GEOCHEMISTRY AND GEOCHRONOLOGY OF THE EYJAFJÖLL VOLCANIC SYSTEM, ICELAND

CHAPTER 1

THE REGIONAL GEOLOGY AND TECTONICS OF ICELAND

Introduction

The work done for this research project has been an attempt to describe the temporal and geochemical development of the Eyjafjöll Volcanic System, in southeastern Iceland, and to investigate magmatic processes at the tip of a propagating rift. Eyjafjöll is located in southeastern Iceland, right along the coastline (Figure 1.01). Just 10 km offshore lies Vestmannæyar (the Westmann Island chain). The name Eyjafjöll means "mountain of the islands" and was given to this volcanic system because when viewing the mainland from Vestmannæyar, it is the prominent mountain.

Iceland is a 100,000 km^2 island which sits astride the Mid-Atlantic Ridge between the British Isles and Greenland. It is the largest volcanic island in the world and is located both at the center of a spreading ridge and above a hotspot. The general geology of Iceland is discussed throughout the rest of this chapter.

General Spatial and Chronological Relationships

Iceland is composed almost entirely of the following three rock types: basalt (80-85%), acidic and intermediate volcanic rocks (10%), and sedimentary rocks of volcanic and glacial origin (5-10%) (Sæmundsson, 1979). The ages of these rocks range from present day volcanic eruptions to the oldest exposed rocks, 16 Ma (Ibid). The ages correspond to location as discussed below and shown in Figure 1.02. The newest volcanism occurs in

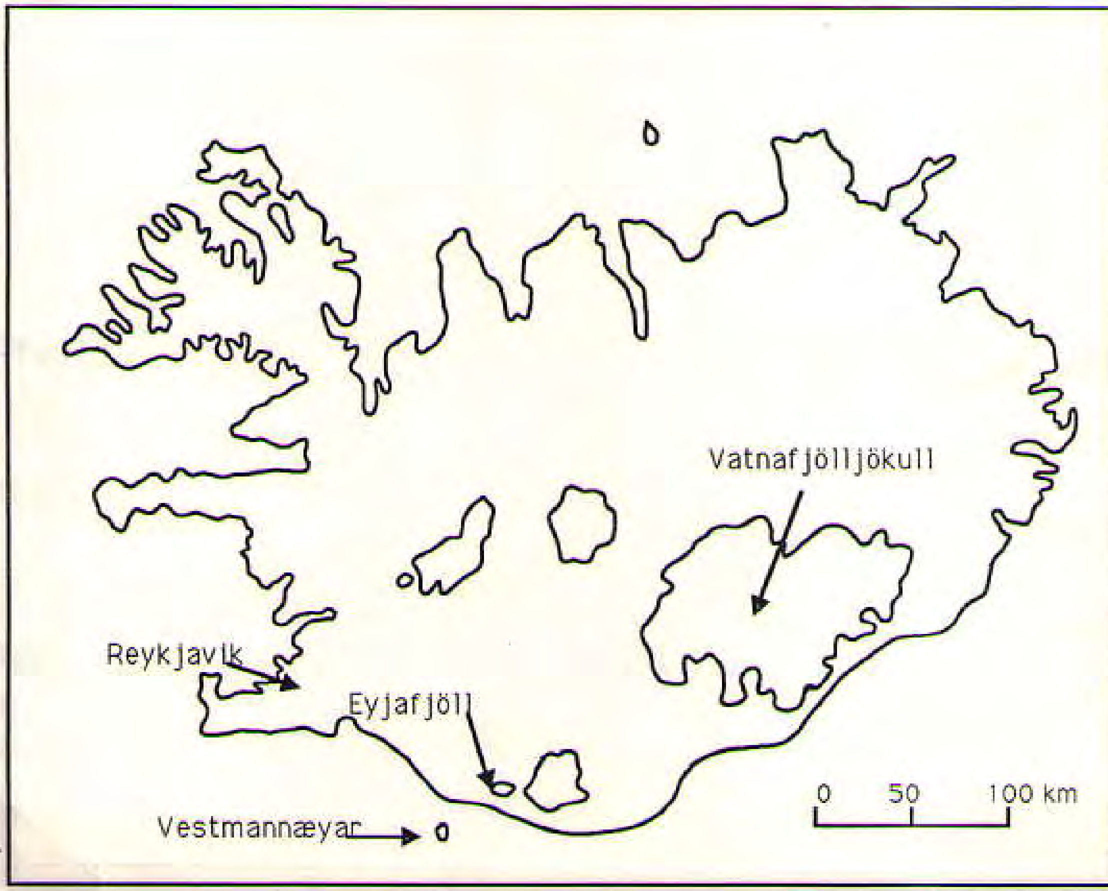

Figure 1.01: Map of Iceland, showing Eyjafjöll (topped by a glacier), Vestmannæyar, and Reykjavik. Enclosed regions within Iceland are glacier fields, such as the largest Vatnafjölljökull.

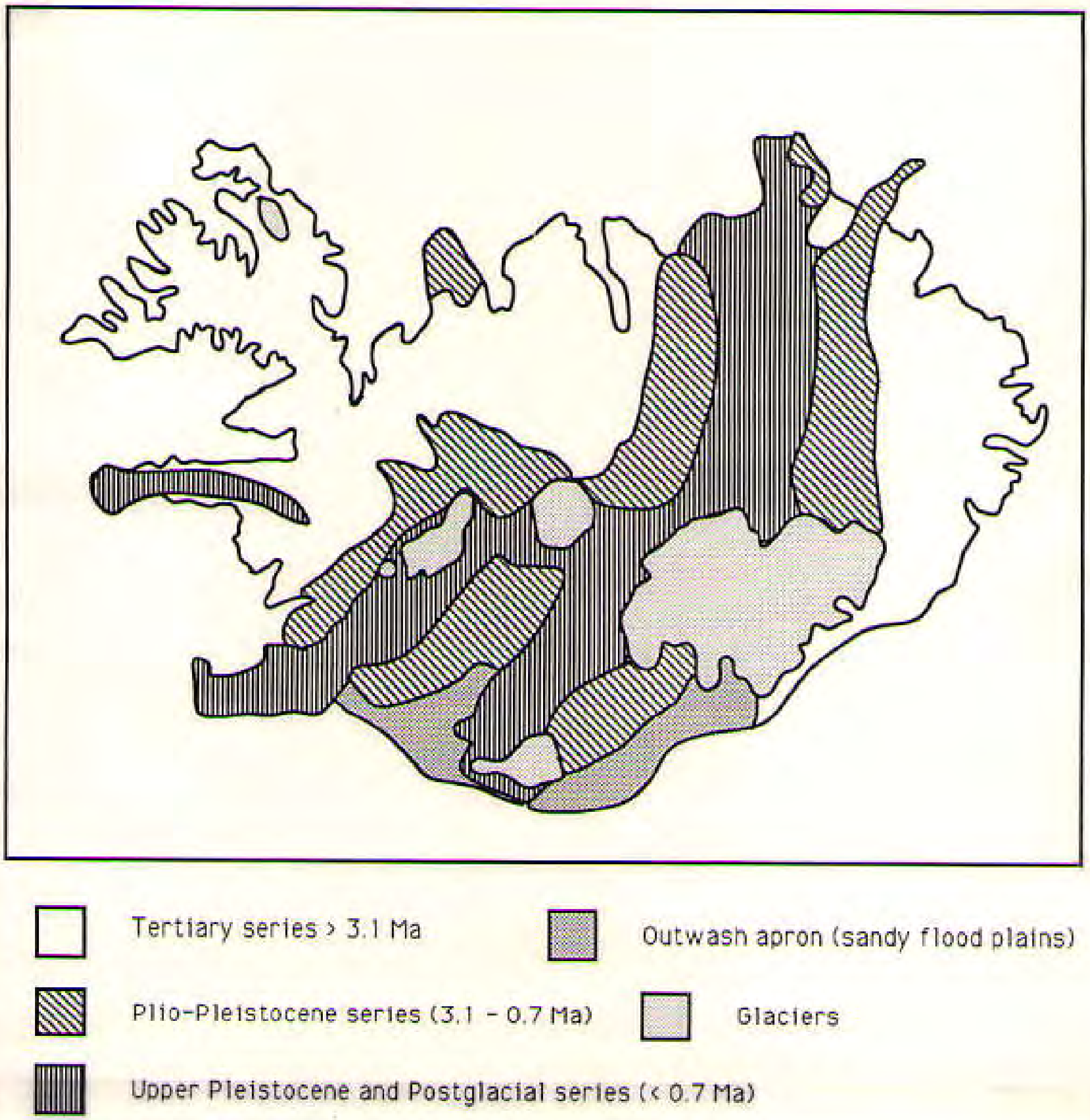

Figure 1.02: Map of Iceland showing the distribution of volcanism according to age, from the Tertiary to present.

distinct volcanic and rifted extensional zones which are named and shown in Figure 1.03. These zones are made up of rocks younger than 1 Ma (upper Pleistocene and postglacial sequences). Outward from these zones of activity the rocks get older, moving from the Plio-Pleistocene, 0.7-3.1 Ma, into the Tertiary sections (older than 3.1 Ma) which are exposed in the eastern and western margins of Iceland.

The Tertiary rocks constitute the majority of the eastern, northwestern, and northern parts of Iceland, with maximum ages of 13 Ma, 16 Ma, and 12 Ma respectively (Helgason, 1985; Schilling et al., 1982). These rocks cover about 50,000 km^2, 1/2 the total area of Iceland, and consist of tholeiitic lavas and genetically associated intermediate and acidic rocks (Sæmundsson, 1979). The older rocks consist of monotonous sequences of lava flows of approximately 5-15 m thickness separated by volcanoclastic beds. These rocks have been well studied and are discussed in full by Sæmundsson (1979).

The Plio-Pleistocene rocks cover an area of about 25,000 km^2, and show more lithologic variety than the Tertiary series. They include a higher number and variety of sedimentary beds (Sæmundsson, 1979). Instead of the primarily subaerial flows seen in the Tertiary, the Plio-Pleistocene units include rocks from subglacial eruptions (hyaloclastites) interstratified with the lava flows. This variation in stratigraphy can be explained by alternating warm and cold climates during which glaciers will correspondingly decrease and increase in size and change the style of eruptions (from subglacially to subaerially erupted lavas). Alternatively, if eruption rates are high enough and/or volumes of magma are great enough, the volcanic pile can grow through and then above the tops of the glaciers, forming pillow lavas, hyaloclastites, and then lava flows, in that order.

