounced bundle of faults has been mapped near the 75 km long eastern border of Skagafjörður District between Norðurárdalur valley and Haganesvík inlet (Jóhannesson and Sæmundsson, 1998b). It is easily recognized in the landscape with long and prominent escarpments indicating late Pleistocene and recent movements.

New bathymetric maps of Skagafjörður and the sea floor north off Skagi, reflect late Pleistocene and Holocene tectonic movements. In the fjord itself N-S trending faults and ridges can be seen. They seem to be connected to the Drangey Pleistocene hyaloclastite formation. (Karl Gunnarsson unpubl. data).

Weak seismic unrest indicates active tectonics in all the area between Hofsjökull glacier and the mouth Skagafjörður, witnessing the final activity of the Skagafjörður zone. Occasionally bigger earthquakes occur. The largest one in the 20th century was 5.0 on Richter scale. It occurred on the sea floor, on 7 km northwest off Drangey, July 11. 1964 (fig 51). An active N – S fault or fault system is seen in seismograms in the Skagafjörður mouth north off Málmey near the junction with the transverse fault area of the Tjörnes fracture zone. A major earthquake, 7 on Richter scale, occurred in March 1963, 30 km NE off Skagi at the western end of the Tjörnes fracture zone. It was followed up by many large aftershocks in the order of 4.0 – 5.5 on Richter (Icelandic Meteorological Office, homepage).
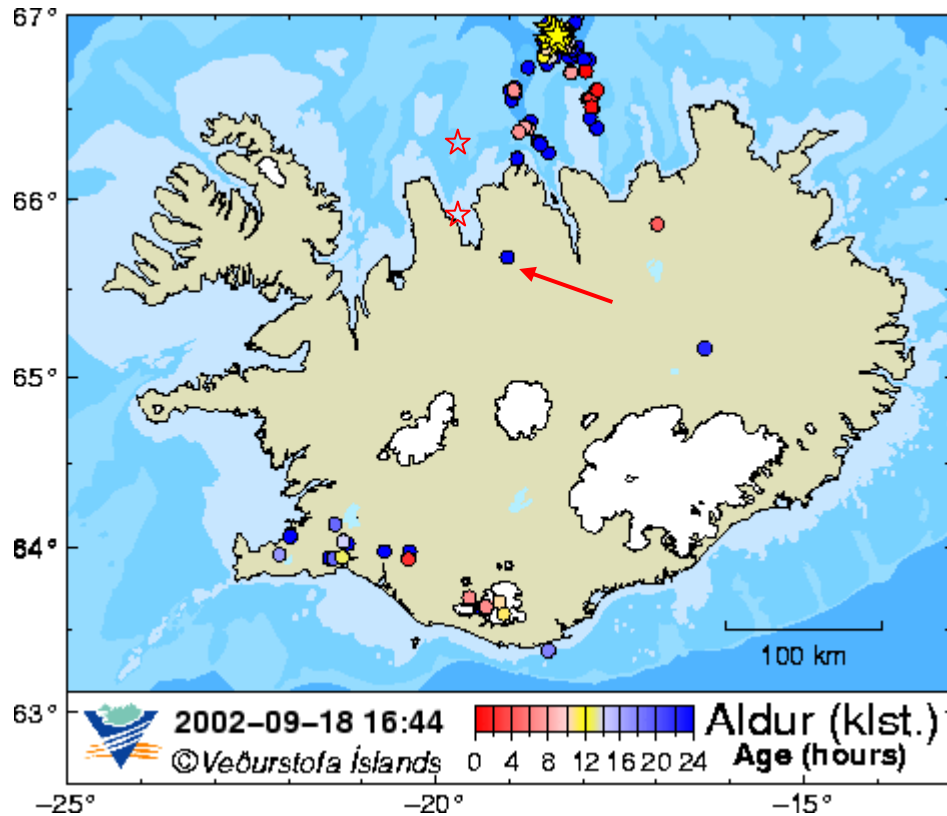
Fig. 51. In mid September 2002 an earthquake pulse occurred north of Iceland with the largest quake reaching 5.5 on Richter scale. Accompanying the main pulse small quakes were detected in Skagafjörður District. One of them appears on the picture, indicated by red arrow. The largest of those (not shown on the picture) occurred in Hjaltastaðafjall September 5th, 2.3 on Richter. Open stars indicate the major earthquakes in 1963 and 1964 (From Icelandic Meteorological Office homepage: www.vedur.is).

The Skagafjörður district is known for its geothermal fields (Karlsdóttir et al., 1991). The activity is far higher than in the adjacent districts according to both heat and utilization. Five municipal heating services operate in the district along with many small private heating centers for house heating and green houses (fig. 50). All the geothermal fields are in the Neogene bedrock but are connected to young tectonics that has broken up the strata pile forming geothermal aquifer systems. No general explanation has been introduced for this high discharge of geothermal heat in the Skagafjörður district but here it is assumed to be due to the Pleistocene rifting of the Skagafjörður volcanic zone.

The high temperatures and utilization of the geothermal fields in Skagafjörður District are a result of the still active fissure systems of the Skagafjörður rift zone.

6.2.3   The Hofsjökull fissure swarm

The Hofsjökull fissure swarm breaks up the highlands north of the Hofsjökull glacier forming long, prominent scarps and grabens. It seems to be connected to the huge ice filled caldera of the volcano. The age of the caldera is not known. Surface disruption, weak seismicity and volcanism under the glacier indicate permanent activity. The length of the fissure swarm is somewhat unclear as it grades into the Pleistocene fault system but it might be around 50 km. This fissure swarm can be seen around Skiptabakki hut in the SW corner of the Bedrock Map. The fault system seems to be young and permeable and cold springs are connected to it. The main trend is N-NNW (Karlsdóttir et al., 1991, Vilmundardóttir et al. 1997).

The Pleistocene – Holocene tectonics in Skagafjörður District and the Hofsjökull fissure swarm are in fact a manifestation of the same phenomena, that is the Skagafjörður axial rift zone. Although it is fading out it is not quite extinct, but the spreading is presumably slowly approaching zero.

7   DYKES AND INTRUSIONS

Structurally there are two main types of dykes in Iceland; regional dykes and inclined cone sheets (cone dykes) (Guðmundsson 1995). The regional dykes often form swarms that are connected to certain central volcanoes. Most dykes were originally vertical or steeply dipping and are rectangular to the layering of the volcanic pile. They are most often horizontally columnar. The dykes can be singular, formed of one columnar row or multiple, formed of two or more columnar rows. The swarms are elongate, often tens of kilometers long and several kilometers wide. Their dimensions are comparable to those of the active volcanic systems of the country. The thickness of the regional dykes is varying from few centimeters to 60 m. The most common thickness is 1 – 3 m.

The cone sheets are confined to central volcanoes. They form swarms that are circular or elliptical, several km in radius, that encircle shallow source magma chamber. Most sheets dip towards the center of the chamber. The sheet thickness is commonly less than half a meter, much less than that of regional dykes, ranging from a few centimeters to 14 m (Guðmundsson 2000).

The Bedrock Map of Skagafjörður Valleys does not give right indication of the distribution of dykes in the area. According to the map they are most common in river canyons and gorges but that is because there they are best visible and mappable but outside these exposures they are mostly hidden an cannot be traced without expensive geophysical equipments and methods. In this work only thickness, direction and some times inclination of dykes have been measured. Their rocktype or paleomagnetism have not been studied. Dykes have been carefully mapped in Austari-Jökulsá canyon from the river junction upstream to the Skatastaðir farm. Farther inland rather few exposures exist at the main river.

7.1.1   *Regional dykes*

As a general rule the dyke intensity becomes higher with increased depth. They are rare at highest exposed levels but can occupy 10 – 20% of the rock volume in dyke swarms at low levels. The highest intensity in Skagafjörður valleys is found in Ábær gorge, 9 dykes in 0.5 km interval or 7% of the total rock volume. In two km interval in Austari Jökulsá canyon up- and downstream of Merkigil gorge 25 dykes have been identified that make up 5% of the total rock volume. This rather low dyke intensity indicates rather low state of erosion and show that the pile never has entered deep into the crust. This supports the same conclusion drawn from the state of alteration of the rock. Dyke swarms as often associate magma chambers and central volcanoes in Iceland have not been recognized in Skagafjörður Valleys. Occasionally a dyke is seen connected to a certain lava layer. The only example of this in Skagafjörður valleys is the feeder dyke belonging to the Austurdalur Pleistocene Volcano already described in the chapter about the volcano. The minority of dykes is however supposed to be such feeder dykes. Most of them never have reached the surface and are therefore non-feeders.

In Skagafjörður valleys the most common dykes are single dykes of one columnar row with thickness ranging between 1 – 5 m. But they also can be multiple, made of two or more columnar rows and reach thickness up to 40 m. The average thickness is around 4 m. An acid dyke is found in Vesturdalur near the caldera rim. It can be traced for 2 km, trending ESE, cutting the canyons of Fossá and Lambá tributaries. It is 10 m thick and leaning 45° like a cone sheet. Composite dykes have not been recognized. A certain dyke can seldom be traced for long distances most often they disappear after few tens of meters. The longest mapped dyke crosses Austurdalur in Merkigil gorge. It can be followed for about 3 km. It is 12-14 m thick where thickest, trending NE.

The main trend of the regional dykes is towards NE but other directions such as ENE, W-E, and NW also occur. Dykes trending N-S are rare. This differs from what has been described for East Iceland and the Tröllaskagi Region.

7.1.2 Cone sheets

Cone sheets (cone dykes) from the final stage of the Tinná Central Volcano are also found in the upper Vesturdalur valley. They could indicate a caldera depression in the area but they have not been closely investigated.

7.1.3 Plugs and bosses

Other intrusions than the dykes are extremely rare. The roots of the Tinná Central Volcano are still unveiled and the erosion has not excavated the plutonic rocks of the old magma chamber. In the north end of Hlíðarfjall is an intrusion of fine grained basalt. The contacts with the country rocks are tuffaceous and brecciated. The total thickness is around 40 m. It is poorly exposed and its extension is unknown. In Giljadalur valley is a minor intrusion or a very thick dyke made of fine-grained rock. Thin sill is found associated with the thick sedimentary layer at the unconformity in the canyon walls of Lambá river in the upper Vesturdalur valley. More intrusions have not been identified. The common reason for the lack of intrusions and dyke swarms is that the volcano is so moderately eroded. Its roots are sitting deep below the valleys.

