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Rifting above a mantle plume: structure and development of the Iceland Plateau

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Abstract

The interaction of the Mid-Atlantic Ridge with the North Atlantic Mantle Plume has produced a magmatic plateau centred about Iceland. The crust of this plateau is 30 km thick on average. This abnormal thickness implies that, unlike other slow-spreading ridges, addition of magmatic material to the crust is not balanced by crustal stretching. The thermal effect of the plume also reduces the strength of the lithosphere. Both mechanisms affect the rifting process in Iceland. A structural review, including new field observations, demonstrates that the structure of the Iceland plateau differs from that of other slow-spreading oceanic ridges. Lithospheric spreading is currently accommodated in a 200 km wide deformation strip, by the development of a system of half-grabens controlled by growth faults. Similar extinct structures, with various polarities, are preserved in the lava pile of the Iceland plateau. These structures are identified as lithospheric rollover anticlines that developed in hanging walls of listric faults. We introduce a new tectonic model of accretion, whereby the development of the magmatic plateau involved activation, growth and decay of a system of growth fault/rollover systems underlain by shallow magma chambers. Deactivation of a given extensional system, after a lifetime of a few My, was at the expense of the activation of a new, laterally offset, one. Correspondingly, such systems formed successively at different places within a 200 km wide diffuse plate boundary. Unlike previous ones, this new model explains the lack of an axial valley in Iceland, the dip pattern of the lava pile, the complex geographical distribution of ages of extinct volcanic systems and the outcrops of extinct magma chambers.

Résumé

L'interaction entre la Ride Médio-Atlantique et le Panache Nord-Atlantique a provoqué la formation d'un plateau magmatique, de 30 km d'épaisseur moyenne, centré sur l'Islande. L'exceptionnelle épaisseur crustale du plateau implique que, contrairement à ce qui se passe sur les autres dorsales lentes, l'addition de matériel magmatique à la croûte n'est pas équilibrée par l'étirement crustal. L'anomalie thermique liée au panache réduit aussi la résistance de la lithosphère. Ces deux mécanismes peuvent influencer le processus de rifting en Islande. À partir d'une synthèse structurale et de nouvelles observations de terrain, nous montrons que la structure de l'Islande diffère de celle des autres dorsales lentes. L'étirement de la lithosphère y est actuellement accommodé, dans une bande de déformation de 200 km de large, par la formation de demi-grabens contrôlés par des failles de croissance. D'anciennes structures similaires sont préservées dans la pile de lave du plateau islandais. Ces structures sont des anticlinaux en rollover d'échelle lithosphérique, qui se sont formés dans le compartiment affaissé de failles listriques. Sur la base de ces observations, nous proposons un nouveau modèle tectonique de rifting à l'aplomb d'un panache. Dans ce modèle, le plateau magmatique se développe par apparition, croissance et extinction de systèmes composés d'une faille de croissance, d'un anticlinal en rollover et d'une chambre magmatique située à l'interface entre croûte fragile et croûte ductile. Après une durée de fonctionnement de quelques Ma, l'extinction d'un système extensif donné est compensée par l'activation d'un nouveau système décalé latéralement. Ainsi, les systèmes extensifs apparaissent

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successivement en différents endroits d'une limite de plaque diffuse de 200 km de large. Contrairement aux modèles antérieurs, ce modèle rend compte de l'absence de vallée axiale en Islande, des pendages variés observés dans la pile de lave, de la distribution géographique complexe des âges des systèmes volcaniques éteints, et de la mise à l'affleurement d'anciennes chambres magmatiques.

Keywords: Ridge; Hotspot; Growth fault; Rollover; Magma chamber

Mots clé :

1. Introduction

Ridge-plume interactions have been widely studied with regard to asthenospheric flow, mantle melting, chemical and petrologic composition of rocks, crustal thickness and surface morphology [1-17]. In contrast, the effects of mantle plumes on the mode of lithospheric extension at oceanic ridges remain largely obscure [18, 19]. A tectonic model of oceanic rifting above mantle plumes is still lacking.

The Mid-Atlantic Ridge is a slow-spreading oceanic ridge. In Iceland, it is located above a mantle plume and rises above sea level. Iceland thus constitutes a unique opportunity to constrain tectonic models of oceanic rifting above mantle plumes. After a brief review of the tectonics of slow-spreading ridges, we introduce the geological framework of Iceland. Then, we critically review previous accretion models. From a structural synthesis, including new field observations, we produce cross-sections of Holocene and extinct volcanic systems. On the basis of these cross-sections, we propose a tentative model for rifting above mantle plumes.

2. Tectonics of slow-spreading ridges

Slow-spreading ridges are linear boundaries between two oceanic lithosphere plates that diverge at less than 3 cm/yr. Slow-spreading ridges are usually composed of an axial valley, 1 to 20 km wide and 1.5 to 3 km deep (Fig. 1). At the bottom of the valley is the Neovolcanic Zone, 1 to 5 km wide, where new mantle-derived material is constantly added to the crust by superficial volcanism and by subsurface magma injection [20]. The newly formed lithosphere is continuously stretched in response to plate divergence [21]. Stretching is accommodated by viscous flow in the lower ductile part of the lithosphere and by normal faulting in its upper brittle part. The faults generally dip towards the spreading axis and bound outwards-tilted blocks. The faults form near the spreading axis, in the Neovolcanic Zone. They drift away, as they are driven by plate separation, and eventually die when they get sufficiently far from the spreading axis. Then the deformation is transferred to faults newly forming in the Neovolcanic Zone. The active spreading axis

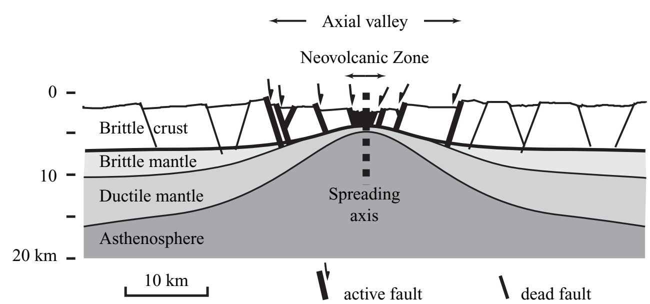

Fig. 1 Typical cross-sectional structure of slow-spreading oceanic ridge (based on [20-22]). In the upper part of the crust, stretching is accommodated by normal faults. The faults dip towards the spreading axis and bound outwards-tilted blocks. The Neovolcanic Zone is narrow, symetric, and is located in an axial valley pinned at the plate boundary.

Figure 1. Structure en coupe d'une dorsale lente classique (schéma basé sur [20-22]). Dans la partie supérieure de la croûte, l'étirement est accommodé par des failles normales. Les failles pendent en direction de l'axe de divergence ; elles limitent des blocs qui sont basculés vers l'extérieur. La Zone Néovolcanique est étroite et symétrique ; elle est située dans une vallée axiale fixée sur la limite de plaques.

thus remains pinned to a narrow strip located at the plate boundary. The balance between addition of material to the crust and stretching of the crust usually yields a symmetrical, steady state, spreading process [21-23] that generally gives rise to 6-7 km thick oceanic crust [24], with lava isochrons distributed symmetrically across the spreading axis.