The upper Pleistocene rocks formed during the Brunhes magnetic period,

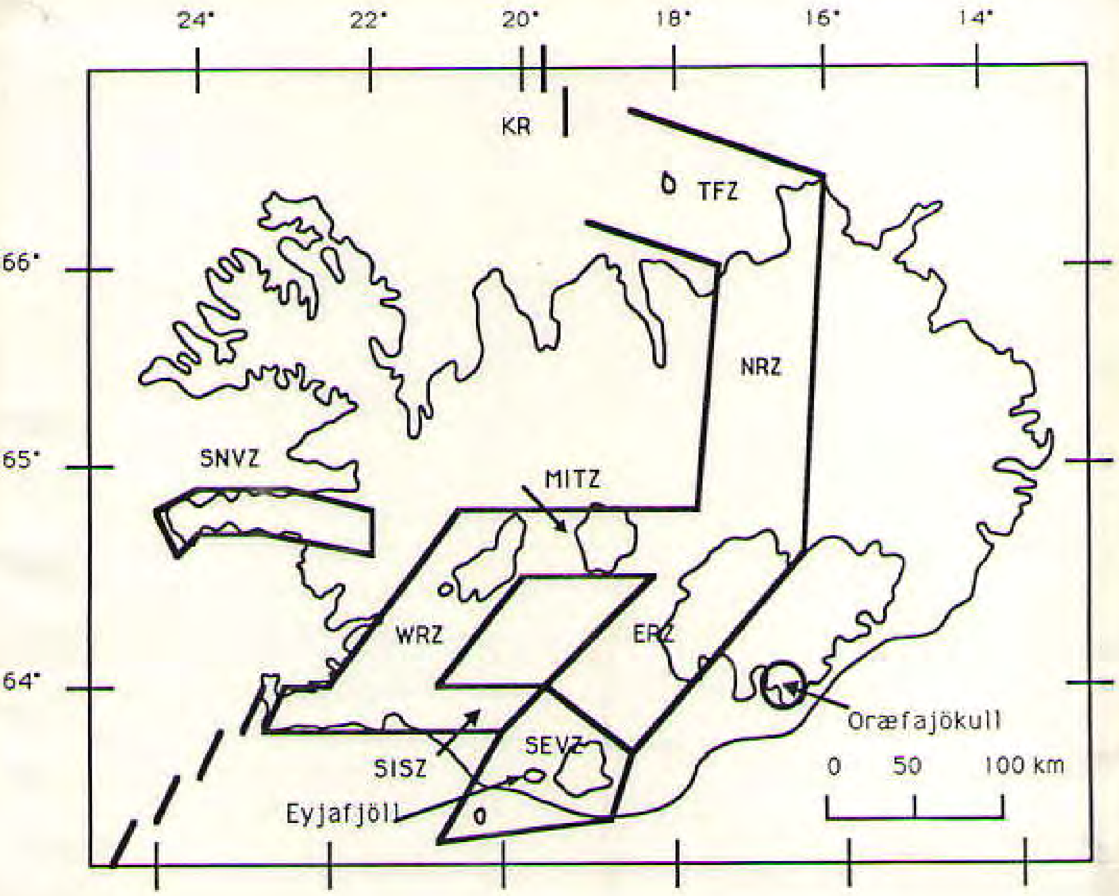

Figure 1.03: Map of Iceland showing the active volcanic and tectonic zones: Reykjanes Ridge (RR), Kolbeinsey Ridge (KR), Tjörnes Fracture Zone (TFZ), Northern Rift Zone (NRZ), Eastern Rift Zone (ERZ), Southeastern Volcanic Zone (SEVZ), Western Rift Zone (WRZ), Snæfellsnes Volcanic Zone (SVZ), Mid-Iceland Transform Zone (MITZ), South Iceland Seismic Zone (SISZ). Enclosed regions within Iceland are glacier fields.

from approximately 0.78 Ma up to the last glacial event (13-8 Ka) (Sæmundsson, 1979; Einarsson and Albertsson, 1988) and are exposed in the area of the active volcanic zones, covering about 30,000 km^2; they show alternating stratigraphy of subaerial and subglacial flows and deposits.

The postglacial rocks have all formed since the last glacial period which occurred 13-8 Ka, depending upon what part of Iceland is being studied (Einarsson and Albertsson, 1988; Sæmundsson, 1979). Because these rocks were never exposed to glacial erosion, most of the flows and deposits still retain their eruptive morphologies. These include surface features of various flows (e.g., pahoehoe, aa, blocky, etc.), and small, fragile, spatter cones aligned along active fissure zones. The total volume of these rocks is 400-500 km^3, covering about 12,000 km^2 of Iceland, over 10% of the surface (Sæmundsson, 1979). Figure 1.04 shows the variety of morphologic and tectonic features found among the active volcanic and rift zones in Iceland.

The general composition of the basalts in all four age groups is tholeiitic. Although some alkalic and transitional (between alkalic and tholeiitic) rocks do occur, they are localized to specific regions described as "flank zones". These are the Snæfellsnes Volcanic Zone, Oræfajökull, and the Southeastern Volcanic Zone. The volcanic system of Eyjafjöll is a member of the Southeastern Volcanic Zone and consists primarily of transitional and alkalic basalts and hyaloclastites younger than 800 Ka.

Hotspots - A Review

Iceland is located above a hotspot, and therefore it is relevant to review the characteristics typically attributed to these features. Hotspots are regions of increased heat flow and magmatism typically associated with mantle plumes. Mantle plumes are hypothesized to originate from deep in the mantle, perhaps

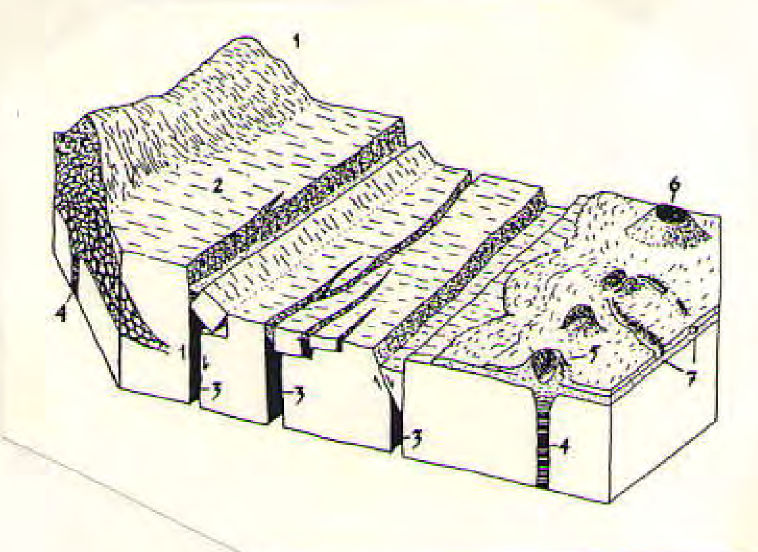

Figure 1.04: Block diagram illustrating the topographic expression of various types of dilational structures seen in the active volcanic zones of Iceland. 1, ridge formed by a subglacial fissure eruption; 2, postglacial lava field, partly covering 1; 3, non-eruptive fissures with local normal faulting; 4, dikes that have fed fissure eruptions; 5, aligned craters along a postglacial fissure; 6, cinder cone; 7, lava channels; (from Williams and McBirney, 1979).

as deep as the core-mantle boundary (Sleep 1990). They consist of high temperature, plastic material, which as a result of its lower viscosity and density (relative to the surrounding material) will rise upward due to buoyancy forces (Griffiths, 1986; Griffiths and Campbell, 1990; Loper and Stacey, 1983). The mechanics involved in this fluid movement have been modeled (Whitehead and Luther, 1975), and the results of analog laboratory experiments are shown in Figure 1.05. A large head (as wide as 1,000 km radius) carries the front of the plume upward, followed by a long, thin conduit that stretches from the head to the place of origin - the thermal boundary layer (Sleep, 1990; Stacey and Loper, 1983). Upon reaching the base of the lithosphere the large head will disperse laterally (Griffiths and Campbell, 1991) and partially melt. The magmas make their way to the surface and erupt in large volumes of superheated mafic material, often described as flood basalts. After the head has dispersed, the tail will continue to supply heat and magma to the surface until the anomaly has ceased.

Commonly known active hotspots and associated flood basalt provinces are: Kerguelen hotspot, Kerguelen plateau; Yellowstone hotspot, Columbia River Basalts; Louisville hotspot, Ontong Java plateau; Réunion hotspot, Deccan Traps (Sleep, 1990; Richards et al., 1989; Duncan and Richards, 1991). Associated with these hotspot regions are a number of characteristics which are quite distinct from other oceanic basalt regions, most notably from mid-ocean ridges. Not only are hotspots associated with higher heat flow, larger magma supply, and deeper melting, but they also have many distinct chemical characteristics.

As hotspots are fed from material deep within the mantle, they are associated with different isotopic signatures than MORB (Mid-Ocean Ridge Basalt) source material. A simplified model for how chemistry varies within

Figure 1.05: A photograph of a laboratory starting convective plume showing the structure caused by conduction of heat and consequent entrainment of surrounding fluid. Key features of the plume are labelled. Dark areas represent dyed material derived from the warmer source material, which becomes wrapped between layers of entrained surrounding material in a bulbous head. Hot source material is continuously supplied through the tail to the top of the plume head. The temperature contrast between the plume and the surrounding material governs the dynamics of the plume ascent, and this temperature contrast decreases as the volume of the head increases with time. The feeder conduit in these laboratory experiments is much broader, relative to the head, than is expected for mantle plumes because the viscosity of the hot conduit material in the experiments is only two orders of magnitude less than that of the surroundings, whereas the viscosity contrast in the mantle may be 10^{-2} to 10^{-5}. Much narrower conduits are observed in experiments with compositional density and viscosity contrasts (Campbell and Griffiths, 1990).

the mantle is described as follows, with a two-layer mantle. The lower mantle is thought to be primitive, and similar in composition to chondrites. The upper mantle is what remained after the melting of the original mantle and formation of the crust, an event which occurred about 3.8 Ga (Faure, 1977, 1986; Davies, 1990). The upper mantle is now depleted, (depleted of all incompatible elements that fractionated during melting) and is the source of MORB. In contrast, hotspot source material originates from the lower mantle.

These two mantle layers differ in composition as a result of their history. The undepleted source (lower mantle) has higher radiogenic Sr and lower radiogenic Nd than the depleted source (upper mantle). (As a reminder, a Rb isotope decays to a "radiogenic" Sr isotope, and a Sm isotope decays to a "radiogenic" Nd isotope). Rb is more incompatible than Sr so that the 3.8 Ga melting event will have removed Rb preferentially from the upper mantle, therefore reducing the amount of radiogenic Sr. This decreases the ratio of radiogenic Sr to stable Sr relative to the lower mantle (Figure 1.06). Similarly, Nd is slightly more incompatible than Sm, so that during the 3.8 Ga melting event, Nd will have been preferentially removed from the upper mantle, increasing the ratio of radioactive Nd to stable Nd. This ratio is therefore higher in the depleted source than the more primitive source, creating an inverse relationship between Nd and Sr ratios for oceanic material (Figure 1.07). This model is, however, too simple; the hotspot isotopic data are not uniform throughout the world as can be seen in this last figure, where the hotspots of Kerguelen, Hawaii and Iceland show different values. Apparently the lower mantle, though different from MORB source material, is not homogeneous. This will be further discussed below.