8   CHEMICAL ANALYSES

In geological mapping projects it often can be difficult to determine certain rock types in hand specimens in the field. Basaltic andesite and andesite e.g. can be hard to distinguish from regional tholeiite lavas and the borderline between andesite and dacite is not so obvious either. The same problem can arise in discerning tholeiite from olivine basalt because of gradation between them. It is therefore a good rule to take a few rock samples to analyze chemically to check out the field determinations.

During the field work in Skagafjörður valleys a total of 42 samples were collected and analyzed, using XRF technique, at Geochemical Laboratories, Earth and Planetary sciences, McGill University in Montreal, Canada. All major elements were analyzed and several trace elements. The samples were believed to cover the entire row of the eruptive rock types both according to petrology and age (S-1 – S-42 in tables 18, 19 and 20). The proposed extensive areas of intermediate rocks that had been found inside the Tinná Central Volcano were of special interest, and therefore many samples were taken from the expected intermediate lavas. The analyses improved the existence of the andesitic formations. In general there was a good agreement between the field determinations and the lab analyses. Locations of samples are shown on the included Bedrock Map.

Samples S-34 – S-42 were collected for the Ar/Ar dating project. They are mainly from the Pleistocene succession.

Fig. 52 shows an alkali - silica plot of the rock sample the analyses according to international standards (IUGS) (Le Maitre et al. 1989, fig. B.13).

The main conclusion of the chemical analysis is that basic, andesitic and acidic rocks from the Tinná Central Volcano follow a tholeiite trend. That is the same trend as has been found for the Thingmúli Central Volcano in East Iceland (Carmichael 1964) and often is used for comparison. The main rhyolite formation of the Tinná Central Volcano is very high in silica, or up to 75% $SiO_2.$

Another conclusion is that the Pleistocene rocks all fall within the basaltic field of the diagram, and all of them belong to low or medium K series as do the Neogene samples (fig. 53, Le Maitre et al. 1989). No alkali basalt is found.

Everts (1975) presented 25 analyses of Pleistocene basalts from Skagi Peninsula and from the east coast of Skagafjörður. Sigurðsson et al. 1978 published 31 analyses of Pleistocene basalds from the area between Langjökull and Skagi Peninsula. All these samples are tholeiites of the same kind as in the Skagafjörður valleys.

Alkalibasalt characterises the recent off-rift zones of Iceland, as the Snæfellsnes zone, the Eastern volcanic zone and the Snæfell zone. Outside these zones it is extremely rare. Its absence among the young Skagafjörður rocks is yet another conformation on the rifting of the Skagafjörður area.

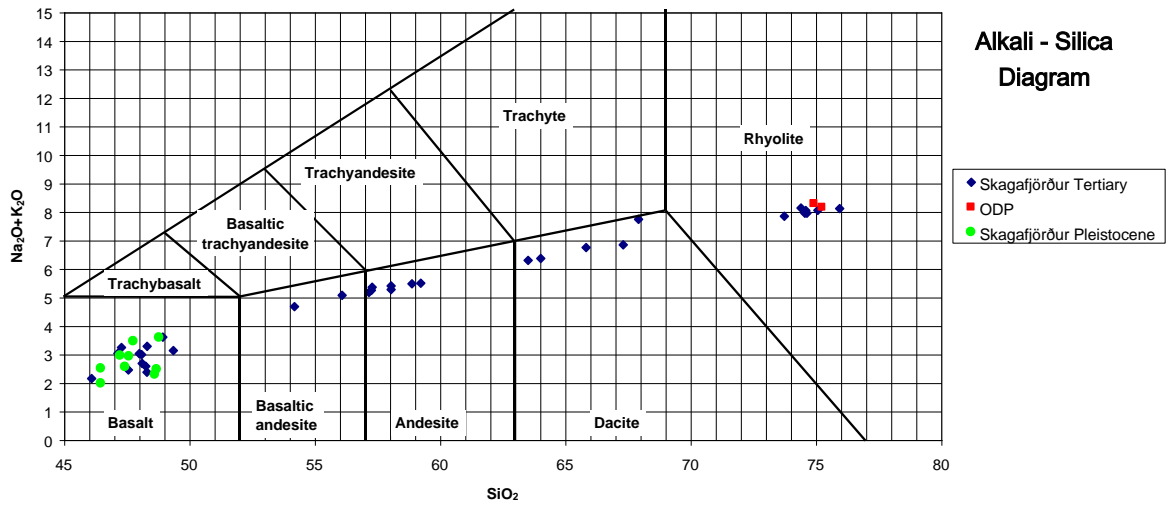Fig. 52 Alkali – silica diagram $Na_2O + K_2O / SiO_2$. The data is from all the analyses listed in table 18 and ODP analyses from Lacasse and Garbe-Schönberg (2001).
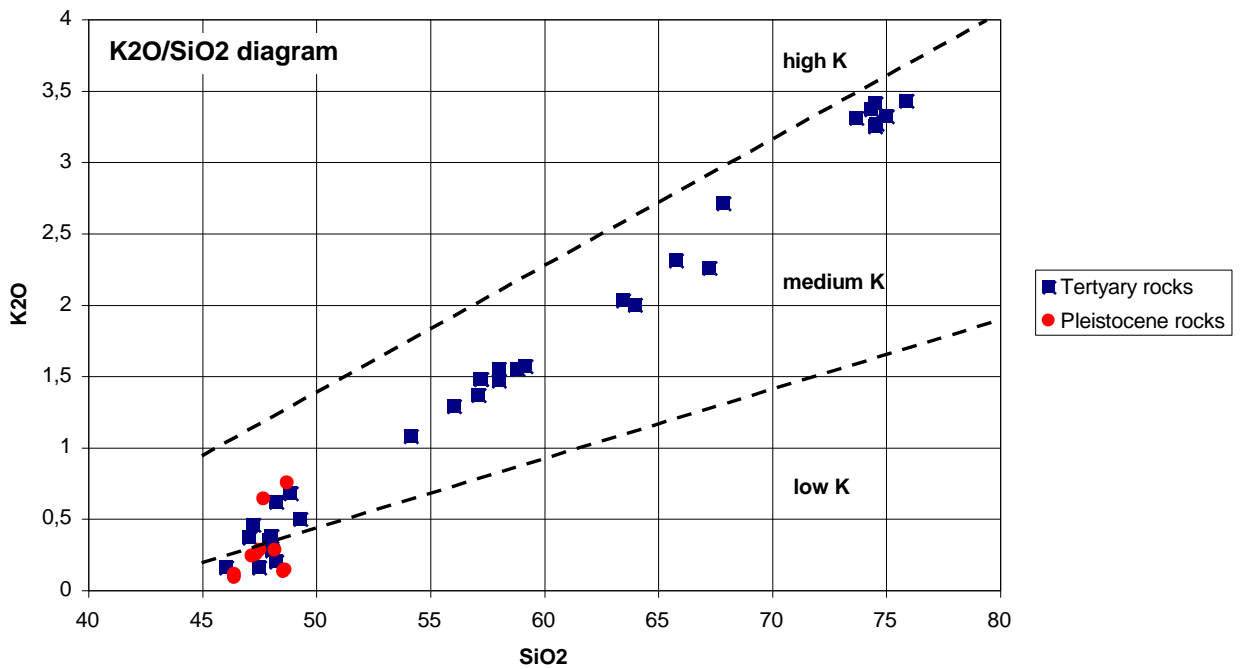

Fig. 53. K_2O/SiO_2 diagram. The data is from all the analyses listed in table 18.

Table 18. Chemical analyses, list of samples

No.	Location	Rock type	m a.s.l.	Formation
S-1	Ytri-Hvítá	andesite	400	ti-4
S-2	Ytri-Hvítá	andesite	420	ti-4
S-3	Ytri-Hvítá	tholeiite	530	nb-1
S-4	Ytri-Hvítá	tholeiite	585	nb-1
S-5	Ytri-Hvítá	tholeiite	590	nb-1
S-6	Ytri-Hvítá	tholeiite	620	nb-1
S-7	Ytri-Hvítá	tholeiite	765	nb-1
S-8	Ytri-Hvítá	porphyritic basalt	950	fo-1
S-9	Ytri-Hvítá	dacite lava	955	nb-3
S-10	Ytri-Hvítá	dacite lava	955	nb-3
S-11	Hvítárdalir dome	rhyolite	750	ti-3
S-12	Tinnárdalur	olivine tholeiite	480	ab-1
S-13	Tinnárdalur	tholeiite	500	ab-1
S-14	Tinnárdalur	porphyritic basalt		ab-1
S-15	Tinnárdalur	olivine tholeiite	550	ti-2
S-16	Skati dome, base, Tinnárdalur	rhyolite, obsidian	650	ti-3
S-17	Skati dome, Tinnárdalur	rhyolite lava	665	ti-3
S-18	Nýjabæjarfjall	andesite	980	nb-2
S-19	Nýjabæjarfjall	andesite	980	nb-2
S-20	Nýjabæjarfjall	andesite	980	nb-2
S-21	Nýjabæjarfjall	porphyritic basalt	1020	fo-1
S-22	Nýjabæjarfjall	andesite	1020	nb-2
S-23	Skati dome, Geldingaskarð	rhyolite lava	745	ti-3
S-24	Austurdalur Pleistocene Volcano	porphyritic basalt	620	ep-5
S-25	Keldudalur dome	dacite	450	
S-26	Afréttarfjall	basaltic andesite	735	nb-2
S-27	Afréttarfjall	dacite	700	nb-2
S-28	Afréttarfjall	dacite	770	nb-2
S-29	Afréttarfjall	basaltic andesite	670	nb-2
S-30	Sandafjall	andesite	925	nb-2
S-31	Skati dome, base, Geldingaskarð,	rhyolite, obsidian	550	ti-3
S-32	Torfufell, Eyjafjörður, Hafrárdalur	rhyolite		
S-33	Torfufell, Eyjafjörður, Hafrárdalur	rhyolite		
S-34	Djúpugil, Goðdalafjall	tholeiite	386	ep2
S-35	Jórgil, Goðdalafjall	porphyritic basalt	370	dd
S-36	Dalakot, Goðdalafjall	porphyritic basalt	345	dd
S-37	Bjarnastaðahlíð, Hlíðarfjall	porphyritic basalt	335	dd
S-38	Bjarnastaðahlíð, Hlíðarfjall	olivine tholeiite	580	ep3
S-39	Goðdalakista	tholeiite	585	eep2
S-40	Austari-Jökulsá near Geldingsá	porphyritic basalt	600	ge2
S-40	Austari-Jökulsá near Geldingsá	porphyritic basalt	620	ge2
S-42	Austari-Jökulsá near Stórihvammur	tholeiite	660	ep1

Table 19: CHEMICAL ANALYSES. Major elements

Table 20: CHEMICAL ANALYSES. Trace elements

xxx

9   PALEOMAGNETISM AND THE AGE OF THE STRATA PILE

9.1   K/Ar and Ar/Ar dates

Already when the pioneer geologist Helgi Pjeturss discovered the young formations in Skagafjord district in the beginning of the 20th century he proposed them to be of Pleistocene age. Later observations have not changed this estimate. It was confirmed by the geological investigations and paleomagnetic work of Trausti Einarsson (1959) and still further strengthened by K/Ar dates (Everts et al. 1972, Everts 1975). Evert's samples were collected in Skagi Peninsula and in young formation at the east coast of Skagafjörður. The dates give the time interval 0.5 - 2.6 Ma, but the error limits for the older dates are high (table 21).