3. Geological setting

3.1. Geodynamic framework

The North Atlantic Ocean has opened in response to the divergence of the European and the North American plates (Fig. 2a). The boundary between these plates is formed by the Mid-Atlantic Ridge. At the latitude of Iceland (65°N), these plates diverge at a half-spreading rate of 0.9 cm/yr and along a path striking N105° [25, 26]. Here the spreading ridge interacts with the North Atlantic Mantle Plume. This plume has been imaged seismically down to a depth of about 400 km [27, 28], throughout the transition zone [29, 30] and more tentatively down to the core-mantle boundary [31, 32].

This plume was activated during the late Senonian (ca. 80 My) and remained active during the opening of the North Atlantic Ocean that had commenced at 54 My [33, 110]. Since then, ridge-plume interaction was responsible for vigorous tectonics and volcanism, which produced the 30 km thick magmatic Iceland plateau between Greenland, Scotland and Norway. This plateau, which is associated with a chemical and topographic anomaly [1, 2, 4, 11, 12, 16, 34], rises above sea level in Iceland, in the prolongation of the Reykjanes and Kolbeinsey Ridges.

3.2. Morphology

A well-defined axial valley marks the axis of the Mid-Atlantic Ridge south of 59° N (Fig. 2b). This axial valley disappears on the Reykjanes Ridge (Fig. 2c), and reappears in the northern part of the Kolbeinsey Ridge (Fig. 2e). Between 59° N and 69° N, the axial valley is replaced by a wide crest, resembling the axial highs of fast-spreading ridges [4, 18]. The lack of an axial valley on the Reykjanes Ridge has been ascribed to the mantle plume [18]. Across Iceland, the axial high evolves into a plateau, more than 1,000km in diameter, and standing 2,000 to 2,500 metres above the surrounding oceanic floor (Fig. 2d). There is no axial valley on this plateau and Holocene volcanic systems do not have a clear topographic expression.

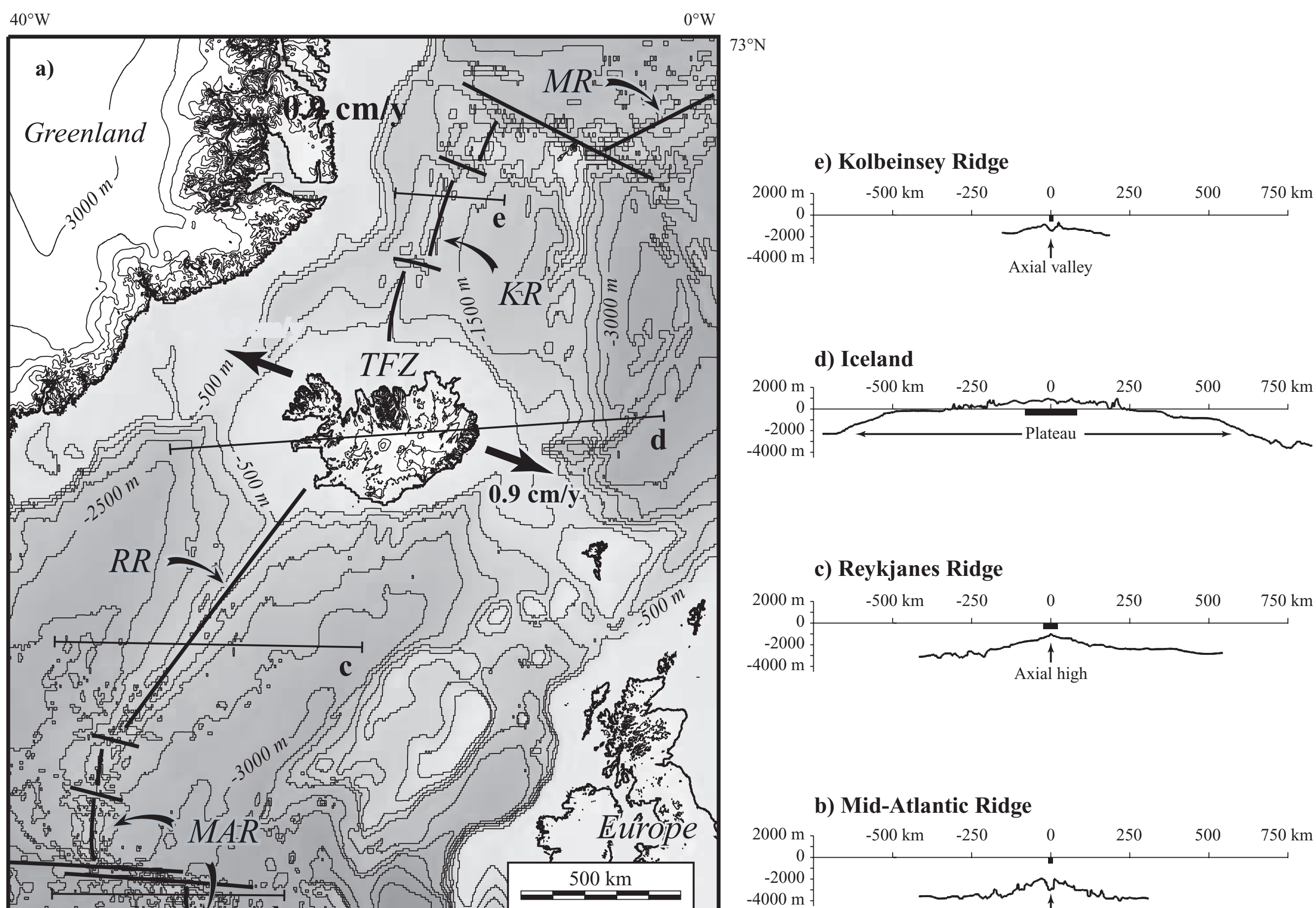

Fig. 2 **a)** Geodynamic framework and topography of North Atlantic Ocean. Spreading axis and transform zones of Mid-Atlantic Ridge indicated by bold lines. Topographic contours at 500 m interval drawn from USGS GLOBE digital elevation models (resolution: 30” onshore and 5’ offshore). World Mercator Projection. MAR: Mid-Atlantic Ridge; RR: Reykjanes Ridge; TFZ: Tjörnes Fracture Zone: KR: Kolbeinsey Ridge; MR: Mohns Ridge. **b-e)** Topographic profiles across Mid-Atlantic Ridge at various latitudes, drawn from same GLOBE data. Width of Holocene active zone indicated by black box on each profile.

Figure 2. a) Cadre géodynamique et topographie de l’Océan Atlantique Nord. L’axe d’accrétion et les zones transformantes de la dorsale Médio-Atlantique sont indiqués par les lignes en gras. Les courbes de niveau (intervalle 500 m) ont été tracées à partir des modèles numériques de terrain GLOBE de l’USGS (résolution : 30’’ à terre, 5’ en mer). Projection Mercator. MAR : Dorsale Médio-Atlantique ; RR : Ride de Reykjanes ; TFZ : Zône de Fracture de Tjörnes : KR : Ride de Kolbeinsey ; MR : Ride de Mohns. **b-e)** Profils topographiques perpendiculaires à la dorsale Médio-Atlantique à différentes latitudes, établis à partir des mêmes données GLOBE. La largeur de la zone active holocène est indiquée sur chaque profil par un rectangle noir.