Pb isotopes work in a similar manner to the Sr and Nd isotopes, but show greater variation within all source regions. (As a reminder, ^{238}U decays to

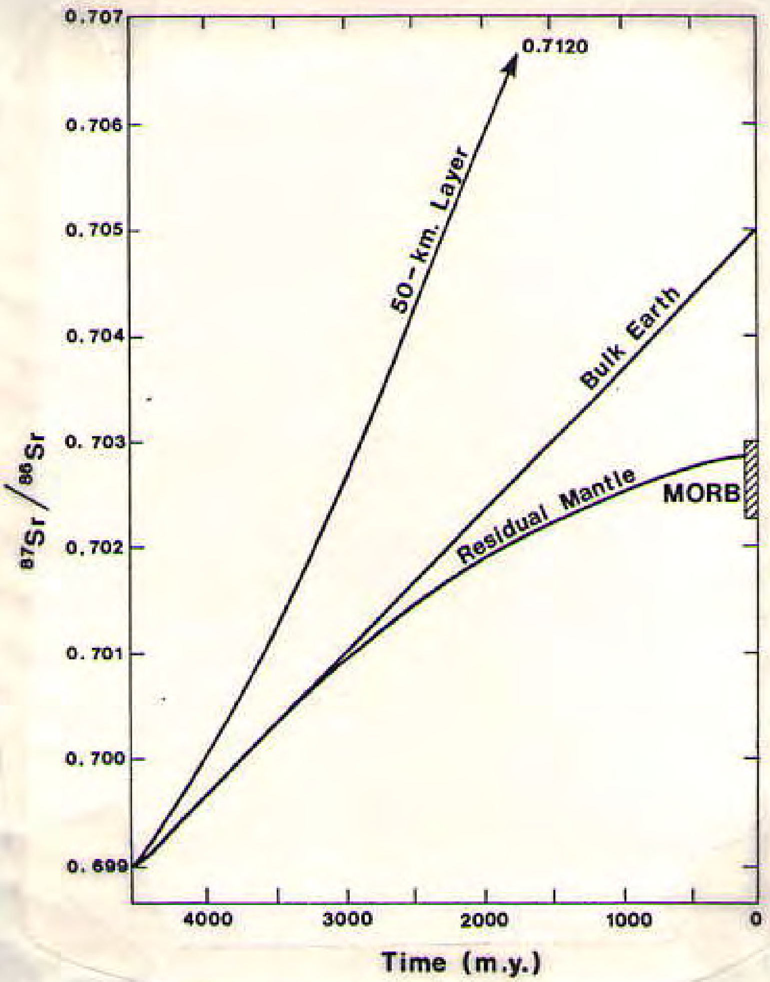

Figure 1.06: A model of the evolution of Sr ratios with time, assuming that half the Earth's mantle has had its low-melting constituents extracted to give the equivalent of a 50 km thick crust, (from Hall, 1987).

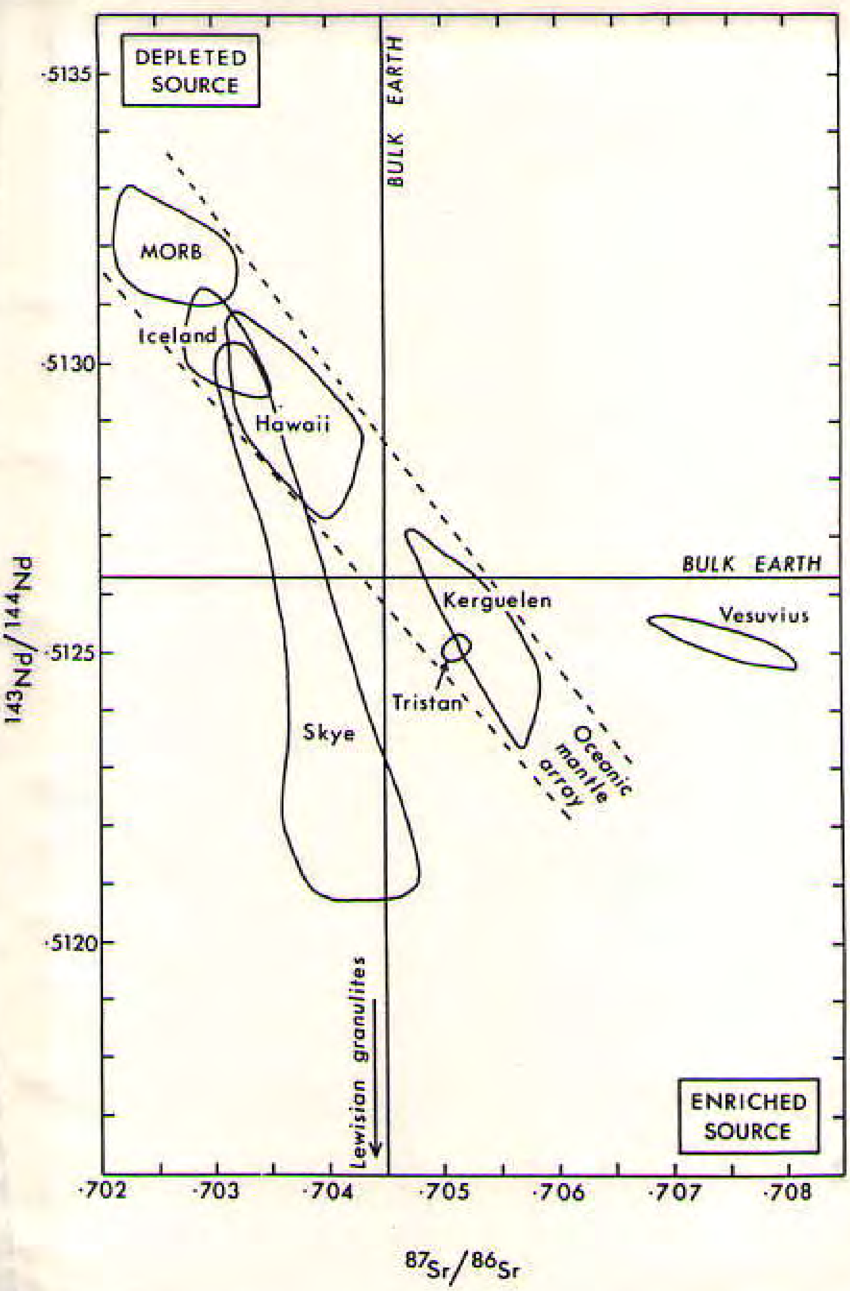

Figure 1.07: Plot of Sr ratios against Nd ratios for basaltic lava suites, (from Hall, 1987).

^{206}Pb; ^{235}U decays to ^{207}Pb; ^{232}Th decays to ^{208}Pb. All three of these systems combined are normally used for characterization of source material). Whenever U or Th are fractionated from one another, the resulting relationships of $^{208}Pb/^{232}Th$, $^{207}Pb/^{235}U$ and $^{206}Pb/^{238}U$ are changed. This will change the isotopic signature of the source region, and is usually described by different values of μ, $^{238}U/^{204}Pb$ ratios (reminder: ^{204}Pb is the nonradiogenic isotope), as shown in Figure 1.08. When plotting MORB and ocean island basalts on Pb diagrams, each region is characterized by a different value of μ. These values even show variation among islands in the same group, indicating heterogeneity of the source region on a local and world-wide scale, (Figure 1.09). Despite this heterogeneity, however, material derived from deep in the mantle, indicative of hotspots, shows Pb ratios more enriched in ^{206}Pb and ^{207}Pb than the MORB source material. This occurs because uranium is more incompatible than thorium and will have been removed from the depleted mantle during crust formation.

According to this model of early mantle differentiation, one would expect that the most incompatible trace elements will be of a higher concentration in the lower mantle than in the upper mantle. These incompatibles include the large ion lithophile elements such as K, Rb, Cs, Ba, U, and Th. Light rare earth elements are more incompatible than the heavy rare earth elements and so these should be relatively depleted in the upper mantle. Finally, $^{3}He/^{4}He$ values are higher from the lower mantle as primordial ^{3}He is lost more easily from the upper regions (Davies, 1990) while in the lower mantle it is constantly being fed by degassing of the core.

As mentioned earlier, this two-layer model of mantle dynamics is too simplified. The lower mantle is not homogeneous evidenced the variation in isotopic signature among different hotspot regions (Fisk et al., 1988;

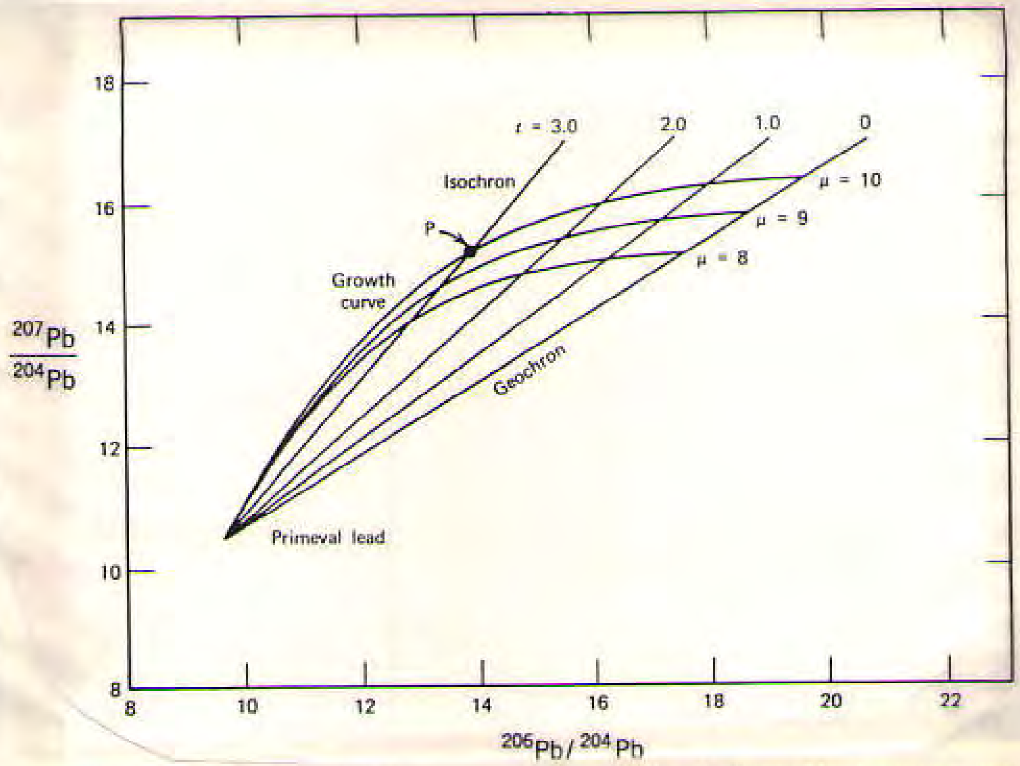

Figure 1.08: Graphical representation of the Holmes-Houtermans model. The curved lines are lead growth curves for U-Pb systems having present-day μ values of 8, 9, and 10. The straight lines are isochrons for selected values of *t*. For example, the coordinates of point P are the Pb ratios of a galena lead that was withdrawn 3.0 x 10^9 years ago from a source region whose present μ is 10.0 (from Faure, 1977, 1986).