Table 21. Ar/Ar and K/Ar dates on lava flows from Skagafjörður valleys

Locality (m a.s.l.)	Rock type	Polarity	No.	Ar/Ar-age	Ref.*
Sólheimafjall PF-48	Ol. tholeiite	N	78-1005	8.88 ± 0.12	1)
Sólheimafjall PF-43	Tholeiite	N	78-1002	8.72 ± 0.10	1)
Sólheimafjall PF-39	Tholeiite	A	75-144	9.04 ± 0.18	1)
Sólheimafjall PF-36	Tholeiite	A	75-141	9.12 ± 0.12	1)
Sólheimafjall PF-34	Porph. basalt	R	75-140	9.16 ± 0.20	1)
Króksbjarg		R	1	1.4 ± 0.4	2)
Tjarnarfjall		R	2	1.1 ± 0.5	2)
Tjarnarfjall		N	3	0.7 ± 0.4	2)
Ketubjörg		R	4	2.6 ± 0.8	2)
Hofsós		N	5	2.7 ± 1.2	2)
Djúpagil innra, 386 m (ib1)	Tholeiite	R	17160	1.482 ± 0.060	3)
Hlídarfjall, 335 m (ep2)	Porph. basalt	R	17169	1.468 ± 0.017	3)
Goðdalakista, 585 m	Tholeiite	R	17172	1.256 ± 0.016	3)
Austari Jökulsá, 600 m (ep2)	Porph. basalt	N	17173	2.841 ± 0.043	3)
Austari Jökulsá, 620 m	Porph. basalt	N	17175	2.486 ± 0.068	3)
Austari Jökulsá, 660 m (ep2)	Tholeiite	R	17177	1.659 ± 0.024	3)
Tinnárdalur, 650 m	Rhyolite	R	17181	5.212 ± 0.016	3)

* 1) Sæmundsson et al. 1980. 2) Everts 1975. 3) This paper.

Sæmundsson et al. (1980) published a series of dates in their extensive geological and paleomagnetic study in Tröllaskagi, North Iceland. Among them are several dates from the northern part of Skagafjörður District indicating an intensive and continuous build up of the strata pile between 9.7 – 8.7 Ma. Their uppermost dated samples were collected

around a polarity reversal in Sólheimafjall that has been correlated to the boundary between the Sólheimar formation and Merkidalur formation in Austurdalur. That is near to the bottom of the pile described in this paper. The dates are given in table 21.
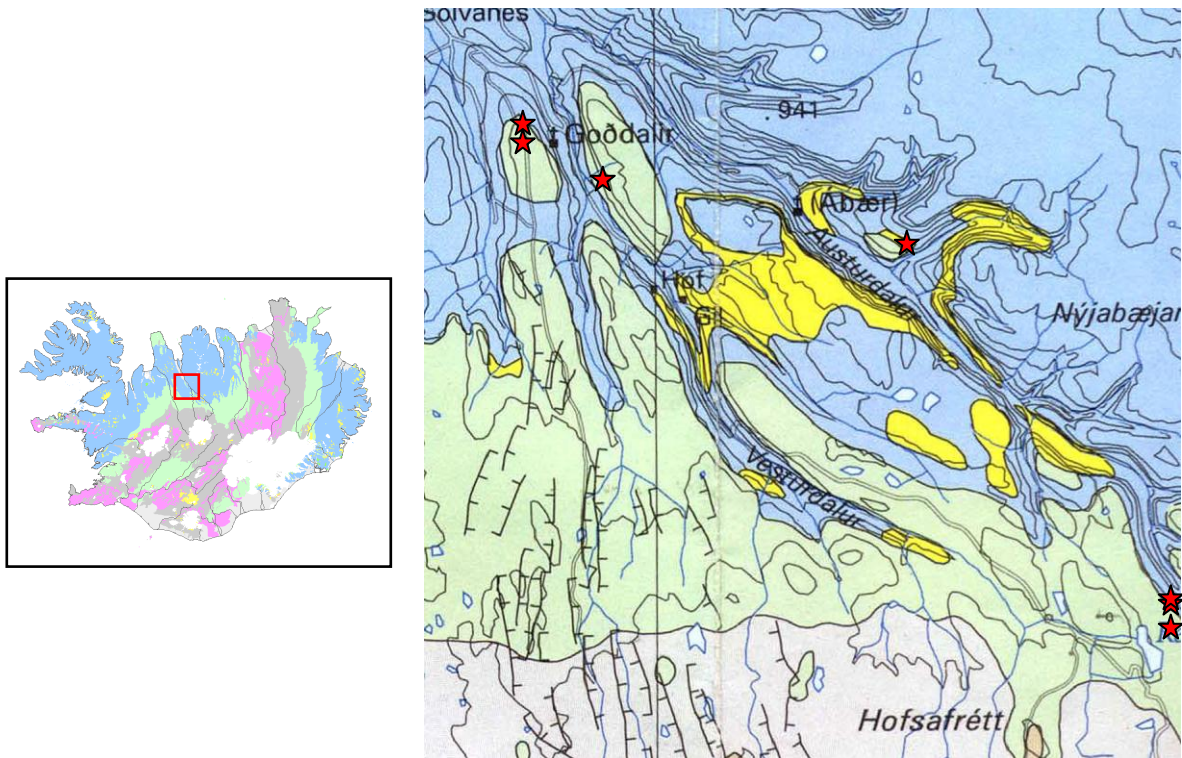

Fig. 54. Location of the dated samples. Blue = The Neogene regions. Green = The Plio-Pleistocene areas (3.3-0.8 Ma). Gray = Late Pleistocene. Yellow = rhyolites of the Tinná Central Volcanoes. (Modified from Jóhannesson and Sæmundsson 1998).

In the year 2000 cooperation was established with Björn S. Hardarson at the University of Edinburgh in order to get some new Ar/Ar dates from the Skagafjörður Valleys. The project was supported by Rannís (Iceland Research Council), BHM (Association of Academics) and Hagþenkir. The aim was to find the timing of the Skagafjörður unconformity, or more exact, to define the age gap included in the unconformity and the sediments accompanying it. Samples were collected in the summer of 2001 and prepared and measured at the University of Edinburgh in 2002. Seven samples were measured, four right above the unconformity, two right below it and one from the Skati rhyolite dome (fig. 54). The distribution of the samples should reveal the supposed widening of the age gap from south to north, from the highlands near Hofsjökull and northwards to the lower Vesturdalur valley. Surprisingly they seem to reveal two unconformities instead of one. The results are given in table 21 and descriptions of the dated samples are in chapter 9.1.1.

The dates from above the unconformity (no. 17160, 17169, 17172 17177) indicate that the activity of the Skagafjörður volcanic zone started in early Pleistocene, 1.6 – 1.7 Ma. Slightly higher ages were obtained from the oldest lavas right above the unconformity in the southernmost exposure in near Geldingsá as in the northernmost exposure in

Vesturdalur near Goðdalir, 40 km away (fig. 55). The dated samples have reverse polarity and harmony with the mid Matuyama geomagnetic field prevailing 1.6 Ma.

The dates above the unconformity gave results close to what was expected. The dates from below the unconformity were on the other hand a surprise.

The dates from Austari Jökulsá near the confluence with Geldingsá (no. 17173 and 17175) right below the unconformity give 2.49 and 2.84 Ma respectively (fig 55). This was in fact quite younger than expected. At that time the Snæfellsnes – Húnaflói volcanic zone had been extinct for several millions of years and the Skagafjörður volcanic zone had not been initiated. This means that the lavas, from which the samples were taken, most probably belong to the North Iceland volcanic zone. It might seem located rather far away from the zone. The axis of the zone is 70 km east of the site today but 2.6 Ma it was 40 km east of it (using 34 km/Ma spreading rate). The volcanic belt is over 50 km wide in Central Iceland (25 km at each side of the axis). The lava, where the sample is collected, is therefore supposed to have flowed 15 km towards west, away from the volcanic belt – and that is quite reasonable.

The lower part of the Skagafjörður strata pile originates in the extinct Snæfellsnes – Húnaflói volcanic belt (Sæmundsson et al. 1980). Somewhere between the upper and lower parts of the pile an unconformity or hiatus should be expected, reflecting the shift, or rift jump, from west towards the North Iceland volcanic belt. This unconformity has not been recognized in the field.
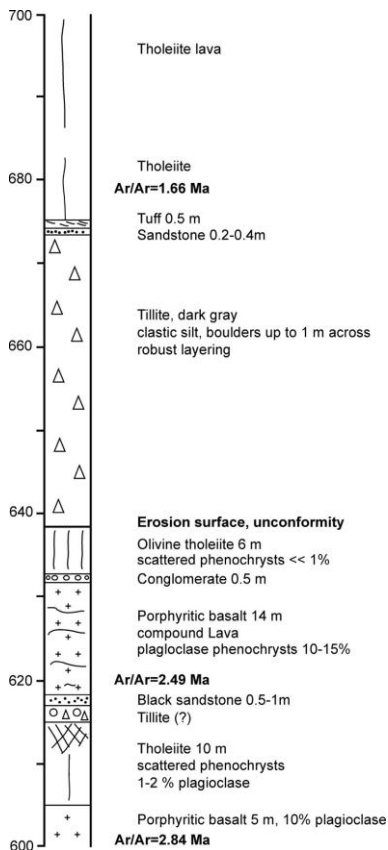
Fig. 55. A section across the unconformity at Austari Jökulsá river near Geldingsá north of Hofsjökull. Samples for dating were collected above and below the suspected unconformity. Here the unconformity includes 0.8 million years.