3.3. Width and location of active zone

Holocene volcanic systems of Iceland consist of fissure swarms connected to central volcanoes (Fig. 3a) [35, 36]. The fissure swarms comprise normal faults, tension fractures and eruptive fissures. The central volcanoes are sites of focused lava emission, which are underlain at depths of 3 to 6 km by magma chambers [36, 114].

At slow-spreading ridges, the active zone is generally a narrow strip pinned at the plate boundary and marked by Holocene volcanic systems (Figs. 1, 2a and 2e). Ages of extinct volcanic systems are distributed symmetrically with respect to the active zone. However in Iceland, Holocene fissure swarms and central volcanoes are scattered in a 200 km wide strip that extends across the island from the south-west to the north (Fig. 3a). The ages of extinct fissure swarms are not distributed symmetrically with respect to Holocene central volcanoes [37, 38].

Seismicity also is scattered throughout this uncommonly wide deformation strip (Figs. 3b and 4a). Earthquakes are particularly numerous in the South Iceland Seismic Zone (SISZ) and in the Tjörnes Fracture Zone (TFZ), which act as transfer zones between the wide deformation strip of Iceland and the narrow active zones of the Reykjanes and Kolbeinsey Ridges, respectively.

In the northern half of the island, the deformation strip comprises seven Holocene fissure swarms. These are associated with the following central volcanoes, from East to West: Kverkfjöll, Askja, Fremri-Namur, Krafla, Theistareykir, Hofsjökull and North-Langjökull (Fig. 3a) [40]. During the Holocene, the latter two have been eruptive in their southern part only; non-eruptive Holocene faults have been recognized in their northern part [40]. An additional swarm of non-eruptive faults has been recognized recently in Eyjafjördur, a fjord north of Akureyri [108]. In the southern half of the island, Holocene eruptive fissure swarms are located along the eastern and western margins of the deformation strip, whereas there are non-eruptive fault swarms in the central part (Fig. 3a) [40]. The latter are connected northwards to the Kerlingarfjöll central volcano, which was active during the Holocene.

The location of subglacial volcanic edifices emplaced during the Bruhnes magnetic epoch indicates that volcanic activity shifted throughout the deformation strip during the last 0.8 My (Fig. 3c) [43-47]. The ages of dikes, reflecting the ages of extinct fissure swarms, also demonstrate that volcanic activity shifted throughout the deformation strip during the last few My (Fig. 3c) [38]. For older lava flows currently located outside the deformation strip, Walker [92] reached the same conclusion by demonstrating that young dike swarms traversing old terrains strike at 30-45° angle with the lava pile. Hence the classical concept of spreading axes (narrow active zone pinned at the plate boundary and age symmetry of extinct volcanic systems) cannot be applied to Iceland. On a time scale of a few My, volcanic and tectonic activity shifted across a 200 km wide diffuse plate boundary, which suggests that the mode of deformation of Iceland differs from that of other slow-spreading ridges.

3.4. Rheological layering

The mode of deformation of rift zones is controlled by the rheological profile of the lithosphere, especially by the depth of the brittle/ductile transition and by the brittle/ductile strength ratio [48-50, 111]. Rheological profiles can be computed from classical brittle/ductile deformation laws, provided that the thickness of the crust, its composition, the temperature gradient and the strain rate are known.

Seismic experiments and gravity modelling indicate that the crust of Iceland is 19 to 42 km thick [41, 51, 52]. This anomalous thickness is due to vigorous magma supply from the mantle. Contrasting with classical slow-spreading ridges that are not influenced by plumes, magma supply in Iceland is not balanced by crustal stretching, as evidenced by the Iceland plateau and the dome-shaped topographic profile of the Reykjanes Ridge (Fig. 2).

The crust comprises lava flows, dikes, sills and intrusive magma bodies. Ancient shield volcanoes, aeolian volcaniclastic deposits, sedimentary layers and plant-bearing lignite horizons are intercalated in the lava pile [35, 53].

The surface temperature gradient varies from 100°C/km in the deformation strip to 50°C/km outside the strip. Within the strip, values of 150°C/km have been measured near Holocene volcanic systems [54]. The temperature of the lower crust is still controversial: downward extrapolation of surface temperature gradients, magnetotelluric measurements and shear wave attenuation suggest a temperature of 1,200°C at the Moho, whereas temperatures estimated from recent seismic studies are in the range of 600 to 950°C (see review of controversy in [41]). The occurrence of magma chambers suggests that temperatures of 1,000-1,200°C are reached locally at depths of 4 to 6 km within the deformation strip. Hence a reasonable value for the mean crustal temperature gradient in the deformation strip is 100°C/km.

If we assume that lithospheric stretching is distributed uniformly throughout the deformation strip, the computed strain rate is 10^{-15} s^{-1}. A simple rheological profile for the deformation strip is shown in Fig. 4c. It was computed with the above values, using Byerlee's law [55] for the brittle behaviour and Weertmann's law [56] for the ductile behaviour. The brittle/ductile transition is located at a depth of 5 km, which is consistent with a sharp decrease in the number of earthquakes below this level (Fig. 4b). Below 12 km, the strength of the ductile crust does not exceed 1 MPa (Fig. 4c). Thus in the deformation strip, the mechanically strong part of the lithosphere comprises merely the uppermost 12 km of the crust; the remainder of the crust (7 to 30 km) has a very low viscosity and is probably underlain directly by the asthenospheric mantle (Fig. 4d).