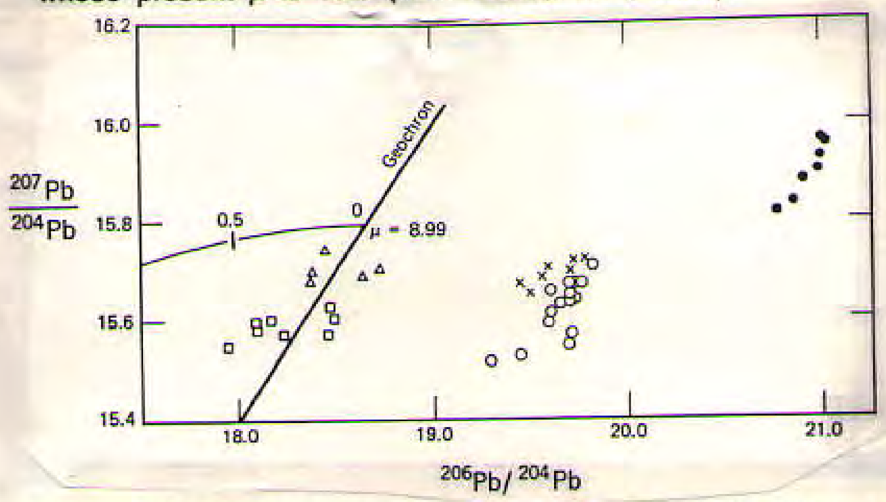

Figure 1.09: Isotope ratios of Pb from young volcanic rocks. Crosses = Ascension Island, triangles = Gough, open circles = St. Helena, closed circles = Tenerife, Canary Islands, squares = Hawaiian Islands. The data show both intra- as well as interisland differences in the isotopic composition of Pb. These variations are evidence for the heterogeneity of U/Pb ratios in the upper mantle and indicate that the source regions of these volcanic rocks have not remained closed systems throughout geologic time (from Faure, 1977, 1986).

Richards et al., 1991). Undoubtedly there must be interaction between the upper and lower mantle and the continental and oceanic crust. In order to explain mantle heterogeneities, which can exist on as small a scale as tens of kilometers (White and Hofmann, 1982a, 1982b), the following theories have been proposed: (1) mixing of the depleted and primitive materials, (2) addition of crustal material (as in subducting lithosphere) with varying amounts of seawater alteration, (3) varying amounts of magmatic depletion of the originally homogeneous mantle during crustal formation, (4) addition of metasomatic fluids (White and Hofmann, 1982a, 1982b; Patchett et al., 1984; Hofmann et al., 1986). Each of these mechanisms has been used to explain isotope and geochemical variations for different oceanic islands (Hotspot regions), but one mechanism alone has been insufficient to explain them all.

<u>Iceland Hotspot</u>

Beneath the crust of Iceland there is a large region of increased heat flow and magma supply associated with the mantle hotspot, with excess temperatures of upwelling magma between 200°C to 250°C (Sleep, 1990; Pálmason and Sæmundsson, 1979). This makes Iceland similar in many respects to other hotspot-fed islands such as Hawaii. Unlike the Hawaiian Island chain, which lies in the middle of a plate the Iceland hotspot is at the center of the diverging plates of the Mid-Atlantic Ridge. This allows for large volumes of lava to accumulate, and the size of Iceland to grow, as the diverging plates move older material away. This is especially important since the rate of divergence is quite slow, about 2 cm/yr with a volume flux of magma of 8 m^3/s (Sleep, 1990; Sæmundsson, 1979). Hawaii has a volume flux of magma of only 4 m^3/s. Plume material melts much more extensively in Iceland than Hawaii, ten times more, because it ascends to shallower depths (Ibid). The increased magmatism and slow spreading rate helps to increase Iceland's size relative to other hotspot volcanic chains elsewhere in the world. (In comparison, Table 1.01 shows data from other hotspot regions).

The initiation of the Iceland hotspot most likely began with the eruption of the flood basalts of the North Atlantic Tertiary Basalt Province, beginning about 62 Ma and peaking at about 59 Ma (White, 1989). Remnant sections of this Paleocene flood basalt province, in the Skaergaard region of Greenland, the Faeroes Islands, the Vøring Plateau and the British Tertiary Province in Scotland and Northern Ireland, show similarities in composition, structure and age (Figure 1.10, White, 1989; White et al., 1987; Nunns, 1981; Vink, 1984; Vink et al., 1985). These regions could have been erupted during one flood basalt episode, concurrent with the opening of the present North Atlantic Ocean (Ibid). The present day Greenland-Iceland Ridge and the Iceland-Faeroes Ridge (Figure 1.11) formed as hotspot traces when spreading of the North Atlantic continued, and initial magma flux decreased.

It appears that this hotspot has been a long-lived, stationary thermal feature of the upper mantle beneath the North Atlantic. It continues today, where it presently sits under south central Iceland, and shows the following anomalies. (1) Iceland is an area of much higher elevation than elsewhere along the Mid-Atlantic Ridge (with its highest point at 2000 meters above sea level). It consists of oceanic crust with a thickness of 4 to 20 km compared to the surrounding crust which is 2-3 km thick (Helgason, 1985; RRISP 77, 1980; Meyer et al., 1985). (2) The Mid-Atlantic Ridge axis is offset in Iceland to the east relative to the northern, Kolbeinsey Ridge and southern, Reykjanes Ridge, in the direction of the hotspot location. (3) The intensity of volcanism decreases along the axial rift zones away from south central Iceland

Table 1.01: Buoyancy, volume, and magma volume fluxes for specific hotspot localities throughout the world.

	Buoyancy Flux = ß (Mg/s)	Volume Flux Flux = Qp (m3/s)	Magma Volume Flux = Qv (m3/s)
Hawaii	8.7	300	4
Easter	3.3	140	-
Réunion	1.9	85	0.77
Tristan	1.7	200	-
Iceland	1.4	63	8
Afar	1.2	53	-
Azores	1.1	51	-
Bermuda	1.1	48	-
Kerguelen	0.5	22	0.007

ß = p(m)*alpha*ΔT*Qp
Qp = V(s)*(L+A/2)*Y

p(m) = density of the mantle
alpha = thermal expansion coefficient
ΔT = average excess temperature
V(s) = full spreading rate
L, A = thickness of lithosphere and asthenosphere away from the ridge.
Y = along-strike distance supplied by the plume

(Data and equations are from Sleep, 1990, and Gerlach, 1990)

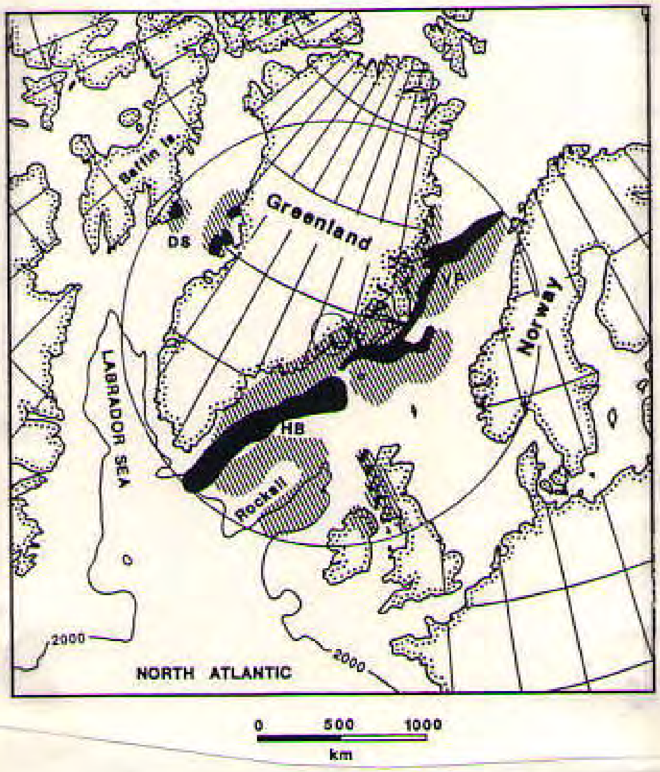

Figure 1.10: Reconstruction of the North Atlantic at magnetic anomaly C23, just after the onset of sea-floor spreading. Positions of extrusive volcanic rock are shown by solid shading, and hatchuring shows extent of early Tertiary igneous activity in the region. Inferred position of the Iceland hotspot mantle anomaly is shown by circle. DS = Davis Strait, VP = Vøring Plateau, HB = Hatton Bank. Projection is equal-area, centered on position of mantle plume. Bathymetric contours in meters (from White, 1989).