The date from the Skati Dome (no. 17181) gives 5.2 Ma. This was also quite younger age than expected. Based on the stratigraphy and paleomagnetic study the Skati Dome and the Tinná Central Volcano were suggested to be 7-8 Ma and to have originated inside the Snæfellsnes-Húnaflói volcanic belt (Hjartarson et al. 1997). No hiatuses, or considerable gaps, were supposed to be hidden inside the strata pile below it and no, or very few, magnetic subchrons were believed to be missing between the well-dated magnetic reversal at top of the Sólheimar formation and the Tinná Volcano. Taking this as valid the date must be considered wrong.
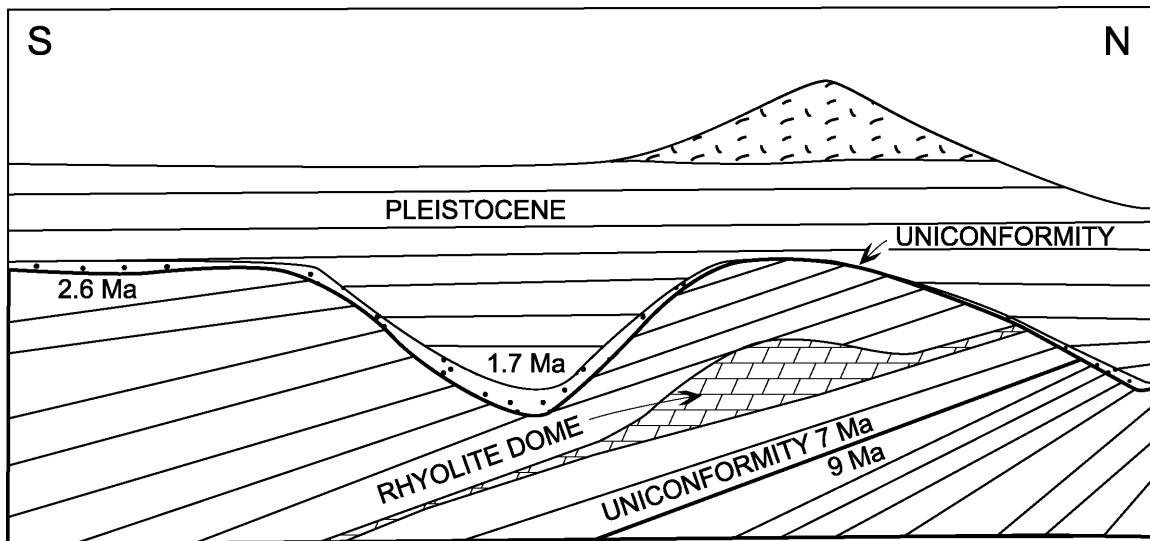

Fig. 56. A schematic cross section indicating the two unconformities in the strata pile of Skagafjörður Valleys. Horizontal lines indicate the Pleistocene succession (with a hyaloclastite hill on top). Skagafjörður Unconformity outlines an ancient valley system in the middle. The dipping Neogene succession is at the bottom including Skati Rhyolite dome and the Hidden Unconformity.

If the date however is correct, the above-mentioned criteria must be reviewed and the young ages on the dates no. 17173 and 17175 support that view. Then the Tinná Central Volcano seems to originate in the North Iceland volcanic zone because 5.2 Ma the Snæfellsnes – Húnaflói zone was extinct (or just about) and the Tinná Volcano was located 80 km east of its axis. The three dates (no. 17173, 17175 and 17181) therefore suggest a major unconformity lurking somewhere inside the pile indicating the momentous rift jump from the Snæfellsnes-Húnaflói zone to the North Iceland zone. It would correlate with the major unconformities of Fnjóskadalur and of the Borgarnes anticline (Sæmundssin 1979). But where is it and how should it be recognized? It should have; 1) erosional surface, 2) sedimentary horizon, 3) accompanying flexure zone and 4) preferably different dips at each side of the unconformity. Two possibilities exist fulfilling three the first requirements, but not the last one:

a) The Merkidalur sedimentary layer

b) The Tinná lignite sediment

New radiometric dates are required to corroborate the existence of the unconformity, its exact location and its duration, but in spite of that an independent opinion will be put forth in this thesis. The unconformity exists and the Merkidalur sediment seems to be a better candidate. Acid rock formations, suggested to be a part of the Tinná Central Volcano, are found below the Tinná lignite sediment and steep dips found in outer Skagafjörður valleys, perhaps representing a flexure, fit better to the Merkidalur than Tinná sediment. It will therefore be stated that that a hereto undiscovered segment of the unconformity between the axial rift zones of Snæfellsnes-Húnaflói and North Iceland, and accompanying flexure zone, crosses Skagafjörður Valleys near Merkigil with E – W trend. But there is some uncertainty about this, and as long as the unconformity has not been detected directly and defined in the field it will be called the Hidden Unconformity (fig 56).

It must be pointed out that there is a discrepancy between the Ar/Ar-age of the Skati Dome and its polarity. The polarity of the rock is reverse (R) but 5.21 Ma the Earths magnetic field was normal, representing the Þverá subchron, the lowest normal interval of the Gilbert polarity chron. Before the Þverá subchron was rather a long reverse subchron, representing the lowest part of Gilbert, 5.23 – 5.89 Ma. Here, the Skati Dome is supposed to belong to the central or the lower part of the subchron and its Ar/Ar-age is therefore slightly too low. Suggested age would be 5.5 Ma.

9.1.1   Description of the dated samples

Sample no. 17160. Locality N 65°19.841 W 19°06.764 386 m a.s.l. It was collected in Djúpugil near Goðdalir farm from the bottom of a tholeiite lava layer. It is columnar jointed, fresh basalt, fine grained and without all phenochrysts. Polarity R. Chemistry: see sample S-34, table 19.

Sample no. 17169. Locality N 65°18.730 W190°3.540 335 m a.s.l. It was collected from the lower part of a compound lava in a gorge in Hlíðarfjall mountain, near Bjarnastaðahlíð farm. This is a 20 m thick, porphyritic lava layer of olivine tholeiite with plagioclase phenoschysts. Polarity R. Chemistry: see sample S-37 table 19.

Sample no. 17172. Locality N65°19.280' W19°08.054', 585 m a.s.l. It was collected from the topmost of Goðdalakista mountain. Polarity R. Chemistry: see sample S-41 table 19

Sample no. 17173. Locality N65°07.462' V18°27.610', 600 m a.s.l. It was collected from the top of a plagioclase porphiric lava in the canyon of Austari-Jökulsá north of the confluence with Geldingsá tributary river. Polarity N. Chemistry: see sample S-40 table 19.

Sample no. 17175. Locality N 65°07.462 W 18°27.610 620 m a.s.l. It was collected from the bottom of a compound lava below the suspected unconformity in the canyon of Austari-Jökulsá north of the confluence with Geldingsá tributary river. This is a 20 m thick, porphyritic compound with plagioclase phenoschysts. Polarity N. Chemistry: see sample S-41 table 19.

Sample no. 17177. Locality N 65°07.208 W 18°27.238 660 m a.s.l. . It was collected from the bottom of a tholeiite lava above the suspected unconformity in the canyon of Austari-Jökulsá north of the confluence with Geldingsá tributary river. Polarity R. Chemistry: see sample S-42 table 19.

Sample no. 17181. Locality N 65°17.243 18°44.291 650 m a.s.l. It was collected from the bottom of the Skati lava dome. This is an obsidian fragment, black and shiny. Polarity R. Chemistry: see sample S-16 able 19.

Paleomagnetism

Geomagnetic polarity time scales have been progressing since the early 1960ies and in Iceland the pioneers in this field, T. Einarsson and Th. Sigurgeirsson started their investigation as early as in the mid 1950ies. At first the time scales for the last few million years were mostly based on results from radiometrically dated igneous rocks. Two of the most cited of these were published by Mankinen and Dalrymple (1979) and McDougall (1979), revised by Spell and McDougall (1992). In the last decade time scales based on mathematical models for the sea floor spreading have been employed. The polarity time scale of Cande and Kent (1995) are probably the best known of the latter type and that is the one used for correlation in this work.

In all profiles described here the paleomagnetic polarity was measured carefully using field magnetometer (fluxgate). As a rule three samples from every lava layer were measured and more of them in the case of irregular or weak polarity. With proper precautions this method has proved to be quite reliable and is of considerable help in stratigraphic correlation between nearby profiles.

Skagi peninsula and Skagafjörður Valleys experienced the first paleomagnetic observations in 1957 (Einarsson 1958, 1959) and in all later investigations on the bedrock paleomagnetic measurements have been part of the study. The best known one is the extensive geological and paleomagnetic study in Tröllaskagi of Sæmundsson et al. (1980), covering the eastern part of Skagafjörður District. In the summer of 2002 a new paleomagnetic project was launched in order to prolong the polarity profiles in Skagafjörður up through the Tinná Group. The project is implemented in cooperation with L. Kristjánsson and Á. Guðmundsson. Profiles in Merkidalur, Fjóslækur and Brennigil have been sampled and measured in a laboratory. Preliminary results allow a comparison between field and laboratory measurements. The congruity is excellent and the only discrepancies occur in transitional samples (Kristjánsson pers. comm).

Table 22. Polarity Periods in Skagafjörður Valleys

Polarity zone	Period Ma
IV	1.8 – 0.8
III	5 – 2.8
II	7 – 5
I	10 – 9

Nineteen reversals have been found in the strata pile of Skagafjörður valleys (Hjartarson et al. 1997), 16 below the Skagafjörður unconformity and 3 above it (table 23). The lowest series, the Sólheimar formation, the Merkidalur formation and the lowest part of Ábær member, were sampled and measured in a laboratory during the project of Sæmundsson et al. (1980).

In the light of the bedrock stratigraphy and the new Ar/Ar dates a fourfold division of the paleomagnetic stratigraphy seems reasonable (table 22):

Polarity zone I is the oldest and it is found below the Hidden Unconformity.

Polarity zone II reaches from the Hidden Unconformity up through the entire Tinná Central Volcano to the boundary between Tinná and Fossárdalur Groups.

Polarity zone III starts after extinction of the Tinná Volcano and ends at the Skagafjörður unconformity.

Polarity zone IV is the youngest one and contains the Pleistocene succession.

9.1.2   Polarity zone I

The marine magnetic anomaly 5 has been well established in central North Iceland (Sæmundsson et al. 1980, Jóhannesson 1991, Jóhannesson and Sæmundsson 1998 a and b). According to Sæmundsson et al. (1980) the top of the anomaly is at the Bóla sediments and at the base of Sólheimar Group. The accumulation rate of the lava pile in the Skagafjörður area proved to be exceptionally high, or 3854 m/Ma (Sæmundsson et al. 1980), the highest so far reported in Iceland. Very few polarity subzones should therefore be missing. The thick reverse pile of the Sólheimar formation, with single thin excursion or subzone, correlates probably to the mostly normal 4Ar chronozone of Cande and Kent (1995). The K/Ar-dates around the reversal at its top give approximately 9.1 Ma. (A satisfactory fit with the 9.3 Ma of the paleomagnetic time scale). The polarity subzones of Merkidalur formation could correspond to chrons C4Ar.1n and C4Ar.1r. The Hidden Unconformity demonstrates a gap in the stratigraphy and a rift jump. Therefore it seems natural to expect a decline in the accumulation rate in the pile below it indicating drift out from a dying volcanic zone. Estimates on the accumulation rates in chapter 9.3 support this.