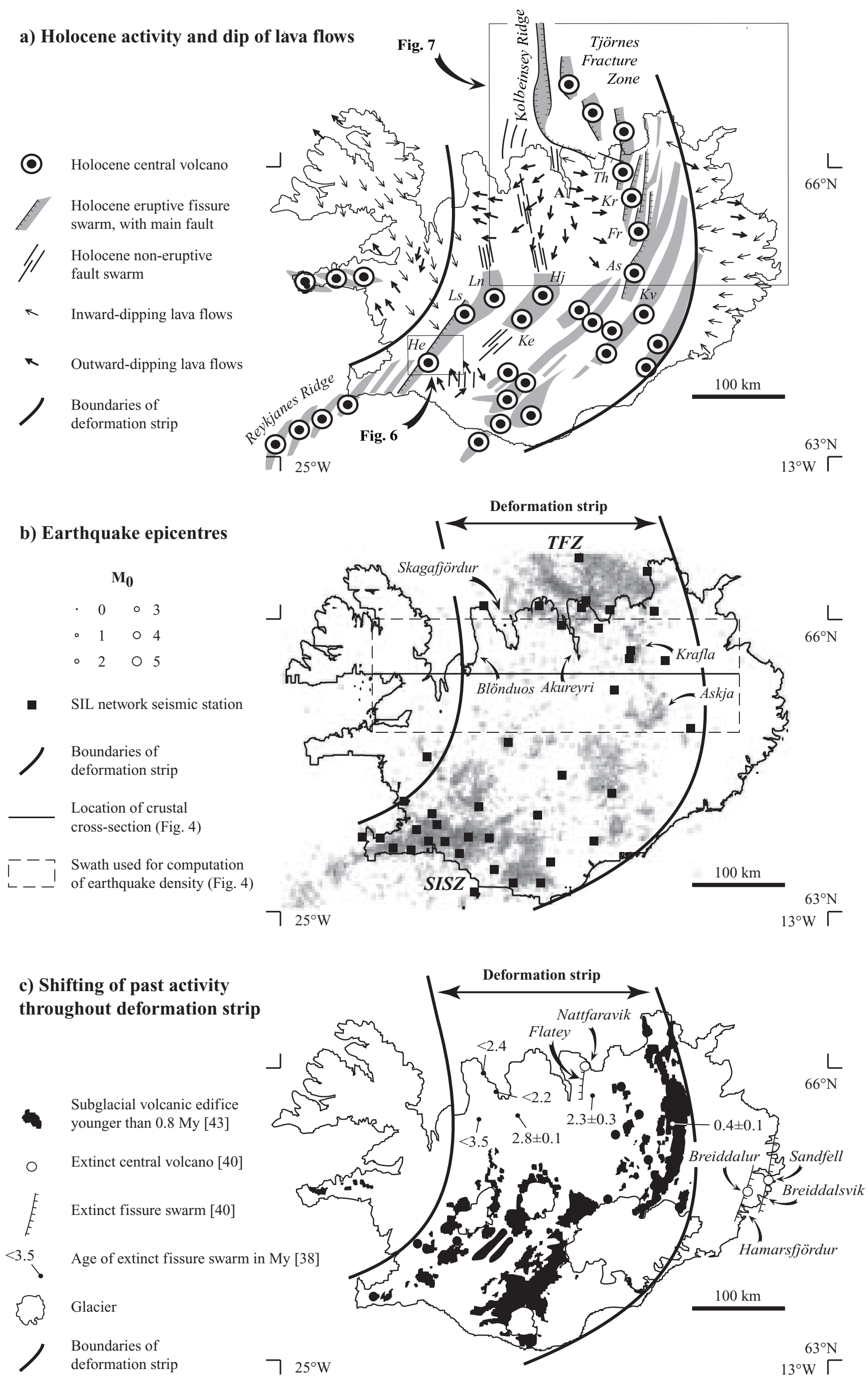

Fig. 3 Long-term deformation strip of Iceland. **a)** Location of Holocene activity (central volcanoes, eruptive fissure swarms and non-eruptive fault swarms) and dip of lava flows. A: Akureyri; As: Askja; Fr: Fremri-Namur; He: Hengill; Hj: Hofsjökull; Ke: Kerlingarfjöll; Kv: Kverkfjöll; Ln: North-Langjökull; Ls: South-Langjökull; Th: Theistareykir. Compilation based on several existing maps [35, 39, 40, 81, 108]. **b)** Epicentres of all earthquakes recorded by SIL seismic network between 01/01/1995 and 15/11/2003 (Icelandic Meteorological Office, Department of Geophysics, http://hraun.vedur.is). SISZ: South Iceland Seismic Zone; TFZ: Tjörnes Fracture Zone. **c)** Ages of extinct fissure swarms and location of subglacial volcanic edifices younger than 0.8 My, indicating that activity shifted throughout deformation strip during last few My. Extinct fissure swarms and central volcanoes discussed in text also indicated.

Figure 3. Bande de déformation à long terme de l'Islande. **a)** Localisation de l'activité holocène (volcans centraux, faisceaux de fissures éruptives et faisceaux de failles non éruptives) et pendage des coulées de lave. A : Akureyri ; As : Askja ; Fr : Fremri-Namur ; He : Hengill ; Hj : Hofsjökull ; Ke : Kerlingarfjöll ; Kv : Kverkfjöll ; Ln : Nord-Langjökull ; Ls : Sud-Langjökull ; Th : Theistareykir. Compilation basée sur diverses cartes existantes [35, 39, 40, 81, 108]. **b)** Épicentres de tous les séismes enregistrés par le réseau sismique SIL entre le 01/01/1995 et le 15/11/2003 (Icelandic Meteorological Office, Department of Geophysics, http://hraun.vedur.is). SISZ : Zone Sismique Sud-Islandaise ; TFZ : Zone de Fracture de Tjörnes. **c)** L'âges des faisceaux de fissures éteints et la localisation des édifices volcaniques sous-glaciaires plus jeunes que 0,8 Ma indiquent que l'activité s'est déplacée à l'intérieur de la bande de déformation durant les derniers Ma. Les faisceaux de fissures et les volcans centraux éteints présentés dans le texte sont aussi indiqués.

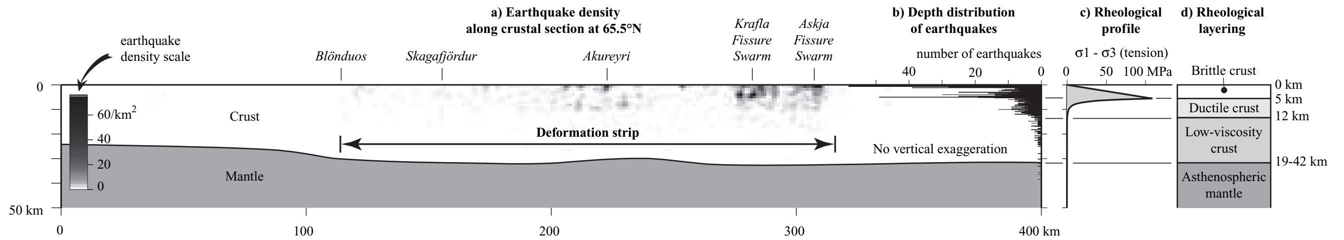

Fig. 4 Rheological layering of deformation strip. **a)** Earthquake density along crustal section at 65.5°N. Density computed as number of earthquakes located in 3 km vertical search radius, successively centred on each cell of 1 × 1 km Cartesian grid placed on cross-section. Includes all earthquakes comprised in 1° wide longitudinal swath centred at 65.5°N. Location of cross-section and of swath indicated on Fig. 3b. Source of data: SIL database (Sep/25/1991 – Nov/6/2003; $0 \leq M_l \leq 4.5$; Icelandic Meteorological Office, Department of Geophysics, http://hraun.vedur.is). Thickness of crust after Foulger *et al.* [41]. **b)** Histogram (500 m classes) showing depth distribution of earthquakes along same swath. **c)** Rheological profile computed from Byerlee's law [55] for brittle behaviour and Weertmann's law [56] for ductile behaviour, assuming linear temperature gradient of 100°C/km, strain rate of $10^{-15}s^{-1}$ and rheological parameters of diabase for whole crust ($A = 2.10^{-4}$ MPa^{-n} s^{-1}, n = 3.4, E = 260 kJ mol^{-1}, ρ = 3,000 kg m^{-3}; [57]). **d)** Inferred rheological layering.