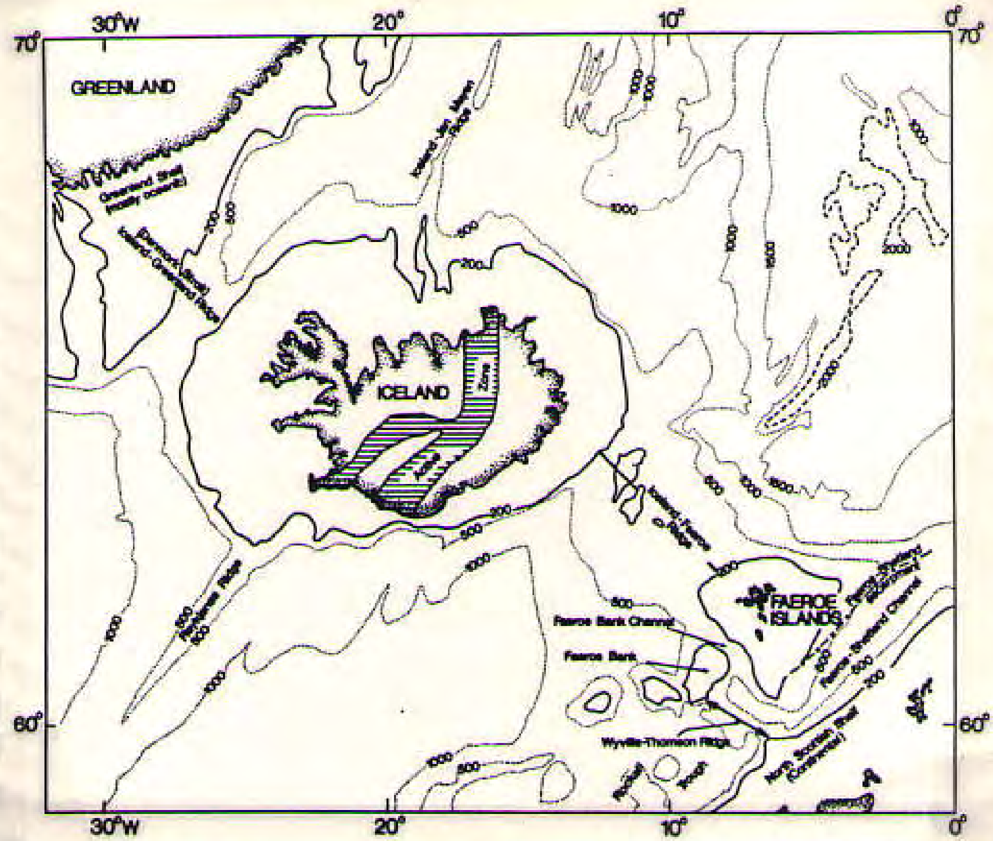

Figure 1.11: The morphology of the Greenland-Scotland ridge (from Bott, 1981).

(Imsland, 1983). (4) The composition of the magmas erupted in Iceland are different from those of the rest of the Mid Atlantic Ridge. The concentrations of the large ion lithophile (LIL) elements (K, Rb, Cs, Ba, U, Th) and light rare earth elements (LREE) in basaltic rocks increase along the Mid-Atlantic Ridge towards Iceland along with Ti and P (Schilling, 1973; Imsland, 1983; Oskarsson et al., 1985; Elliott et al., 1991). (5) Radiogenic Sr and ^{206}Pb increase towards Iceland while radiogenic Nd decreases (Ibid). (6) Starting with typical Mid-Atlantic Ridge ratios of ≤ 8 x atmospheric, $^{3}He/^{4}He$ values increase towards Iceland and near the location directly above the hotspot they reach values as high as 26 x atmospheric (Kurz et al., 1985; Poreda et al., 1986). (7) Geophysical evidence from Iceland shows an increased heat flow under the region and a deeper and wider domain of partially melted mantle material down past 50 km and believed to continue to 250 km (RRISP 77, 1980). There is a low velocity domain extending below Iceland from 0 to 75 km. This zone narrows between the depths of 75 to 125 km and then widens again to reach depths below 375 km, (Figures 1.12, 1.13, and 1.14, Tryggvason et al., 1983). (These low velocities for P-waves, between 7.0 to 7.8 km/s, are indicative of molten mantle material).

Dynamics of Mantle Melting

As Iceland is located in such a unique tectonic setting, with two different potential source compositions, that of the hotspot (relatively undepleted mantle) and that of the Mid-Atlantic Ridge upper mantle (depleted), any model for mantle dynamics beneath Iceland must describe the interaction between the two. Ultimately, magma from the hotspot source must make its way through upper depleted mantle in order to reach the surface. Just how this migration occurs, and what the compositional effects of this are, is the basis for a number

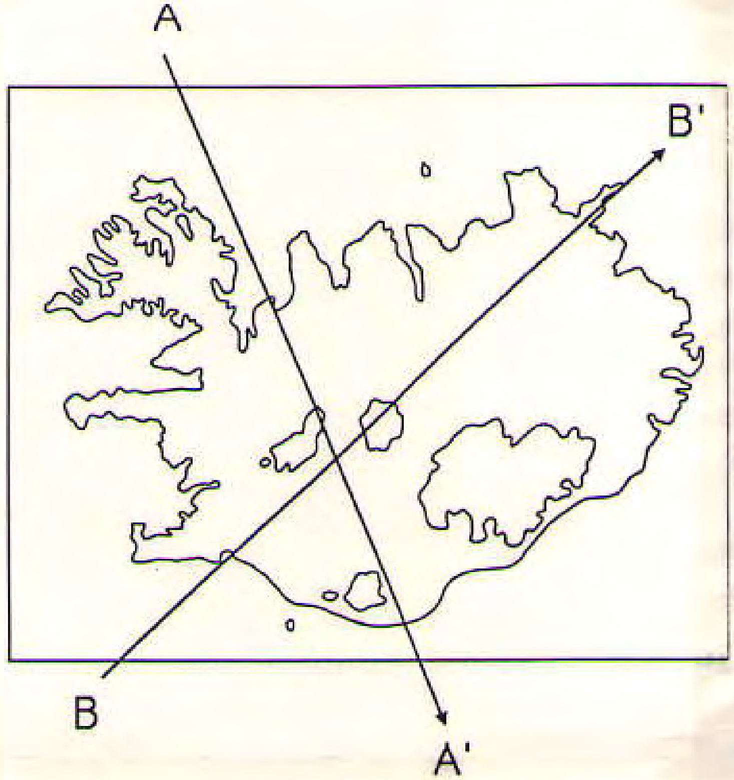

Figure 1.12: Cross section lines for figures 1.13 and 1.14.

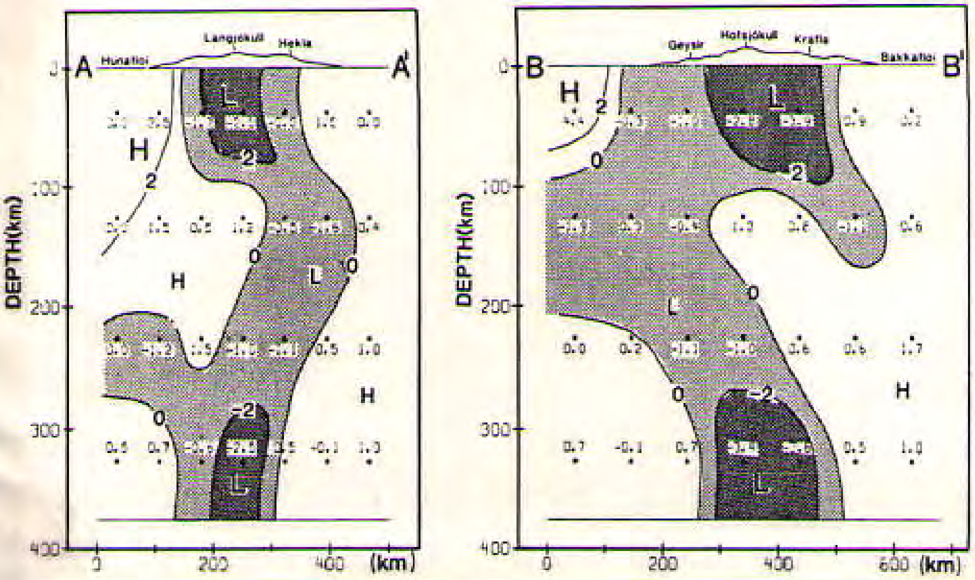

Figure 1.13: Cross-sections spanning latitude range from 63.0°N to 66.0°N, from NW to SE (A-A') and from SW to NE (B-B'). Capital letters indicate areas of high (H) and low (L) velocities, while numbers indicate P-wave velocity perturbations (%) (Tryggvason, 1983).

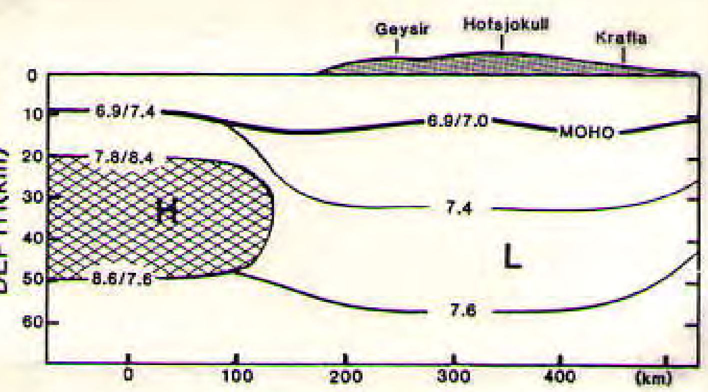

Figure 1.14: Generalized crust and upper mantle cross-section derived from the RRISP main profile. The strike of the profile roughly coincides with the B-B' cross-section of figure 1.11 (modified by Tryggvason, 1983, from RRISP 77, 1980).

of different models (Meyer et al., 1985; Schilling et al., 1982; Wood et al., 1979; Wood, 1981; Zindler, et al., 1979; Hemond et al., 1988). Inherent to these models is the need to explain the variations in lava composition throughout Iceland, from major element variations to differences in isotopic content, (Figures 1.15 and 1.16, Elliott et al., 1991). Geographical variations are seen, and with them the proposed source material is thought to be spatially heterogeneous (Zindler et al., 1979; Wood, 1981). Elliott et al. (1991), explain the chemical variations as the result of mixing of magmas formed at different pressures within the mantle. This would explain why FeO varies in the high MgO rocks. Another explanation for the observed variations is the result of mixing of plume (undepleted mantle) and normal MORB derived upper mantle, in varying degrees (Meyer et al., 1985; Kurz et al., 1985; Wood, 1981; Schilling et al., 1982; Sun et al., 1975).

It has been suggested that crustal contamination may play a major role, especially in forming the alkali basalts seen in select regions of Iceland by melting of amphibolite lower crust (Oskarsson et al., 1982, 1985) which has undergone past hydrothermal alteration and underplating (Hemond et al., 1988; Steinthórsson et al., 1987). More recent work, however, has shown that this is not necessary, that there are other methods of producing these alkali basalts, such as greater depths of melting and smaller percentages of partial melt (Jaques and Green, 1980; Stolper, 1980; Thy et al., 1991b). Helium isotope data (Kurz et al., 1985; Hilton et al., 1990) suggest that assimilation can't provide the observed helium ratio variations, but that changes in mantle source composition (specifically in terms of varying amounts of undepleted and depleted mantle) can. This will be discussed further in Chapter 2.