9.1.3   Polarity zone II

The stratigraphy above the Hidden Unconformity seems to indicate continuous accumulation, perhaps with an intermezzo during the formation of the Tinná lignite sediment. As discussed in chapter 2.3 the lifetime of the volcano is estimated to have been 0.5 – 1 Ma, spanning three polarity subchrons. The Skati Dome and the reverse lavas around it are supposed to represent the lowest Gilbert subchron (C3 r) spanning the interval 5.23-5.89 Ma. Thus the polarity reversal below the Skati Dome represents the Gilobert/Epoch 5 boundary.

Above the Skati Dome is the Nýjibær member with acid and intermediate lavas interpreting the normal Þverá subchron (5.23 – 4.98 Ma). The Þverá subchron was first defined and given a name by McDougall 1977. Its type locality is in Borgarfjörður, West Iceland. The stratigraphy of the Gilbert polarity zone in Borgarfjörður, including Þverá subzone, doesn't show any influences of glaciation. In east Iceland, on the other hand, diamictite layers have been found in the lower Gilbert zone that are believed to indicate cooling climate and local glaciations (Hjartarson and Vilmundardóttir 1998). The stratigraphy in Skagafjörður thus shows closer relation to west than east Iceland.

If no polarity subchron is missing between Skati Dome and the Hidden Unconformity, then the lowest polarity zones above it belongs to Epochs 5 and 6. The lowest one represents the normal subchron C3Bn, 6.94 – 7.09 Ma. Hence, the polarity zone II spans the interval 7.09 – 4.98 Ma or roughly 2 million years.

Now it is possible to estimate the duration of the Hidden Unconformity. It spans 2 million years or the interval 7 – 9 Ma in late Miocene.

Table 23. Polarity Subchrons in the Skagafjörður Valleys

Member	Polarity	Subchrons	Age (Ma)	NB
Late Pleistocene	N	C1n	0 – 0.78	Bruhnes
Mid Pleistocene	R	C1r.1r	0.78 – 0.99	Matuyama
Lambá member	N	C1r.1n	0.99 – 1.07	Jaramillo
Early Pleistocene	R	C1r.2r		Mid Matuyama
Plio-Pleistocene			2	Sediment
Unconformity	--	----------	2.6 – 1.6	----------
Geldingsá member	N	C2An.1n	2.58 – 3.04	Gauss Ar/Ar = 2.6
Hölkná member	R			
Hölkná member	N			
Fossárdalur memb.	R			
Nýjibær formation	N	C3n.4n	4.98 – 5.23	Þverá
Tinná formation	R	C3 r	5.23 – 5.89	Ar/Ar = 5.21
Ábær	N	C3An.1n	5.89 – 6.14	Epoch 5
Ábær	R			Discontinuous
Ábær	N	C3An.2n	6.27 – 6.57	
Ábær	R			Epoch 6, discont.
Ábær	N	C3Bn	6.94 – 7.09	
Merkidalur (me3)				Sediment
Unconformity	--	----------	7 – 9	----------
Merkidalur (me2)	R	C4Ar.1r	9.03 – 9.23	
Merkidalur (me1)	N	C4Ar.1n	9.23 – 9.31	
Sólheimar	R	C4A r	9.7 – 9.31	

9.1.4   Polarity zone III

The Tinná Central Volcano must have been a dominating mountain in the Skagafjörður area in its days rising high above its environment. After it became extinct a long time must have lasted before it was immersed by the regional lava flows. Moreover the build up rate is predicted to have been low during the formation of Fossárdalur group (Fossárdalur, Hölkná and Geldingsá formations), and many magnetic subchrons must be missing. But according to the Ar/Ar-dates, and the age interpretations in last chapters,

these formations represent 2.4 million years or the interval 5 Ma (top of Þverá) to 2.6 Ma containing 12 subchrons (Cande and Kent 1995), including Síðufjall, Nunivak, Cohiti, Mammoth and Caena. Of these 12 only 4 polarity subchrons are found.

Table 23 gives preliminary estimates of how the stratigraphy and polarity log of Skagafjörður Valleys fits to the polarity time scale of Cande and Kent (1995).

9.1.5   Polarity zone IV

Polarity zone IV contains the entire Skagafjörður Pleistocene volcanic succession. Four magnetic subzones have been recognized inside it. The oldest one is the mid Matuyama reverse zone (C1r2r), between Olduvai and Jaramillo, composed of several lavas and sedimentary layers in between, making up extensive cover in the western part of the area shown on the included Bedrock Map.

The second subzone is a thin discontinuous horizon represented by 1 – 3 lavas with normal polarity. This is most probably the Jaramillo subzone that lasted for 80 thousand years around one million years ago. Lavas from Jaramillo have been mapped in several places in Iceland: In Northeast Iceland (Hjartarson and Vilmundardóttir, 1998) in Hreppar, South Iceland (Kristjánsson et al., 1998), in Ingólfsfjall, South Iceland (Kristjánsson et al., 1988) and in Southwest Iceland (Friðleifsson, 1985). In all these sites only a few lavas (1-3) comprise normal polarity. The sections through Jaramillo in Skagafjörður valleys show interglacial lavas sandwiched by glacial sediments.

Above Jaramillo is the uppermost Matuyama subzone made of lavas, hyaloclastites and sediments. Finally, formations from the Brunhes polarity zone cover extensive areas south of Hofsjökull glacier. The Brunhes/Matuyama reversal marks the early/late Pleistocene boundary, a divide that can be seen on all geological maps of Iceland.

Table 24. Accumulation rate in Iceland

Area	Thick-ness m	Accumulation rate m/Ma	Eruption frequency	References
E Iceland	7000	620	16000	Watkins and Walker 1977
Tröllaskagi, lower		1000		Sæmundsson er al. 1980
Tröllaskagi, upper		3850		Sæmundsson er al. 1980
Flateyjarskagi, upper	500	500	50000	Jancin et al. 1985
Flateyjarskagi, lower	700	1000	17000	Jancin et al. 1985
Langidalur	1150	1000		Kristjánsson et al. 1992
NV Peninsula W	4055	1816	5000	McDougall et al. 1984
NV Peninsula E	3165	670	12000	McDougall et al. 1976
Borgarfjörður		730		McDougall et al. 1977
Neskaupstaður		690		McDougall 1976
Skagaf. Valleys, Sólheimar group	1300	1700		This work
Skagaf. Valleys, Tinná group	1000	500		This work
Skagafj. Valleys, top of Neogene pile	260	100		This work
Skagafjörður, Pleistocene	200	200		This work

9.2   Accumulation Rate

The rate of accumulation during the formation of the Neogene succession in Skagafjörður Valleys can be calculated on basis of the general geological section shown on the included Bedrock Map. There the average thickness of every formation is given. The total thickness of the Neogene pile is around 1550 m. If it accumulated during 6.7 million years, between 9.3 and 2.6 Ma, the rate was 230 m/Ma. Compared to reported accumulation rates elsewhere in Iceland, where it is 600 – 1200 m/Ma in most places (table 24), this value is very low. The main explanation is of course the major unconformities hiding in the pile.

 In the Skagafjörður Valleys a great difference could be expected between the accumulation rate in the upper and lower parts of the pile. In the lower part the rate is expected to have been high. Then the Snæfellsnes – Húnaflói rift zone was in full action and intense volcanism seems to have indicated the area. Sæmundsson et al. (1980) found exceptionally high accumulation rate for the upper part of their Tröllaskagi section (between 9.5 and 8.9 Ma) right below and partly overlapping the section in Skagafjörður valleys, or about 3800 m/Ma. If the division and age estimates described in the chapter

about the paleomagnetism are correct (chapter 9.2, table 23) the accumulation slowed down during the formation of Sólheimar group. Then it was 1700 m/Ma, and if only the upper part of the group is taken in account (the Merkidalur formation) then the accumulation rate is 1000 m/Ma or the same as in the lower Tröllaskagi pile (Sæmundssom et al. 1980).

Above the Sólheimar group an unconformity is expected to hide inside the pile demonstrating 2 million years of erosion and ablation of the bedrock. After that the volcanism started up again when the North Iceland rift zone was initiated. During the first one or two million years of the zone the Snæfellsnes – Húnafói rift zone might have been active simultaneously resulting in rather low production in both zones.

The accumulation during the 2 million years interval of the Tinná group was rather slow. The pile is 1000 m thick so the rate is 500 m/Ma. This is lower than might have been expected inside a central volcano.

During the formation of the topmost part of the pile in Skagafjörður Valleys, Fossárdalur group, the accumulation rate is expected to have been exceptionally low or only 100 m/Ma. Hiatuses and gaps in the strata pile explain this.

The accumulation rate during the formation of the Pleistocene succession is hard to estimate. In Skagi and in the areas north of Hofsjökull the thickness of the succession might be approximately 200 m, and extruded during a million years or so.

10 REFERENCES

Abelson, M. and Agnon, A., 2001: Hotspot activity and plume pulses recorded by geometry of spreading axes. Earth and Planetary Science Letters 189, 31-47.

Aðalsteinsson, H., 1985: Lífvist í tjörnum og vötnum á Hofsafrétt. OS-85046/VOD-04, 48 p.

Arnórsson, S. and Sveinbjörnsdóttir, Á.E., 2000: Age and chemical evolution of groundwater in Skagafjördur, Iceland. Journal of Geochemical Exploration 69-70, 459-463.

Bernstein, S., Kelemen, P.B., Tegner, C., Kurz, M.D., Blusztajn, J. and Brooks, C.K., 1998: Post-breakup basaltic magmatism along the East Greenland Tertiary rifted margin. Earth and Planetary Science Letters 154, 221-235.

Björnsson, B. J., 1975: Villinganes. Jarðfræðiyfirlit. Orkustofnun, OS-ROD 7501, 13 pp.

Björnsson, B. J., 1975: Jökulsá eystri, Skagafirði. Gangar og brotalínur. Jarðfræðikort. Unpubl. data from Orkustofnun.

Cande, S. C. and Kent, D. V., 1995: Revised calibration of the geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. Journal of Geophysical Research 100, 6093-6095.