Figure 4. Stratification rhéologique de la bande de déformation. **a)** Densité de séismes le long d'une coupe crustale à 65,5°N. La densité est calculée comme le nombre de séismes situés à l'intérieur d'un cercle vertical de 3 km de rayon, centré successivement sur chaque nœud d'une grille cartésienne de 1 × 1 km placée sur la coupe. Tous les séismes situés à l'intérieur d'une bande de calcul longitudinale, de 1° de large et centrée à 65,5°N, ont été pris en compte. La coupe et la bande de calcul sont localisées sur la Fig. 3b. Source des données : base de donnée SIL (Sep/25/1991 – Nov/6/2003 ; $0 \leq M_l \leq 4{,}5$; Icelandic Meteorological Office, Department of Geophysics, http://hraun.vedur.is). Épaisseur de la croûte d'après Foulger *et al.* [41]. **b)** Histogramme (classes de 500 m) montrant la distribution des séismes en fonction de la profondeur, dans la même bande de calcul. **c)** Profil rhéologique calculé à partir de la loi de Byerlee [55] pour le comportement fragile et de la loi de Weertmann [56] pour le comportement ductile, en supposant un gradient thermique linéaire de 100 °C/km, un taux de déformation de $10^{-15}s^{-1}$ et en utilisant les paramètres rhéologique de la diabase pour l'ensemble de la croûte ($A = 2{,}10^{-4}$ MPa^{-n} s^{-1}, n = 3,4, E = 260 kJ mol^{-1}, ρ = 3 000 kg m^{-3} ; [57]). **d)** Stratification rhéologique déduite.

4. Previous accretion models

4.1. Models with volcano-tectonic activity pinned at plate boundary

Iceland was first recognized as a region of crustal accretion by Bödvarsson and Walker in 1964 [58]. These authors suggested that the whole basaltic pile developed by injection of dikes and emplacement of lava flows in the currently active zone, followed by symmetric drift and subsidence (Fig. 5a).

Palmason [59, 60] developed thermo-kinematic numerical models based on the ideas of Bödvarsson and Walker [58]. Two plates cool down as they symmetrically drift apart at a specified rate (Fig. 5b). The processes of subsidence and of accretion (intrusion of dikes and sills, emplacement of lava flows, opening of fractures and development of normal faults) were not further specified and were integrated in an overall density function centred at the plate boundary. Isotherms, particle trajectories and lava isochrones predicted by the model are illustrated in Fig. 5b. This model, which was recently slightly modified to take into account mass and heat transfer between the crust and the mantle [61], was supported by Hardarson and Fitton [62] on the basis of petrological data.

Daignières *et al.* [63] proposed a kinematic numerical model (Fig. 5c), based on a sequence of four accretion processes (opening of fractures, development of normal faults, injection of dikes and emplacement of lava flows). Each accretion process is defined *a priori* by the following geometric parameters: (1) location of structures, (2) width of dikes and fractures, (3) thickness and lateral extension of lava flows, and (4) dip and throw of faults. This model arbitrarily creates antithetic pairs of coeval faults with equal throws. Accretion processes proceed successively, according to a probability law. Particle trajectories and lava isochrones predicted by this model are similar to those predicted by the model of Palmason [59, 60]. In both models, crustal accretion is assumed to proceed in a narrow strip pinned at the plate boundary and the width of the strip is kept constant through time.

4.2. Models with shifting volcano-tectonic activity

Gibson and Piper [64] demonstrated that the lava pile of Iceland is composed of juxtaposed flow packages, partially overlapping each other and produced by distinct fissure swarms. Based on this observation, Gibson and Piper [64] developed an alternative model (Fig. 5d) according to which magma injection in a given fissure swarm produces a lenticular package of lava flows. After a given time lap, the fissure swarm and the lava package drift laterally, as a second fissure swarm is activated, controlling the development of a new lava lens adjacent to and partially overlapping the earlier formed package. Successive fissure swarms develop alternately on the western and eastern sides of a long-term deformation strip. Gibson and Gibbs [65] later refined this model by assuming that the development of a given fissure swarm is controlled by a listric fault (Fig. 5e). Based on a study of interstratified distal and proximal lava flows, Helgason [45, 46] supported the idea that the active zone shifts frequently (Fig. 5f). In these alternative models, volcano-

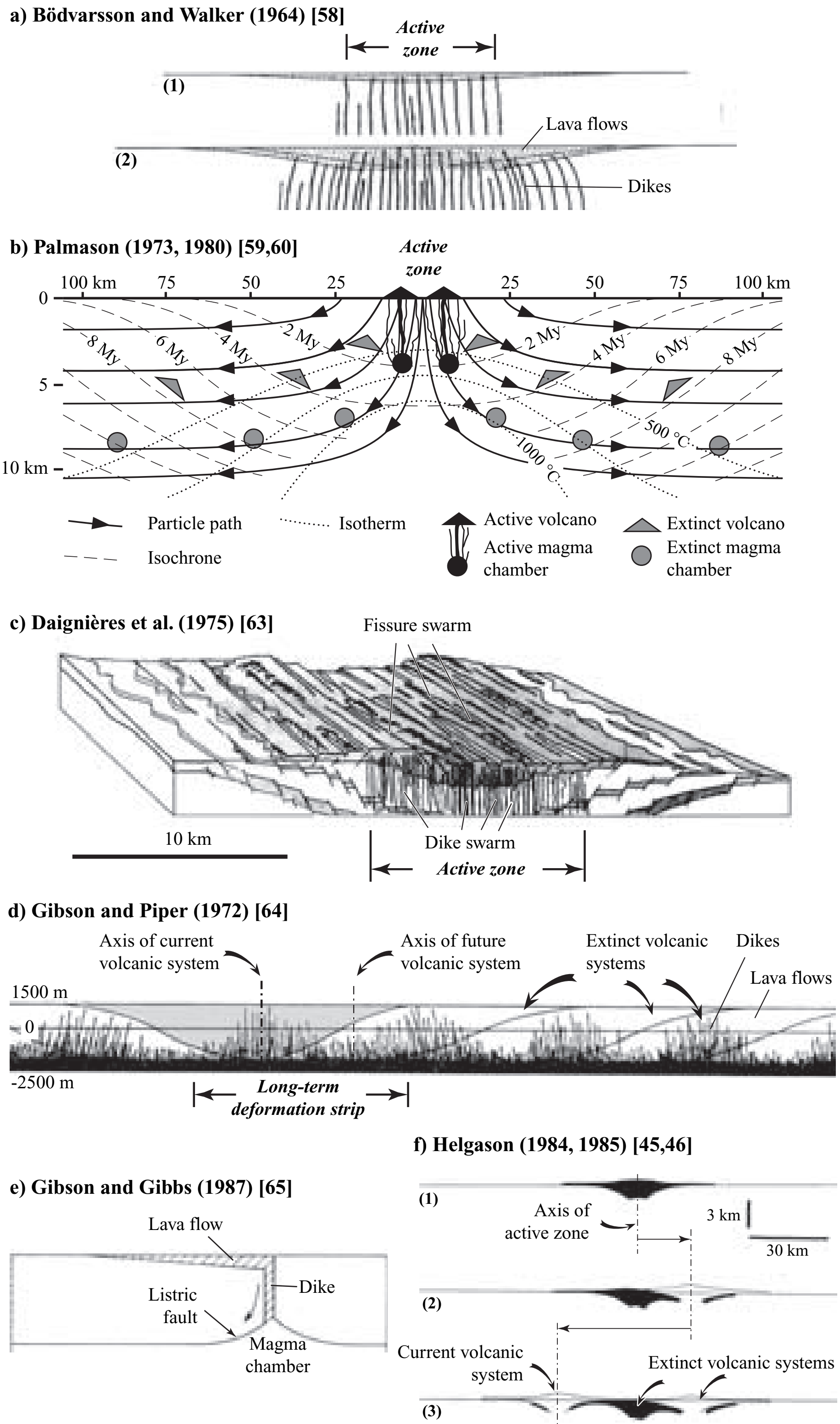

Fig. 5 Previous accretion models (redrawn with modifications from original figures). **a)** Two successive steps in development of lava pile according to Bödvarsson and Walker [58]. **b)** Palmason [59, 60]. **c)** Daignières *et al.* [63]. **d)** Gibson and Piper [64]. **e)** Gibson and Gibbs [65]. **f)** Three successive steps in development of lava pile according to Helgason [45, 46]. Volcano-tectonic activity remains pinned at narrow plate boundary in models a, b and c. Volcano-tectonic activity shifts laterally in models d, e and f.