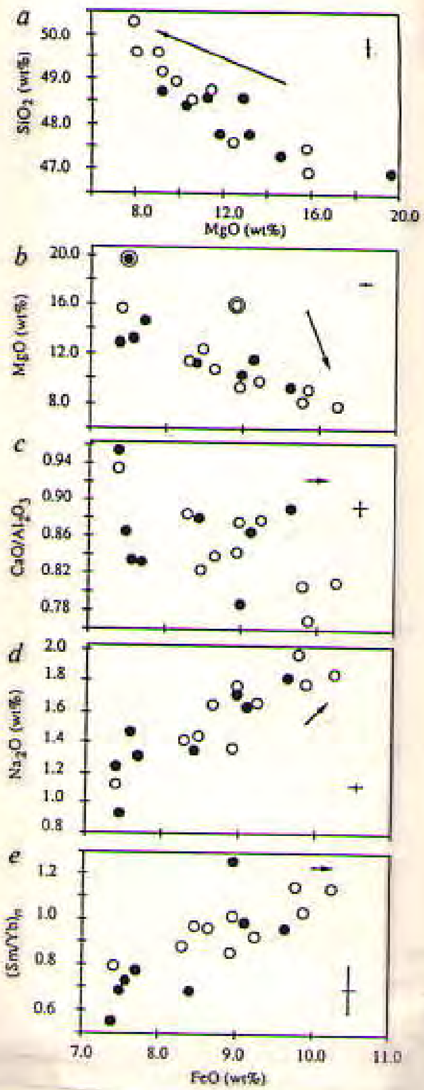

Figure 1.15: Major and trace element variation within Iceland (Elliott et al., 1991). Circled symbols in (b) are petrographically identified as olivine cumulates. Sm/Yb in (e) is normalized to chondritic abundances. Closed symbols are samples from the Reykjanes Peninsula; open symbols are from Theistareykir. Vectors show the effects of 15% fractional olivine crystallization.

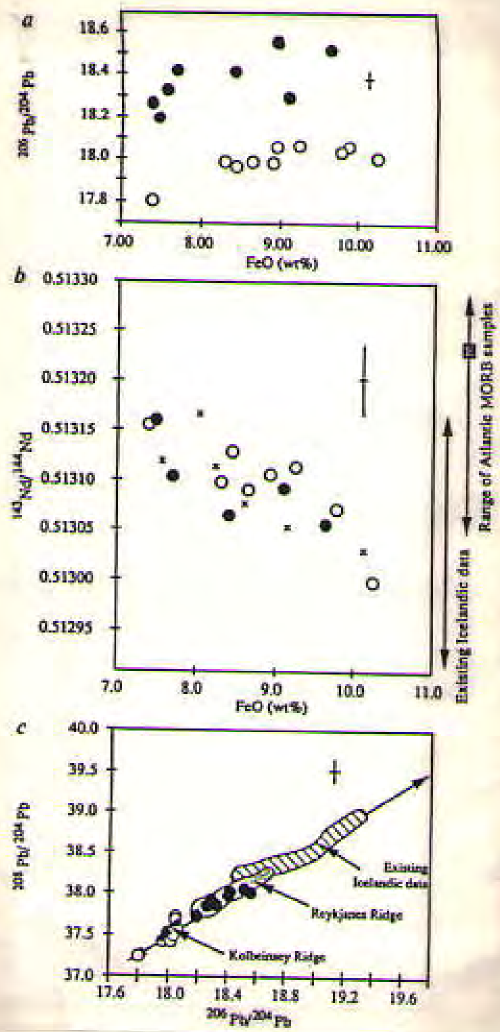

Figure 1.16: Pb and Nd variation within Iceland. (b) Crosses show data for other similar Reykjanes Peninsula lavas. The shaded box indicates Nd ratios for MORB endmembers for Reykjanes Ridge and Kolbeinsey Ridge.
(c) The hatchured region shows existing data for post-glacial Icelandic basalts; the shaded region shows data for the submarine Reykjanes Ridge basalts, and the bar shows the range of lead isotope values for samples from the Kolbeinsey Ridge. The double-headed arrow shows the range of lead isotope values for Atlantic MORB and is aligned to give the best fit to the data. Symbols for Reykjanes Peninsula and Theistareykir suites are as in Figure 1.14 (from Elliott et al., 1991)

Mid-Atlantic Spreading Ridge

The Mid-Atlantic Ridge extends north of Iceland with the Kolbeinsey Ridge, and south with the Reykjanes Ridge and crosses Iceland with the presently active volcanic zones which, for the most part, are described as active axial rift zones (Figure 1.03). The Western Rift Zone extends north from the Reykjanes Ridge and terminates where it intersects the Mid-Iceland Transform Zone. The Northern Rift Zone connects with the Kolbeinsey Ridge by the Tjörnes Fracture Zone. It extends south, past the Mid-Iceland Transform Zone, as the Eastern Rift Zone, the southernmost extension of which is the Southeast Volcanic Zone. The Eastern Rift Zone and the Western Rift Zone run parallel with one another and are believed to be a pair of propagating and dying rifts respectively. They are joined again at the southern ends where the South Iceland Seismic Zone connects the Reykjanes Peninsula with the Southeast Volcanic Zone.

A propagating rift is one of a pair of parallel spreading ridge segments which connect with one another by means of a transform fault zone (Figure 1.17). It is an indicator of movement of the ridge axis away from its previous position. The originally continuous ridge has been separated into two segments and then displaced across this separation by a transform fault. One segment, the dying or failing rift, is slowly declining in volcanic activity as the source material is being drained away. The other segment, the propagating rift, is increasing in activity as it taps into the magma supply of the dying rift. The dying rift will cease activity starting from its tip and moving backwards, while the propagator moves forward and increases in activity, eventually surpassing the dying rift entirely. The front tip of the propagator is where the newest volcanism occurs, through older crust, while the tip of the dying rift has the least amount of volcanism and eventually becomes extinct.

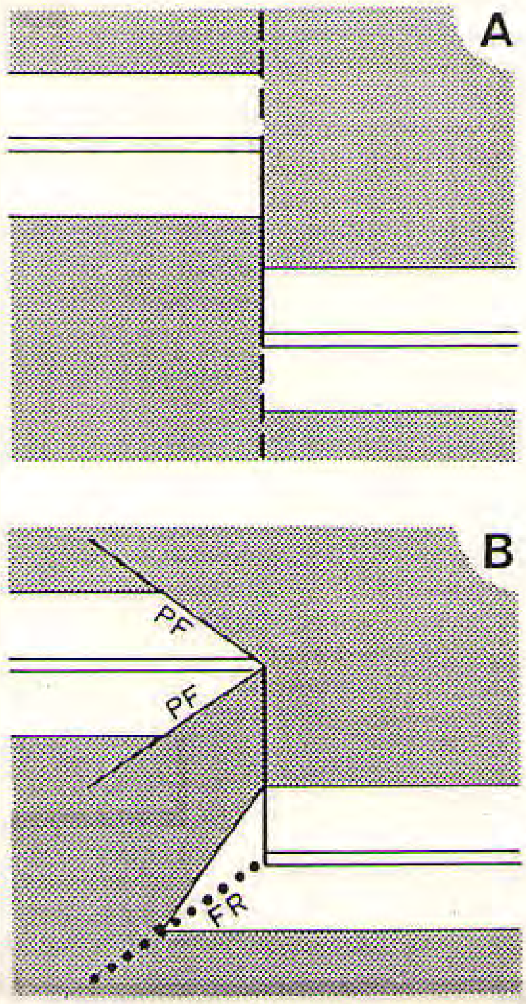

Figure 1.17: Schematic representation of normal ridge (A) and propagating ridge (B) transform intersections. Stippled areas denote all crust older than the axial anomaly which is unpatterned. Tectonic features specific to propagating rift systems include pseudofaults (PF) and a failed rift (FR). Note that propagating rift segments are bounded by old crust across the pseudofaults as well as across the transform (from Sinton et al., 1983).

The zone that joins up the two segments is characterized by "pseudofaults" (Figure 1.17) which can be seen by changes in the magnetic anomalies, (Sinton et al., 1983; Phipps Morgan and Kleinrock, 1991; Calvert et al., 1990). The Eastern Rift Zone is thought to be a propagating rift, and the Western Rift Zone the corresponding dying or failing rift. They both run parallel to one another and the Eastern Rift Zone has a higher magma supply rate. The tip of the Eastern Rift Zone, the Southeastern Volcanic Zone, appears to be propagating forward, where new, young volcanism is occurring in Vestmannæyar (the Westmann Islands, off the southeast shore of Iceland). The South Iceland Seismic Zone corresponds to the southern transform fault zone connecting the Eastern and Western Rift Zones. It will discussed in more detail, later on in this chapter.

The entire Iceland plate boundary system appears to be moving west-northwest relative to the hotspot, resulting in relocation of the rift zones further east. This is most evident when comparing the plate's present location with its past location, when initiation of the hotspot began, in what is now eastern Greenland, (Figure 1.10).

Maximum extension in the rift zones occurs perpendicular to the spreading ridge axes. These zones are dominated by linear extensional tectonic features such as dike and normal fault swarms, and fissure eruptions (Figure 1.04). The axes are dotted with a number of central volcanic systems. Study of the presently active volcanic system of Krafla, in the northeast of Iceland gives a spreading rate for the Northern Rift Zone of 2 cm/year (Sæmundsson, 1979). The topography of the rift zones also consists of broad valleys bordered by taller, older, ridges which slope down to the valley floor via down-dropped normal faulting, (well-developed axial valleys).

Active seismic regions such as the Mid-Iceland Transform Zone and the South Iceland Seismic Zone connect the axial rifting zones with shear zones characterized by oblique faulting (Einarsson and Björnsson, 1979; Einarsson, 1991). The Mid-Iceland Transform Zone consists of NW-SE trending wrench faults and N-S grabens, with volcanic fissure swarms in an en echelon pattern (Sæmundsson, 1979). The combined movement is suggestive of a right-lateral transform zone (Einarsson and Björnsson, 1979; Einarsson, 1991). The South Iceland Transform zone is indicative of a left-lateral transform connecting the Reykjanes Ridge in the west to the Eastern Rift Zone and Southeast Volcanic Zone in the east (Ibid). It will be discussed in more detail in the next section. The seismic zones seen in Iceland are often described as subaerial analogs to the transform faults seen along the rest of the Mid-Atlantic Ridge (Sæmundsson, 1979; Einarsson and Björnsson, 1979; Hackman et al., 1990; Einarsson, 1991).

Other zones of present day volcanic activity that don't have the same tectonic features and/or chemical composition as the previously mentioned ones are termed "flank zones". These consist of the Snæfellsnes Volcanic Zone, the Southeastern Volcanic Zone, and Oræfajökull, and have been described, by Sæmundsson (1979) as follows. Snæfellsnes Volcanic Zone, and the Southeastern Volcanic Zone erupt lavas which are transitional to alkalic in composition, as opposed to the primarily tholeiitic composition of the rift zones. All three zones lack well-developed extensional features. The Snæfellsnes Volcanic Zone has yet to have any successful tectonic interpretation though it has been referred to in the past as an old transform zone (Steinthórsson et al., 1985). It is characterized by a large volcanic edifice, Snæfellsjökull, and has experienced faulting in a trend nearly perpendicular to that of the Western Rift Zone. Oræfajökull is 40 to 50 km east of the main

Eastern Rift Zone. It is 2119 m high and nearly circular, with no fissure swarms. It overlies older basement and is right along the edge of the center of the hotspot. It erupts transitional and tholeiitic lavas. The Southeastern Volcanic Zone is believed to be the propagating tip of the Eastern Rift Zone, and is discussed further below.