Carmichael, I. S. E. 1964: The petrology of Thingmúli, a Tertiary volcano in E Iceland. Journal of Petrology 5, 435-460.

DeMets, C., Gordon, R.G., Argus, D.F. and Stein, S., 1990: Current Plate motions. Geophys. J. Int. 101, 425-478.

Dunbar, C. O. and Rodgers, J., 1957: Principles of Stratigraphy. John Wiley & sons, New York, 365 pp.

Einarsson, P. 1991: Umbrotin við Kröflu 1975 – 1989. In (ed.) Garðarsson and Einarsson: Náttúra Mývatns. Hið íslenska náttúrufræðifélag, Reykjavík, 97-139.

Einarsson, T., 1958: Landslag á Skagafjallgarði, myndun þess og aldur. Náttúrufræðingurinn 28, 1-56.

Einarsson, T., 1958b: A Survey of the Geology of the area Tjörnes – Bárðardalur. Societas Scientarium Islandica XXIX, 1-62.

Einarsson, T., 1959: Jarðeldasvæði um norðanverðan Skagafjörð. Aldursákvörðun á landslagi á Miðnorðurlandi. Náttúrufræðingurinn 29, 113-168.

Einarsson, T., 1959b: Studies of Pleistocene in Eyjafjörður, Middle Northern Iceland. Societas Scientiarum Islandica XXXII, 1-79.

Einarsson, T., 1962: Upper Tertiary and Pleistocene rocks in Iceland. Societas Scientiarum Islandica XXXVI. Reykjavik, 197 pp + maps.

Einarsson, T., 1965: Remarks on crustal structure in Iceland. Geophysical Journal of the Royal Astronomical Society 10, 283-288.

Einarsson, T., 1968: Submarine ridges as an effect of stress fields. Journal of Geophysical Research 73, 7561-7576.

Einarsson, T., 1971: Eðlisþættir jarðarinnar og jarðsaga Íslands . Almenna Bókafélagið, Reykjavík, 267 pp.

Einarsson, Th. 1968: Jarðfræði. [Geology]. Mál og Menning, Reykjavík

Einarsson, Th. 1991: Myndun og mótun lands. [Formation and figuration of the landscape]. Mál og Menning, Reykjavík.

Eiríksson, J. 1981: Lithostratigraphy of the upper Tjörnes sequence, north Iceland: The Breiðavík Group. Acta Naturalia Islandica 29, 37 pp.

Eiríksson, J., Guðmundsson, A. I., Kristjánsson, L. and Gunnarsson, K 1990: Paleomagnetism of Pliocene-Pleistocene sediments and lava flows on Tjörnes and Flatey, North Iceland. Boreas 19, 39-55.

Everts, P., 1975: Die Geologie von Skagi und der Ost-Küste des Skagafjörðurs (Nord-Island). Sonderveröffentlichung des Geologischen Instituts der Universität Köln, 114 pp. + maps.

Everts, P., Koerfer, L.E. and Swartzbach, M., 1972: Neue K-Ar-Datierungen islandischer Basalte. Neues Jahrbuch Geol. Palaeontol. Monatsh.5, 280.

Flores, R. M., Jónsson, J. and Jóhannesson, H., 1981: A short account of the Holocene tephrochronology of the Snæfellsjökull central volcano, western Iceland. (Ljósu öskulögin frá Snæfellsjökli), Jökull 31, 23-30.

Friðleifsson, I.B., 1985: Jarðsaga Esju og nágrennis. Árbók Ferðafélags Íslands, 141-172.. OS-84089/JHD-16, Orkustofnun, Reykjavík, 134 p + maps.

Gale, N. H., Moorbath, S., Simons, J. and Walker, G. P. L. 1966. K-Ar ages of acid-intrusive rocks from Iceland. Earth and Planetary Science Letters 1, 284-288.

Garcia, S., Angelier, J., Bergerat, F. and Homberg, C. 2002: Tectonic analysis of an oceanic transform fault zone based on fault-slip data and earthquake focal mechanisms: the Húsavík-Flatey Fault zone, Iceland. Tectonophysics 344, 157-174.

Garcia, S, Arnaud, A.O., Angelier, J, Bergerat, F and Homberg, C. 2003. Rift jump process in Northern Iceland since 10 Ma from 40Ar/39Ar geochronology. EPSL 214, 529-544.

Geirsdóttir, Á. 1990: Diamictites of late Pliocene age in western Iceland, Jökull 40 3-25

Geirsson, H. 2003. Continuous GPS measurements in Iceland 1999 – 2003. Report 03014, Icelandic Meteorological Office, Reykjavik.

Grönvold, K., Oskarsson, N., Johnsen, S. J., Clausen, H. B., Hammer, C. U., Bond, G. and Bard, E., 1995: Ash layers from Iceland in the Greenland GRIP ice core correlated with oceanic and land sediments. Earth and Planetary Science Letters 135, 149-155.

Guðmundsson, A. T., 2001: Íslandseldar. Mál og menning, Reykjavík.

Guðmundsson, Á., 1991: Jökulsár í Skagafirði. Geological map, 1:50.000. Unpubl. manuscript, Orkustofnun, Reykjavík.

Guðmundsson, Á., 1990: Dyke emplacement at divergent plate boundaries. In: A.J.Parker, P.C.Rickwood and D.H.Tucker (eds.), Mafic Dykes and Emplacement Mechanisms, p. 47 – 62, Rotterdam.

Guðmundsson, Á., 1995: Infrastructure and mechanics of volcanic systems in Iceland. Journal of Volcanology and Geothermal Research 64, 1-22.

Guðmundsson, Á., 1986: Mechanical Aspects of Postglacial Volcanism and Tectonics of the Reykjanes Peninsula, Southwest Iceland. Journal of Geophysical Research, 91, 12711-12721.

Guðmundsson, Á., 2000: Dynamics of Volcanic Systems in Iceland: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge System. Annual Review of Earth and Planetary Sciences 28, 107-140.

Hallgrímsson, H., Pálsson, J., Björnsson, H., Kristinsson, H. og Haraldsson, Þ., 1982: Nátúrufarskönnun á virkjunarsvæði Héraðsvatna við Villinganes í Skagafirði. Orkustofnun, Reykjavík, OS82047/VOD08, 186 pp. + app.

Hallgrímsson, J., [1989]: Ritverk Jónasar Hallgrímssonar I-IV. Svart á hvítu, Reykjavík.

Harðarson, B. A., 1983. Kjarnagreining og sýnataka. Endurskoðuð vinnulýsing. Greinargerð OS BAH-83/01, 27 p.

Harðarson, B. A., 1989. Blönduvirkjun. Jarðgöng. Reynsla af breggæðamati. Orkustofnun, Reykjavík, OS-89021/VOD-03, 16 p.

Harðarson, B. A. and Guðmundsson, Á., 1986: Stafnsvatnavirkjun. Mannvirkjajarðfræði. Orkustofnun, Reykjavík, OS-86039/VOD-14 B, 63 pp. + maps.

Harðarson, B.S., Fitton, J.G., Ellam, R.M. and Pringle, M.S., 1997: Rift relocation – a geochemical and geochronological investigation of a paleo-rift in NW Iceland. Earth and Planetary Science Letters 153,181-196.

R.A. Duncan and J. Helgason. Precise dating of the Holmatindur cooling event in eastern Iceland: Evidence for mid-Miocene bipolar glaciation. JGR 103 (1998) 12397-12404.

Hjartarson, Á., 1973: Rof jarðlagastaflans milli Eyjafjarðar og Skagafjarðar og ísaldarmenjar við utanverðan Eyjafjörð. B.Sc. – thesis (unpublished). University of Iceland 37 pp. + map

Hjartarson, Á., 1994: Environmental changes in Iceland following the Great Þjórsá Lava Eruption 7800 14C years BP. In: Environmental Change in Iceland. Ed. J.Stötter & F.Wilhelm, Munchen, 147-155.

Hjartarson Á (1995) Hekla and its surroundings (Á Hekluslóðum). Árbók Ferðafélags Íslands 1995, Reykjavík1-236

Hjartarson, Á., 1999: Yfirlit um jarðfræði í Ölfusi og grend. ÁH 99/06, Orkustofnun, Reykjavík, 11 pp. + map.

Hjartarson, Á., 2000: Þjórsárhraunið milka, þykktir, flatarmál, rúmmál. Orkustofnun, Reykjavík, ÁH 2000/02, 5 pp.

Hjartarson, Á., 2002: The Skagafjord Ephemeral Volcanic Zone. In (ed. S. S. Jónsson). The 25th Nordic Geological Winter Meeting. January 6th – 9th, 2002 – Reykjavík, Iceland. Abstract Volume.

Hjartarson, Á. and Hafstað, Þ. H., 1997: Sviðinhornahraun. Berggrunnsrannsóknir og kort. OS-97016, Orkustofnun, Reykjavík. 32 pp. + maps.

Hjartarson, Á. and Vilmundardóttir, E.G., 1998: Vesturöræfi - Hraun. Samræming jarðfræðikorta á Austurlandi. OS-98027, Orkustofnun, Reykjavík, 32 pp + map.

Hjartarson, Á., Friðleifsson, G. Ó. and Hafstað, Þ. H., 1997: Berggrunnur í Skagafjarðardölum og jarðgangaleiðir. OS-97020 Orkustofnun, Reykjavík, 55 pp. + maps.

Hjartarson, Á. and Hafstað, Þ. H., 1999: Jarðfræði á Hofsafrétt í Skagafirði. OS-99031, Orkustofnun, Reykjavík, 13 pp. + maps.

Hjartarson, Á. and Jónsdóttir, H.E., 1999: Akureyri. Geological map 1:50.000. OS-99118, Orkustofnun, Reykjavík, 18 pp + map.

Icelandic Meteorological Office homepage: *www.vedur.is*

Ingólfsson Ó (1998) Isavsmältning och israndlinjer på Island. In Andersen and Pedersen: Israndslinier i Norden. TemaNord 1998:584. Nordisk Ministeråd, København, pp 109-124

Jakobsson, S.P., 1972: Chemistry and distribution pattern of recent basaltic rocks in Iceland. Lithos 5, 365-386.

Jakobsson, S.P., 1979a: Petrology of recent basalts of the Eastern volcanic zone, Iceland. Acta Naturalia Islandica 26, 103 p.

Jakobsson, S.P., 1979b: Outline of the petrology of Iceland. Jökull 29, 57-73.

Jakobsson, S. P., Jónsson, J. and Shido F., 1978: Petrology of the Western Reykjanes Peninsula, Iceland. Journal of Petrology 19, 669-705.