Figure 5. Modèles d'accrétion antérieurs (redessinés avec des modifications d'après les figures originales). **a)** Deux étapes successives du développement de la pile de lave d'après Bödvarsson et Walker [58]. **b)** Palmason [59,60]. **c)** Daignières *et al.* [63]. **d)** Gibson et Piper [64]. **e)** Gibson et Gibbs [65]. **f)** Trois étapes successives du développement de la pile de lave d'après Helgason [45, 46]. L'activité volcano-tectonique reste fixée dans une limite de plaques étroite dans les modèles a, b et c. Elle se déplace latéralement dans les modèles d, e et f.

tectonic activity is not pinned at the plate boundary, but shifts throughout the island.

4.3. Major weaknesses of previous models

4.3.1. Dip of lava flows

At slow-spreading ridges, lava flows generally dip outwards (Fig. 1). This pattern has been successfully explained by classical ridge models, in which outwards-tilted blocks are controlled by inwards-dipping faults [21-23]. In contrast, previous models for Iceland predict that lava flows will symmetrically dip towards the spreading axis if activity is pinned at the plate boundary (Figs. 5a, b, c).

In Iceland though, lava flows do not systematically dip towards the present spreading axis (Fig. 3a). They do not systematically dip outwards either. Thus the dip of the lava flows in Iceland can be explained neither by classical tectonic models for slow-spreading ridges, nor by models, previously designed for Iceland, with volcano-tectonic activity pinned at the plate boundary.

4.3.2. Depth of burial of lava flows and acid magmatism

Models with volcano-tectonic activity pinned at the plate boundary predict that lava flows and volcanoes emplaced at a given time, would get rapidly buried beneath subsequent lava flows (Fig. 5b). However, as already pointed out by Helgason [45, 46], this is inconsistent with outcrops of ancient volcanoes, magma chambers and dikes in the western and eastern parts of the island (Fig. 3c). If the lava flows and the interlayered sedimentary and plant-bearing lignite horizons, which actually crop out in western and eastern Iceland [35, 53] had been buried so deeply and were subsequently exhumed by erosion, they would have experienced temperatures of 500 to 1,000°C and pressures of 100 to 150 MPa (Fig. 5b). Metamorphic minerals, indicative of such pressure and temperature conditions, would have grown in pores. However the pores of some of the oldest lava flows in western and in eastern Iceland are actually filled by zeolites (chabazite, analcite, mesolite and laumontite), which grow at temperatures of between 180° and 300° [66]. As in hydrothermal systems such as volcanoes, these temperatures are reached at very shallow levels [66], this speaks against deep burial of some of the old lava flows. By contrast, models with shifting volcano-tectonic activity do not predict deep burial of lava flows and of volcanoes, as successive lava packages do not pile up along a pinned axis.

The crust of Iceland comprises a small but significant amount of acid magmatic products (10% according to Jakobsson [67]). Acidic rocks include rhyolitic to dacitic lava flows, emplaced near central volcanoes, and dioritic to granodioritic bodies intruding the lava pile. Acidic rocks are rather uncommon in oceanic settings. The production of acidic magmas in Iceland has been attributed to partial remelting of subsided lava flows, in the vicinity of shallow magma chambers [68-71].

Hence tectonic models for Iceland must take into account the fact that (1) some solidified lava flows have melted again after they subsided, as suggested by the production of acidic magmas, while (2) others remained close to the surface, as indicated by the lack of metamorphic minerals in their pores.

4.3.3. Location of activity at a given time and distribution of ages of extinct volcanic systems

In models with volcano-tectonic activity pinned at the plate boundary, a symmetrical distribution of ages of extinct volcanic systems across the Holocene active zone is expected. In models with shifting volcano-tectonic activity, it is assumed that one fissure swarm only is active at a given time (Figs. 5d-f). These assumptions are inconsistent with the actual pattern of Holocene activity along multiple fissure swarms (Fig. 3a) and with the asymmetric distribution of ages of extinct fissure swarms [37, 38].

4.3.4. Faults

As normal faults are apparently less numerous than fissures and the throws of faults do not exceed 100 metres at the surface of Holocene fissure swarms [72, 73], faults have been ignored or arbitrarily defined in previous models. The reason for the small apparent throws however, is that the faults are growth faults that become smoothed out time and again by lava flows [74]. Hence they cannot be detected from surface observations. Faults with throws larger than 100 m are exposed in natural cross-sections of the lava pile however [36].

Now it has been plainly demonstrated that deformation at oceanic ridges is dominated by faulting [20-23]. Hence a tectonic model for Iceland must embrace faulting. As in such a model, faults must be consistent with structures observed in the field, we describe in the following the structure of characteristic Holocene fissure swarms in order to constrain a faulted model.

5. Structure of Holocene fissure swarms

5.1. Southern Iceland

In southern Iceland, the deformation strip comprises several Holocene fissure swarms striking NNE (Fig. 3a). Glaciers and thick subglacial volcanic products cover the easternmost swarms, which make it difficult to observe the underlying structure (Fig. 3c). In part of the westernmost (Thingvellir) fissure swarm, however, subglacial and postglacial volcanic productions failed to keep pace with faulting and subsidence; hence the deep structure of this swarm may be inferred from surface observations. The swarm joins the Hengill and South-Langjökull central volcanoes (Fig. 3a). It is covered by lava flows younger than 0.8 My. The Thingvellir fissure swarm is asymmetrical and bordered to the West by a sharp tectonic contact between lava flows younger than 0.8 My and older basalts (Fig. 6). This contact is formed by narrowly spaced faults with throws of several tens of metres. These are growth faults, with vertical offsets increasing from about 100 m in ~0.1 My old formations, to 400 m in ~0.3 My old formations [74]. In contrast, the eastern margin of the Thingvellir fissure swarm corresponds to the stratigraphic contact between lava flows emplaced during the Bruhnes magnetic epoch, and older basalts dipping westwards. A series of step faults, with vertical offsets of a few metres only, occur in a 20 km wide belt adjacent to the western border of the swarm [74-78]. In conclusion, the Thingvellir fissure swarm is a westwards dipping half-graben (Fig. 6b); it is controlled by a major east dipping growth fault that extends to the Hengill and Langjökull central volcanoes.

5.2. Northern Iceland

In northern Iceland, the deformation strip comprises two non-eruptive fault swarms in the west and five eruptive fissure swarms in the east (Fig. 3a). A non-eruptive fault swarm has been recently recognized in the central part of the deformation strip in Eyjafjördur, a fjord north of Akureyri [108]. The five eruptive swarms strike NNE and they form an *en-échelon* array (Fig. 7). Central volcanoes associated with these eruptive swarms fall into a curved line striking from North-South to NW-SE. This suggests that magma is preferentially supplied from a strip that trends oblique to the eruptive fissure swarms and is located along their western boundary. Two central volcanoes at least have developed calderas underlain by magma chambers: Askja and Krafla [36, 40].