Southeastern Volcanic Zone

The Southeastern Volcanic Zone (SEVZ) consists of the volcanic systems of Hekla, Vatnafjöll, Torfajökull, Tindfjöll, Katla, Eyjafjöll, and Vestmannæyar (Figure 1.18). This zone continues southward from the Eastern Rift Zone where it connects with the South Iceland Seismic Zone (Figure 1.19). The volcanic systems display no "well-developed" axial valleys like the rift zones do, though faults and fissure swarms are present. All faults trend in a direction which is subparallel to the rifting of the Eastern Rift Zone. In particular there appear to be sets of conjugate shear faulting with NNE-SSW and ENE-WSW trends, especially prominent in the more northern systems such as Hekla. The orientation of the volcanic axes of the systems in this zone trend in an ENE-WSW or E-W direction, the former direction describing Hekla, Vatnafjöll, Torfajökull, and Katla (more northerly) whereas the latter direction describes Eyjafjöll and Tindfjöll (towards the south).

The SEVZ seems to derive its tectonic features from a combination of Eastern Rift Zone spreading and left-lateral stress along the South Iceland Seismic Zone, with maximum tension subparallel to the spreading direction of the Eastern Rift Zone (Sæmundsson, 1979). The strike of the linear features (such as fissure swarms, and faults) changes southward, from ENE-WSW to E-W, coinciding with a variation in rock composition from tholeiitic in the northeast through transitional to alkalic in the south. There is also a decrease

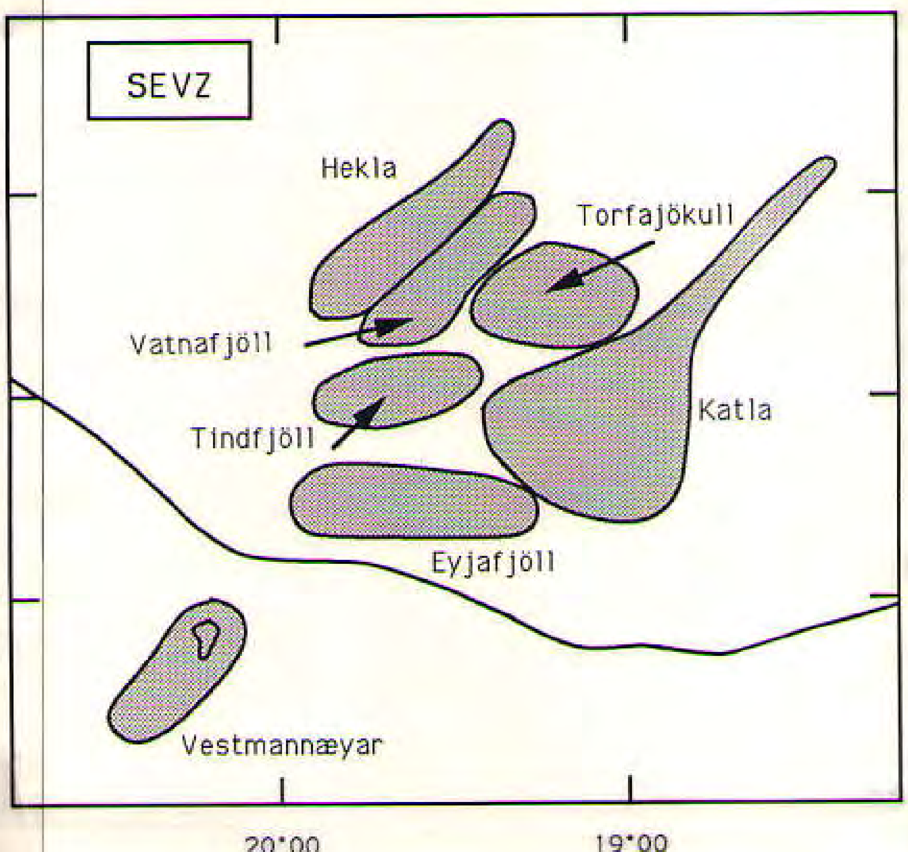

Figure 1.18: Southeastern Volcanic Zone (SEVZ), with the volcanic systems (dotted fields) of Hekla, Vatnafjöll, Torfajökull, Tindfjöll, Eyjafjöll, Katla, and Vestmannæyar (after Jakobsson, 1979).

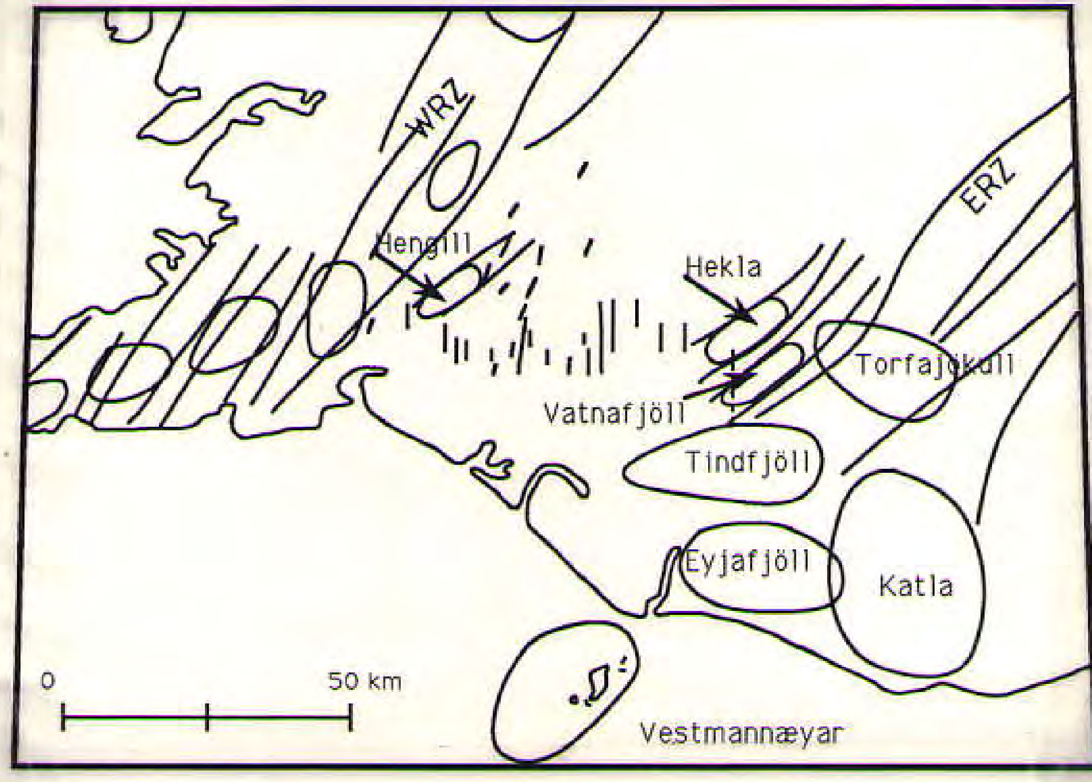

Figure 1.19: The South Iceland Seismic Zone (SISZ) and the Southeastern Volcanic Zone (SEVZ) showing mapped surface breaks and regions in which over half of the buildings were destroyed in historic seismic events. The north-south dashed line near Vatnafjöll indicates the estimated location of the fault on which the May 25, 1987 earthquake occurred (from Hackman et al., 1990). The Eastern Rift Zone (ERZ) and Western Rift Zone (WRZ) continue northward of this region. Circled regions are outlines of volcanic systems.

in volcanic productivity southwards. The volcanic systems show no rifting (spreading) and sit on crust from 10 to 20 km thick, some of the thickest crust in Iceland (RRISP 77, 1980).

South Iceland Seismic Zone

The South Iceland Seismic Zone (SISZ) seems to be the subaerial analog to a transform fault. It consists of SW-NE trending en echelon fissures in the west near the Reykjanes Peninsula, moving through N-S trends to ENE-WSW trends in the east (Figure 1.19, Einarsson, 1991). Comparison with previously published literature on "bookshelf" faulting (Phipps Morgan and Kleinrock, 1991) provides a means for interpreting the activity in this zone. "Bookshelf" faulting is a term applied to en echelon faults by which shearing is accommodated with strike-slip faulting and rotation at a high angle to the overall shear direction. Shortening is accommodated by reverse faulting and rotation of normal faults (Figure 1.20). In the SISZ, there is no morphological expression of a throughgoing transform fault, only a series of north-south trending strike slip faults. Phipps Morgan and Kleinrock (1991), suggest that this transform zone might migrate through "bookshelf faulting". Using this explanation, the north-south trending faults are a method of reducing east-west lateral stress.

Hackman et al. (1990) compared seismic moment release data derived from earthquake magnitudes in the SISZ to their own models and found an even greater amount of transform activity than was predicted, suggesting not only greater seismicity, but also a greater depth for the brittle layer, up to 14 km, rather than 10 km as previously noted.

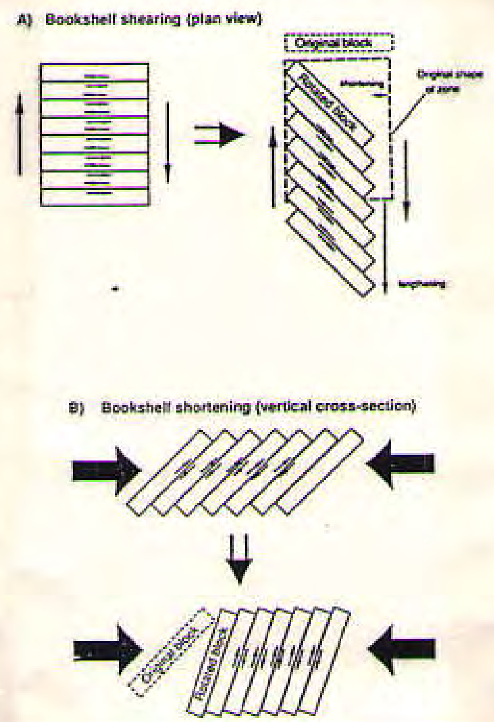

Figure 1.20: Bookshelf faulting. a) Bookshelf shearing is accommodated by strike-slip faulting and rotation of en echelon faults which are at a high angle to the overall shear direction. Large arrows show overall shear direction, and small arrows show local slip directions on faulting between individual "books". Note the strain incompatibility produced by this process (lengthening in overall shear direction and shortening perpendicular to this direction). b) Bookshelf shortening can be accommodated by reverse faulting and rotation of en echelon normal faults (from Phipps Morgan and Kleinrock, 1991).