Jancin, M., Young, K. D., Voight, B., Aronson, J. L. and Sæmundsson, K., 1985: Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland. Journal of Geophysical Research 90, 9961-9985.

Japsen, P., Bonow, J., Klint, K. E. and Jensen, K., 2002: Neogene uplift, erosion and resedimentation in West Greenland. Field report summer 2002. Geological survey of Denmark and Greenland, rapport 2002/71.

Jones, S.M. 2003: Test of ridge-plume interaction model using oceanic crustal structure around Iceland. Earth and Planetary Science Letters 6564, 1-14 (in press)

Jóhannesson, H., 1980: Jarðlagaskipan og þróun rekbelta á Vesturlandi. Náttárufræðingurinn 50, 13-31.

Jóhannesson, H., 1989: Aldur Hallmundarhrauns í Borgarfirði. Fjölrit Náttúrufræðistofnunar Íslands nr. 9. 12 p.

Jóhannesson, H., 1991: Yfirlit um jarðfræði Tröllaskaga (Miðskaga). Árbók FÍ, 1991, 39-56.

Jóhannesson, H and Sæmundsson K., 1998a: Geological map of Iceland 1:500.000. Bedrock Geology. Icelandic Institute of Natural History, Reykjavik (2nd edition)

Jóhannesson, H and Sæmundsson K., 1998b: Geological map of Iceland 1:500.000. Tectonics. Icelandic Institute of Natural History, Reykjavik.

Jónsson, J., 1978: Jarðfræðikort af Reykjanesskaga. OS-JHD 7831, Orkustofnun, Reykjavík, 333 pp. + maps.

Jónsson, B., Egilson, D. and Zóphóníasson, S., 1977: Villinganes. Mannvirkjajarðfræði. Orkustofnun, Reykjavík, OS-ROD 7709, 19 p.

Jull, M. and McKenzie,D., 1966: The effort of deglaciation on mantel melting beneath Iceland. Journal of Geophysical Research 101, 21815-21828.

Kaldal, I. and Víkingsson, S., 1978: Jökulsár í Skagafirði I. Jarðfræði. Orkustofnun, Reykjavík, OS-ROD 7805, 33 pp. + maps.

Kaldal, I. and Víkingsson, S., 1979: Jökulsár í Skagafirði II. Jarðgrunnskort. OS-79044/ROD-16. (Two maps in an album).

Karlsdóttir, R., Haraldsson, G. I., Ingimarsdóttir, A., Guðmundsson, Á. and Hafstað, Þ. H., 1991: Skagafjörður. Jarðfræði, jarðhiti, ferskvatn og rannsóknarboranir. Sérverkefni í fiskeldi, 1987. Orkustofnun, Reykjavík, OS-91047/JHD-08, 96 pp. + maps.

Kjartansson, G., 1965: Jarðfræðikort af Íslandi. Mið-Ísland. Náttúrugripasafn Íslands, Reykjavík.

Kjartansson, G., 1966: Rúmmál hraundyngna. Náttúrufræðingurinn 36, 125

Knudsen, K. L. and Eiríksson, J. 2002: Application of tephrochronology to the timing and correlation of palaeoceanographic events recorded in Holocene and Late Glacial shelf sediments off North Iceland. Marine Geology, 191, 165-188.

Kristjánsson, L., Jóhannesson, H. and McDougall, I., 1992: Stratigraphy, age and paleomagnetism of Langidalur, Northern Iceland. Jökull, 42, 31-44.

Kristjansson, L., Guðmundsson, Á. and Haraldsson, H., 1995: Stratigraphy and paleomagnetism of a 3-km-thick Miocene lava pile in the Mjoifjördur region, eastern Iceland. Geol. Rundsch. 84, 813-830.

Kristjánsson, L., Jóhannesson, H., Eiríksson, J. and Guðmundsson, A.I., 1998: Brunhes – Matuyama paleomagnetism in three lava sections in Iceland. Canadian Journal of Earth Sciences 25, 215-225.

Kristjánsson L. and Jónsson, G., 1998: Aeromagnetic results and the presence of an extinct rift zone in Western Iceland. J. Geodyn. 25, 99-108.

Kristjánsson, L., Duncan, R.A., and Guðmundsson, Á., 1998: Stratigraphy, paleomagnetism and age of volcanics in the upper regions of Þjórsárdalur valley, central south Iceland. Boreas 27, 1-13.

Lacasse, C., Paterne, M., Werner, R., Wallabre-Adams, H.-J., Sigurðsson, H., Carey, S. and Pinte, G., 1996a: Geochemistry and origin of Pliocene and Pleistocene ash layers from the Icelandic plateau, site 907. p. 309-331. In: Thiede et al. 1996: Proceedings of the Ocean Drilling Program, Scientific Results, Vol. 151, 309-331.

Lacasse, C., Sigurðsson, H., Carey, S. N., Paterne, M. and Guichard F. 1996b: North Atlantic deep-sea sedimentation of late Quaternary tephra from the Iceland hotspot. Marine Geology 129, 207-235.

Lacasse, C. and Grabe-Schönberg, C.-D., 2001: Explosive silicic volcanism in Iceland and Jan Mayen area during the last 6 Ma: sources and timing of major eruptions. Journal of Volcanology and Geothermal Research 107, 113-147.

Larsen, G., 2000: Holocene eruptions within the Katla volcanic system, south Iceland: Characteristics and environmental impact. Jökull 1- 28.

Larsen, G. and Thorarinsson, S., 1977: H4 and other acid Hekla tephra layers. Jökull 27, 28-45.

Larsen, L. M., Watt, W. S. and Watt, M. 1989: Geology and petrology of the Lower Tertiary plateau basalt of the Scoresby Sund region, East Greenland. Grønlands Geologiske Undersøgelser, bulletin 57, 164 pp.

Lawver, L.A. and Müller, R.D. 1997: Iceland hotspot track. Geology 22, 311-314.

Le Maitre, R.W. (Editor), Bateman, P., Dudek, A. Keller, J., Lameyre, J., Le Bas, M. J., Sabine, P. A., Schmid, R., Sörensen, H., Streckeisein, A. Wolley, R. W. and Zanettin, B., 1989. A Classification of Igneous Rocks and Glossary of Terms, Recommendations of the International Union of Geological Sciences, Subcommission on the Systematics of Igneous Rocks, 193 p.

Líndal, J. H., 1940: Mælifellshnjúkur. Náttúrufræðingurinn 10, 51-67.

Líndal, J. H., 1941: Drangey og hvernig hún er til orðin. Náttúrufræðingurinn 11, 41-46.

Líndal, J. H., [1964]: Með huga og hamri. Jarðfræðidagbækur og ritgerðir. 410 p.

Magnússon, G., 1980: Um nokkra surtarbrandsstaði í Vesturdal í Skagafirði. Týli 10, 7-12.

Magnússon, G., 1981: Um nokkra surtarbrandsstaði í Austurdal. Týli 11, 53-58.

Mankinen, E. A., and Dalrymple, G. B. 1979: Revised geomagnetic polarity time scale for the period 0-5 My BP. Journal of Geophysical Research 84, 615-626.

McDougall, I.1979: The present status of the geomagnetic polarity time scale. In McElhinny, M. W. (ed.): The Earth, its origin, structure and evolution, 543-565. Academic Press, London.

McDougall, I., Watkins, N.D., Walker, G.P.L. and Kristjansson, L., 1976: Potassium-Argon and Paleomagnetic Analyses of Icelandic Lava Flows: Limits on the Age of Anomaly 5. Journal of Geophysical Research 81, 1505-1512.

McDougall, I., Saemundsson K., Johannesson, H., Watkins, N. D. and Kristjansson, L. 1977: Extension of the geomagnetic polarity time scale to 6.5 m.y.: K-Ar dating, geological and paleomagnetic study of a 3,500-m lava succession in western Iceland. Geological Society of America Bulletin 88, 1-15.

McDougall, I., Kristjansson, L. and Saemundsson K. 1984: Magnetostratigraphy and geochronology of Northwest Iceland. Journal of Geophysical Research 89,7029-7060.

Moorbath, S., Sigurdsson, H. and Goodwin, R., 1968: K:Ar ages of the oldest exposed rocks in Iceland. Earth and Planetary Science Letters 4, 197-205.

NACSN (North American Commission on Stratigraphic Nomenclature) 1983: North American Stratigraphic Code. The American Association of Petroleum Geologists Bulletin 67, 841-875

Ninkovich, D., Sparks,R.S.J. and Ledbetter, M.T., 1978: Trhe exceptional magnitude and intensity of the Toba eruption Sumatra; An example of the use of deep sea tephra layers as a geological tool. Bulletin of Volcanology 41, 1-13.

Ólafsson, E., [1943]: Ferðabók Eggerts Ólafssonar og Bjarna Pálssonar um ferðir þeirra á Íslandi 1752-1757 II. bindi. Reykjavík, 317 p.

Pálmason, G. 1980. A continuum model of crustal generation in Iceland: Kinematic aspects. Journal of Geophysics 47, 7-18.

Pálmason, G. 1981. Crustal rifting and related thermomechanical processes in the lithosphere beneath Iceland. Geol. Rundsch. 70, 244-260.

Pálmason, G. and Sæmundsson, K., 1974: Iceland in Relation to the Mid-Atlantic Ridge. Annual Review of Earth and Planetary Sciences 2, 25-50.

Pálsson, J., Haraldsson, Þ., Jóhannsson, B. and Kristinsson, H., 1990: Jökulsár í Skagafirði - Hofsafrétt. Gróðurfar á áætluðum virkjunarsvæðum. Könnun gerð af Náttúrufræðistofnun Norðurlands. OS-90051/VOD-04, 55 p.

Pálsson, S., [1983]: Ferðabók Sveins Pálssonar

Pedersen, A. K., 1985: Lithostratigraphy of the Tertiary Vaigat Formation on Disko, central West Greenland. The Geological Survey of Greenland Report No. 124. Kopenhagen 30 pp.

Pedersen, A. K., Larsen, L. M., Riisager, P. and Dueholm, K. S. 2002: Rates of Volcanic deposition, facies changes and movements in a dynamic basin: the Nuussuaq Basin, West Greenland, around the C27n-C26r transition. In: Jolley, D. W. and Bell, B. R. (eds.) The North Atlantic Igneous Province: Stratigraphy, Tectonic, Volcanic and Magmatic Processes. Geological Society of London, Special publications, 197, 1-25

Pjetursson, H., 1905: Om Islands Geologi. Meddelser fra Dansk Geologisk Forening Nr. 11, 1-104.

Salvador, A. (ed) 1984: International Stratigraphic Guide, Second Edition. Geological Society of America, 220 pp.