Most parts of the fissure swarms have been covered by subglacial volcanic products and Holocene lava flows, thus rendering it difficult to observe the underlying structure. However the deep structure of the fissure swarms can be inferred from observations in the vicinity of the Tjörnes peninsula. The eastern flank of the Tjörnes peninsula is a sharp tectonic contact between the volcanic cover younger than 0.8 My to the east, and older basalts to the west (Figs. 7 and 8). Fault scarps have been preserved along this contact [79]. The faults dip to the east, and their cumulative vertical offset is more than 1,000 m [80]. This fault zone can be followed offshore on seismic profiles (Fig. 9) [80-82] and can be traced southwards to the Theistareykir central volcano (Fig. 7).

In contrast, along the eastern margin of the deformation strip, the volcanic cover younger than 0.8 My rests conformably on older basalts that dip westwards (Figs. 7 and 8) [40]. A positive magnetic anomaly is associated with outcrops of basalts younger than 0.8 My (Fig. 8). The western flank of this anomaly is sharp and coincides with the location of the fault zone described above. This anomaly decreases gently eastwards, except for local departures above subglacial volcanic ridges [83, 84]. Piper [85] interpreted the shape of the anomaly as due to eastwards-thinning of the positively-magnetized volcanic cover younger than 0.8 My. Eastwards-thinning of this volcanic cover on-shore is consistent with eastwards-thinning of the sedimentary cover offshore (Fig. 9). In conclusion, the Holocene fissure swarms of north-eastern Iceland are controlled by a major east-dipping fault zone that extends along the eastern flank of the Tjörnes penisula (Fig. 8). The hanging wall dips westwards; it is filled by Bruhnes lava flows onshore and by sediments offshore.

5.3. Structural changes towards Kolbeinsey Ridge

The deformation strip of Iceland is connected to the Kolbeinsey Ridge by a complex transfer zone comprising the Husavik-Flatey right-lateral fault and the Tjörnes Fracture Zone (Fig. 7) [19, 86]. In the latter area, WNW striking seismic lineaments were previously interpreted as right-lateral strike-slip faults [87]. However recent accurate seismotectonic analysis demonstrated that earthquake clusters define left-lateral planes, striking north, and not WNW [88]. Bathymetric, magnetic, and seismic surveys had previously revealed that the Tjörnes Fracture Zone consists of three *en echelon* northerly striking troughs [4, 81]. These troughs are controlled by east-dipping faults marking their western borders (Figs. 7 and 9) [19, 44, 80]. Submarine volcanoes and geothermal fields are associated with these faults [4, 80-82]. Sediments fill the hanging walls of these faults, whereas the basaltic basement crops out in the footwalls (Fig. 9). Sediments in the hanging walls thin eastwards, thus defining in cross-section a wedge-shaped geometry that resembles the geometry inferred for the volcanic cover younger than 0.8 My onshore (Fig. 8).

Moving northwards to the Kolbeinsey Ridge, i.e. away from the Iceland plume, the width of the deformation strip decreases from 200 km to 20 km. There is a gradual transition from a series of synthetic eastwards-dipping growth-faults, controlling westwards-dipping half-grabens (Fig. 9a), to a pair of symmetrical outwards-dipping listric growth-faults (Figs. 9b-c), up to a faulted, 50 km wide crest resembling the axial highs of fast-spreading ridges (Fig. 9d). Further north, the classical structure of a slow-spreading ridge is recovered (Fig. 2e).

5.4. Synthesis: structure of Holocene fissure swarms

From surface observations, we inferred the deep structure of some Holocene fissure swarms of Iceland. These swarms form an array of *en echelon* half-grabens that are controlled by east-dipping faults. In the hanging walls, the Bruhnes volcanic cover (onshore) and the sedimentary cover (offshore) dip westwards and thin eastwards. Central volcanoes are located in the vicinity of the major faults, suggesting that magma ascent to the surface is closely linked to faulting. If continuous at depth, the magma plumbing system strikes oblique to the fissure swarms, however.

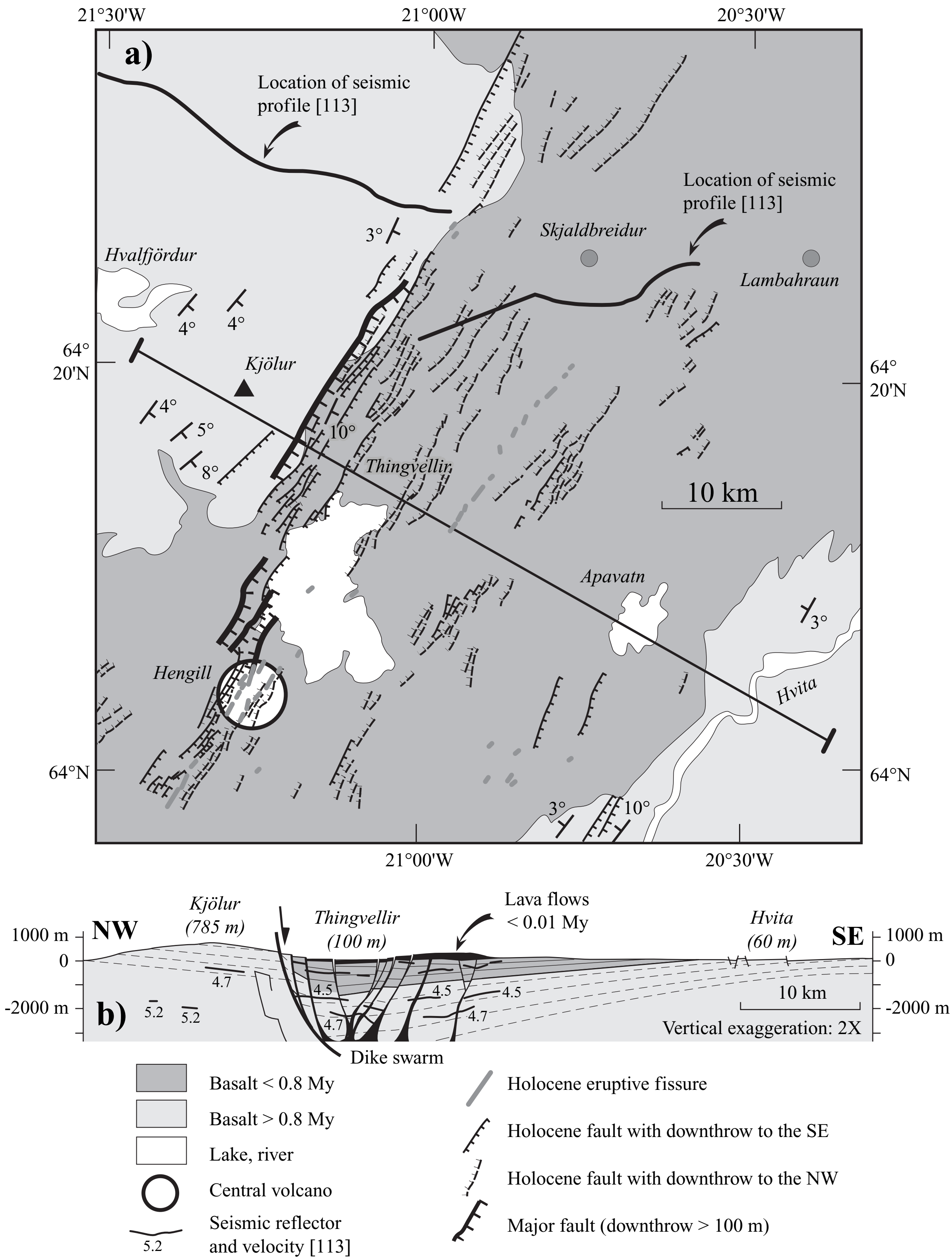

Fig. 6 **a)** Structural map [76, 77] and **b)** interpretative cross-section of Thingvellir fissure swarm (SW-Iceland). The fissure swarm is a half-graben tilted westwards. Seismic reflectors and velocities projected along strike from a seismic profile [113] located 20 km north of interpretative section. Location of map indicated by box in Fig. 3a.