Mechanics of Axial Spreading

The mechanics of the axial spreading in Iceland are much debated with two main contending models: a continual movement away from the central axis, and a discontinuous movement occurring as jumps of the axis. The Tjörnes Fracture Zone which connects the Kolbeinsey Ridge, north of Iceland, with the Northern Rift Zone is believed to have formed about 4 Ma (Sæmundsson, 1974). This zone evolved due to the jump of the rift axis from its former location north of the Western Rift Zone (along that zone's present strike up to Skagi) to its present position. This boundary shift is evidenced by an unconformity of 4 Ma lavas over the older Tertiary lava pile along the present day Northern Rift Zone, and by evidence of older transform zone fault scars where the old Northern Rift Zone was proposed to have been. Most geologists accept this jump of the plate boundary, but maintain that it is a rare occurrence and that steady-state activity is characterized by a stationary boundary (Pálmason, 1980).

An alternative theory of rift mechanics suggests that the boundaries of the plates shift frequently (Helgason et al., 1984, 1985). This idea was developed in order to explain the following irregularities. 1) There is a variable depth distribution to seismic layer 3, ranging from 2-11 km throughout Iceland, a depth which should be constant if invoking stationary plate boundaries with continual symmetric spreading (Helgason, 1985) (Figure 1.21). 2) Iceland has an excess width, about 40% wider than would be expected by constant spreading rates from a single boundary, and asymmetric isochrons, with ages of 16 Ma in Western Iceland and 13 Ma in Eastern Iceland, (Figure 1.22) (Helgason, 1985; Schilling et. al., 1982). This fails to correlate with spreading of the Mid-Atlantic Ridge systems north and south of Iceland which show symmetric spreading patterns up to 20 Ma (Helgason, 1985) (Figure

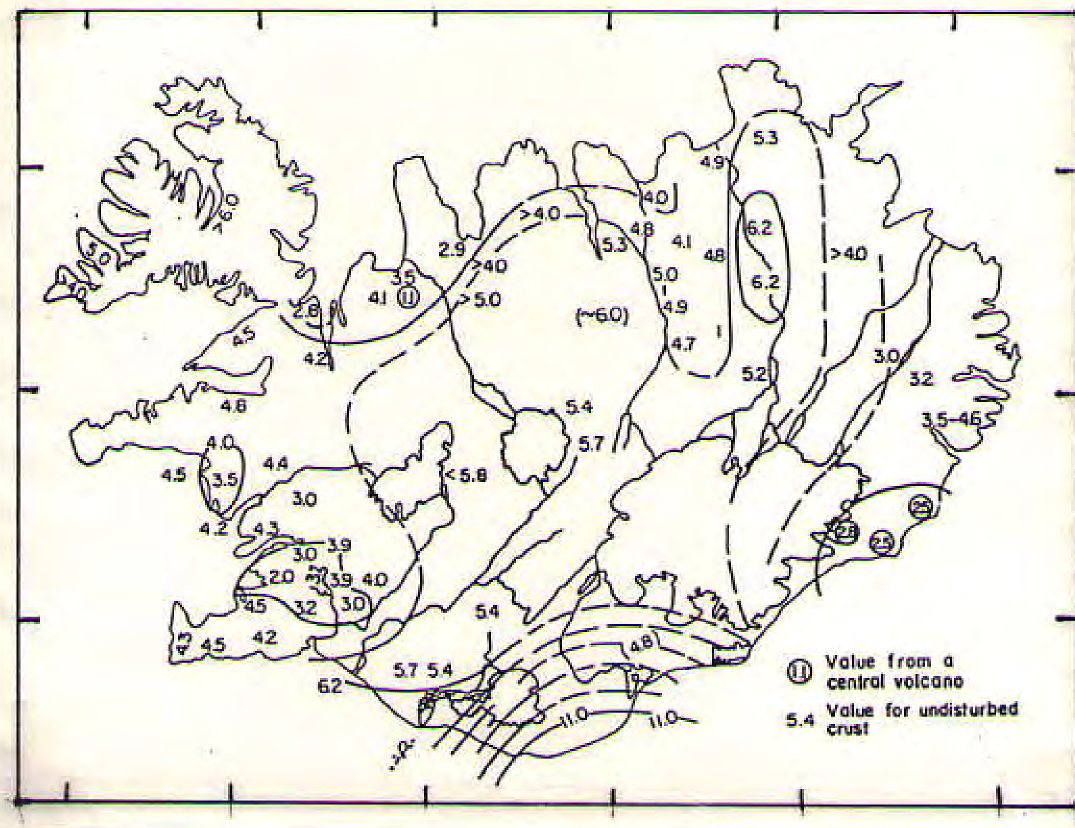

Figure 1.21: Depth to the lower crust (layer 3) in Iceland (in km). Velocity is assumed to increase continuously with depth (from Flovenz, 1980).

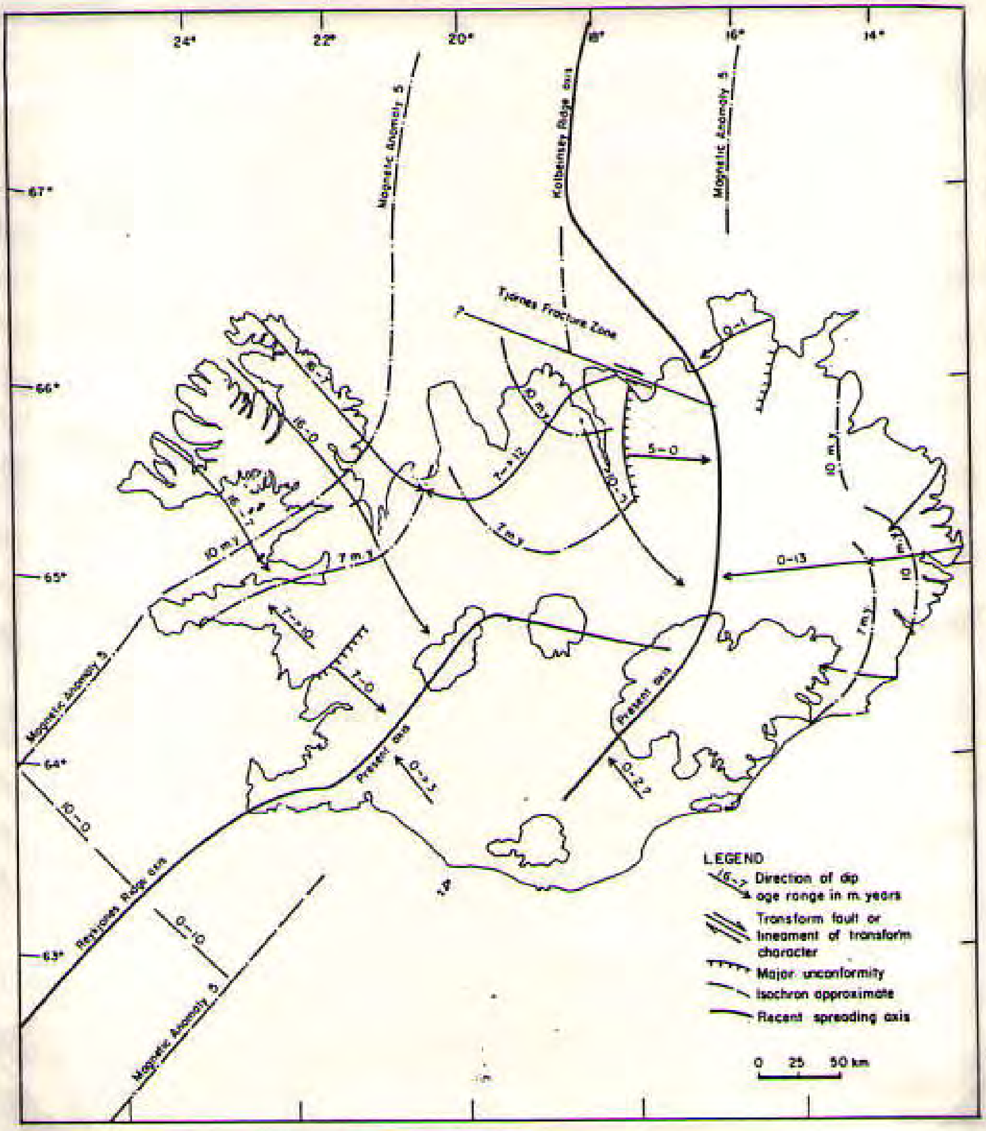

Figure 1.22: Development of the present geometry of the tectonically active zones in Iceland. Traces of extinct rift zones are left as structural synclines in the Tertiary areas. Major unconformities have developed between sequences produced within extinct rift zones and those produced from the presently active ones (from Sæmundsson, 1979).

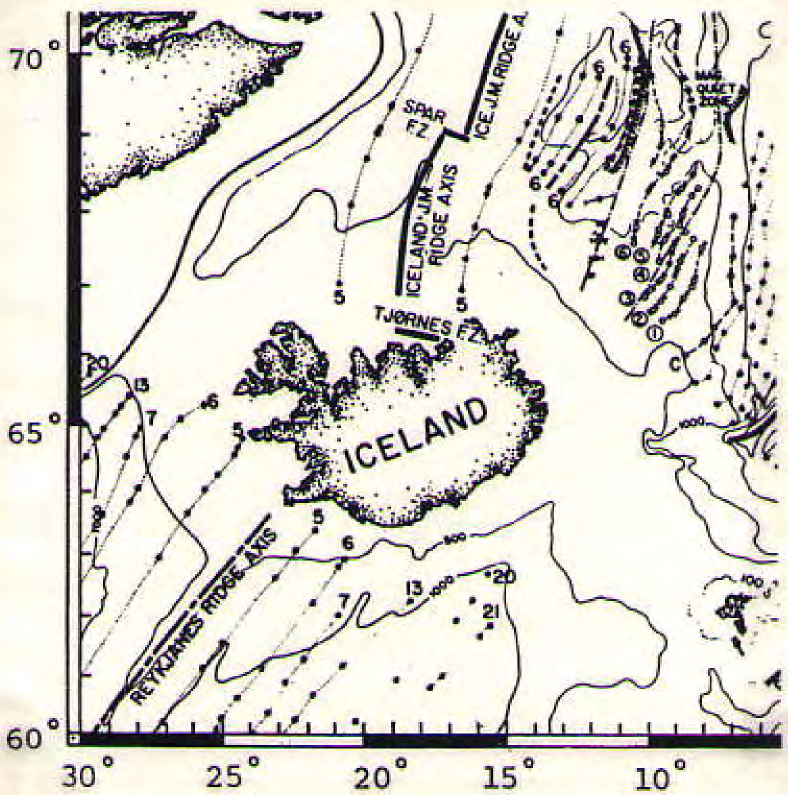

Figure 1.23: Some major identified magnetic anomaly lineations in the ocean surrounding Iceland (from Björnsson, 1981).