Schilling, J.-G., Sigurðsson, H. and Kingsley, R.H., 1978. Skagi and Western Neovolcanic Zones in Iceland: 2. Geochemical Variations. Journal of Geophysical Research 83: 3983-4002

Sen, A., Kendall, C. G. St. G. and Levine, P., 1999: Combining a computer simulation and eustatic events to date seismic sequence boundaries: a case study of the Neogene of Bahamas. Sedimentary Geology 125, 47-59.

Serle, R., 1976: Spreading rate in Iceland. Nature 261, 75.

Sigmundsson, F. 1991: Post-glacial rebound and astenosphere viscosity in Iceland. Geophysical Research Letters 18, 1131-1134.

Sigmundsson, F., Einarsson, P., Bilham, R. and Sturkell, E. 1995. Rift transform kinematics in south Iceland: Deformation from Global Positioning System measurements 1986 to 1992. JGR 100, 6235-6248.

Sigurðsson, F., Sæmundsson, K. 1984: Surtarbrandur á Vestfjörðum: Orkustofnun Reykjavík, OS-84039/ODB-02, pp 43.

Sigurðsson, H., 1970: Structural origin and plate tectonics of the Snaefellsnes volcanic zone, western Iceland. Earth and Planetary Science Lettars 10, 129-135.

Sigurðsson, H., Schilling, J.G. and Meyer, P.S., 1978. Skagi and Langjökull Volcanic Zones in Iceland. 1. Petrology and Structure. Journal of Geophysical Research 83: 3971-3982

Sigurðsson, O., Zóphóníasson, S., Hannesdóttir, L. and Þorgrímsson, S. 1975: Dettifdossvirkjun, jarðfræðiskýrsla, OS-ROD-7526, Orkustofnun, Reykjavík, pp 1-62 + maps

Sigvaldason, G. E., 1992: Recent hydrothermal explosion craters in an old hyaloclastite flow, central Iceland. Journal of Volcanology and Geothermal Research 54, 53-63.

Sigvaldason, G. E., 2002: Volcanic and tectonic processes coinciding with glaciation and crustal rebound: an early Holocene rhyolitic eruption in the Dyngjufjöll volcanic center and the formation of the Askja caldera, north Iceland. Bulletin of Volcanology 64, 192-205.

Sigvaldason, G. E., Annertz, K. and Nilsson, M., 1992: Effect of gracier loding/deloding on volcanism: postglacial volcanic production rate of the Dyngjufjöll area, central Iceland. Bulletin of Volcanology 54, 385-392

Sigvaldason, G. E. and Steinþórsson, S. 1974: Chemistry of tholeiitic basalts from Iceland and their relationship to the Kverkfjöll hot spot. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, p. 155-164.

Sinton, J. 2001: Postglacial volcanics of the Western Volcanic Zone. Geologic map. University of Hanaii, Honolulu, HI, USA.

Sinton, J., Grönvold, K., 2001: Post-Glacial Eruptive History of the Western Volcanic Zone, Iceland. (Poster)

Spell, T.L. and McDougall, I., 1992: Revisions to the age of the Brunhes-Matuyama boundary and the Pleistocene geomagnetic polarity time scale. Geophysical Research Letters, 19, 1181-1184.

Steinþórsson, S., 1967: Tvær nýjar Cl4-aldursákvarðanir á öskulögum úr Snæfellsjökli, Náttúrufræðingurinn 37, 236-238.

Sæmundsson, K., 1973 Straumrákaðar klappir í kringum Ásbyrgi. Náttúrufræðingurinn 43, 52-60.

Sæmundsson, K., 1974: Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. Geological Society of America Bulletin 84, 495-504.

Sæmundsson, K., 1979: Outline of the geology of Iceland. Jökull 29, 7-28.

Sæmundsson, K., 1992: Geology of the Thingvallavatn area. In: ed. Jónasson 1992, Ecology of oligotrophic, subarctic Thingvallavatn, pp. 434

Sæmundsson, K., 1995: Um aldur stóru dyngnanna á utanverðum Reykjanesskaga. In: Björn Hróarsson, Dagur Jónsson og Sigurður Sveinn Jónsson (eds): Eyjar í eldhafi. Gott mál, Reykjavík, pp.165-172.

Sæmundsson, K., Kristjánsson, L., McDougall, I. and Watkins, N. D.1980: K-Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. Journal of Geophysical Research 85, 3628-3646.

Thorarinsson S., 1951: Laxárgljúfur and Laxárhraun a tephrochronological study. Geografiska annaler 1951, 1-89.

Thorarinsson, S., 1958: The Öræfajökull eruption of 1362. Acta Naturalia Islandica II, 1-100

Thorarinsson, S., 1968: Síðustu þættir Eyjaelda. Náttúrufræðingurinn 57, 113-135.

Thorarinsson S., 1979: Skeljungssteinn, Náttúrufræðingurinn 43 183-185

Thorarinsson S., 1979: The postglacial history of the Mývatn area. Oikos 32, 17-28.

Thordarson, Th., Self, S., 1993: The Laki (Skaftar fires) and Grimsvotn eruptions in 1783 – 1785. Bull. Volcanol. 55, 33-63.

Thordarson, T., Miller, D. J., Larsen, G., Self, S. and Sigurdsson, H., 2001: New estimates of sulfur degassing and atmospheric mass-loading by the 934 AD Eldgja eruption, Iceland. Journal of Volcanology and Geothermal Research 108, 33-54.

Torfason, H., Hjartarson, Á., Jóhannesson, H., Jónsson, J. and Sæmundsson, K. 1997: Bedrock map, Viðey 1613 III NV-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason, H., Hjartarson, Á., Jóhannesson, H., Jónsson, J. and Sæmundsson, K. 1999: Bedrock map, Vífilsfell 1613 III SA-B, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Torfason, H., Hjartarson, Á., Jóhannesson, H. and Friðleifsson, I. B. 2000: Bedrock map, Mosfellssveit 1613 III NA, 1:25.000. Landmælingar Íslands, Orkustofnun, Hafnarfjarðarbær, Garðabær, Kópavogsbær, Seltjarnarnesbær and Reykjavíkurborg.

Verkfræðistofa Sigurðar Thoroddsen, 1975: Virkjun Héraðsvatna I. Frumáætlun við 32 MV virkjun við Villinganes. Orkustofnun, Reykjavík, OS-ROD 7522, 37 pp.

Verkfræðistofa Sigurðar Thoroddsen, 1984: Virkjun Austari-Jökulsár. Stafnsvatnavirkjun. Forathugun. Orkustofnun, Reykjavík, OS-84105/VOD-11, 37 pp. + maps.

Vilmundardóttir, E. G., 1977: Tungnárhraun. OS ROD 7702, Orkustofnun, Reykjavík, 156 pp. + map.

Vilmundardóttir, E. G. and Sigurðsson, F., 1984: Tillögur um staðal fyrir berggrunnskort OS-VOD í mælikvarða 1:50.000. OS-84004/VOD-04, Orkustofnun, Reykjavík., 19 pp.

Vilmundardóttir, E. G. and Larsen, G.1986: Productivity pattern of the Veiðivötn fissure swarm, southern Iceland,in postglacial time. Preliminary results. 17. Nordiska Geologmötet , Helsingfors.

Vilmundardóttir, E. G., Snorrason, S. P., Larsen, G. and Guðmundsson, Á., 1988: Bedrock map. Sigalda – Veiðivötn 1:50.000

Vilmundardóttir, E. G., Hafstað, Þ. H., Kaldal, I. og Víkingsson, S., 1997: Vestari-Jökulsá. Berggrunnur og jarðgrunnur. jarðfræðikortlagning, 1997. OS-97068, Orkustofnun, Reykjavík. 9 pp. + maps.

Víkingsson, S., 1987: Stafnsvatnavirkjun. Byggingarefni. Orkustofnun, Reykjavík, OS-87047/VOD-05, 37 pp.

Vogt, P.R., 1971: Astenosphere motion recorded by the ocean floor south of Iceland. Earth and Planetary Science Letters, 13: 153-16.

Vogt, P.R. and Avery, O.E. 1974: Detailed magnetic surveys in the Northeast Atlantic and Labrador Sea. Journal of Geophysical Research 79, 263-389.

Völksen, C. 2000. Die Nutzung von GPS für die Deformationsanalyse in regionalen Netzen am Beispiel Islands. Wissenschaftliche Arbeiten der Fachrichtung Vermessungswesen der Universität Hannover, nr. 237.

Walker, G.P.L., 1959: Geology of the Reyðarfjörður area, eastern Iceland. Geological Society of London Quaternary Journal 114, 367-393.

Walker, G. P. L., 1960: Zeolite zones and dike distribution in relation to the structure of the basalts of eastern Iceland. Journal of Geology 68, 515-528.

Walker, G.P.L., 1963: Breiddalur Central Volcano, eastern Iceland. Quarterly Journal of the Geological Society of London, 119, 29-63.

Walker, G.P.L., 1963: Some aspects of Quaternary volcanism in Iceland. Trans. Leicester Lit. Philos. Soc. 49, 25-40.

Walker, G.P.L., 1971: Compound and Simple Lava Flows and Flood Basalts. Bulletin of Volcanology 35, 579-590.

Walker, G.P.L. 1974. The structure of Eastern Iceland. In Kristjánsson (ed.): Geodynamics of Iceland and the North Atlantic Area. D. Reidel , Hingham, Mass, 155-164

Walker, G.P.L., 1975: Excess spreading axes and spreading rate in Iceland. Nature 255, 468-470.

Wallrabe-Adams, H.-J. and Werner, R., 1999: Chemical composition of middle Miocene – lower Pliocene ash from sites 982 and 985. In: Raymo et al. 1999: Proceedings of the ODP, Scientific Results, vol. 162.

Ward, P. L. 1971: New Interpretation of the Geology of Iceland. Geological Society of America Bulletin 82, 2991-3012.

Watkins, N.D. and Walker, G.P.L., 1977: Magnetostratigraphy of Eastern Iceland. American Journal of Science 277, 513-584.

Wendt, K., Möller, D: and Ritter, B. 1985: Geodetic measurements of surface deformations during the present rifting episode in NE-Iceland. JGR 90, 10163-10172.

Wilson, R. L. and McElinny, M. W. 1974: Investigation on the large scale paleomagnetic field over the last 25 million years. Eastward shift of the Icelandic spreading ridge. Geophysical Journal of the Royal astronomical Society 39, 570-586.

Winchester, J.A. and Floyd, P. A.,1977. Geochemical discrimination of different magma series and their differentiation products using immobile elements. Chemical Geology, vol. 20, pp. 325-343.