Figure 6. a) Carte structurale [76, 77] et **b)** coupe interprétative du faisceau de fissures de Thingvellir (SO de l'Islande). Ce faisceau de fissures est un demi-graben basculé vers l'Ouest. Les réflecteurs et les vitesses sismiques ont été projetés parallèlement aux structures, à partir d'un profil sismique [113] situé 20 km au nord de la coupe interprétative. Localisation de la carte indiquée par un rectangle sur la Fig. 3a.

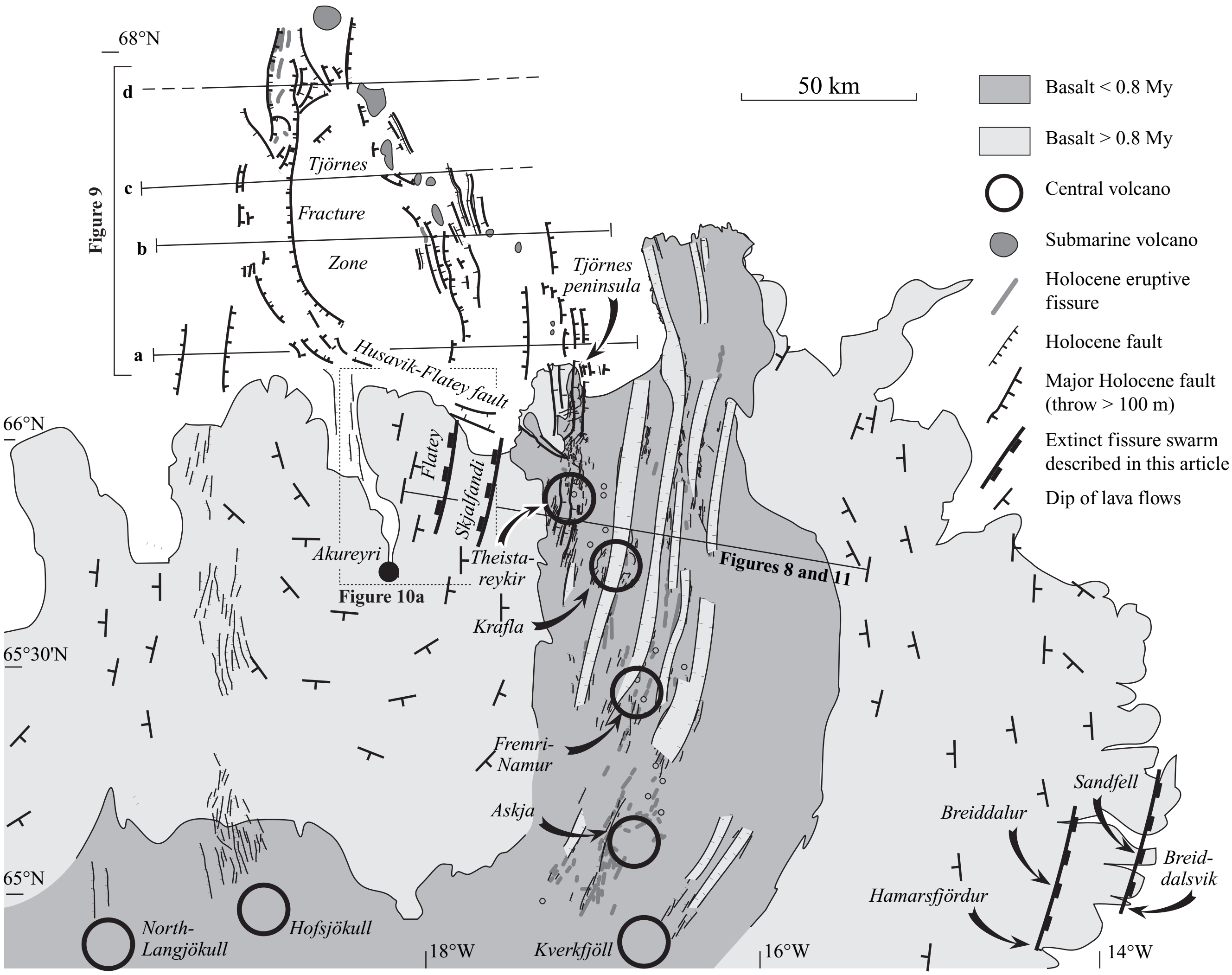

Fig. 7 Structural map of Holocene eruptive fissure swarms of NE-Iceland [77, 81, 108], with locations of cross-sections (Figs. 8, 9, 10 and 11) and of extinct fissure swarms described in the article (Flatey, Skjalfandi, Sandfell, Breiddalur).

Figure 7. a) Carte des faisceaux de fissures éruptives holocènes du NE de l'Islande [77, 81, 108], avec la localisation des coupes (Figs. 8, 9, 10 et 11) et des faisceaux de fissures inactifs présentés dans l'article (Flatey, Skjalfandi, Sandfell, Breiddalur).

As a conclusion, the rifting process is fundamentally asymmetrical and affects an area 200 km wide in Iceland. As the distance to the mantle plume increases, the width of the deformed area decreases and the structure becomes more symmetrical.

6. Deep structure of extinct fissure swarms

Glacial incision has excavated deep valleys along the coast of Iceland. These valleys provide natural cross-sections, both parallel and orthogonal to the structural grain, that are suitable for observing the deep structure of the plateau. The lava pile generally displays a wedge-shaped geometry with older flows dipping steeper (8-12°) than younger ones (0-2°) along a vertical section, and with flows getting thinner updip [89-95]. Locally, the dip of the lava pile increases up to 35°, forming anticlinal folds [40, 53, 92, 96].

Dike swarms, striking North to NE and cutting across the lava flows, are laterally connected to basic and/or acidic magma bodies [40, 89-92]. These dike swarms have been interpreted as the deep parts of ancient fissure swarms and the magma bodies as shallow magma chambers, once feeding ancient central volcanoes, that are now exposed by erosion [36]. Cross-sectional relationships between lava flows, faults, dike swarms and magma bodies therefore provide constraints on the deep structure of fissure swarms. To illustrate these relations, we describe four characteristic field cross-sections.

6.1. Flatey and Skjalfandi Valleys

The Flatey Valley is located in central northern Iceland and parallels the structural grain (Fig. 7). Folded lava flows

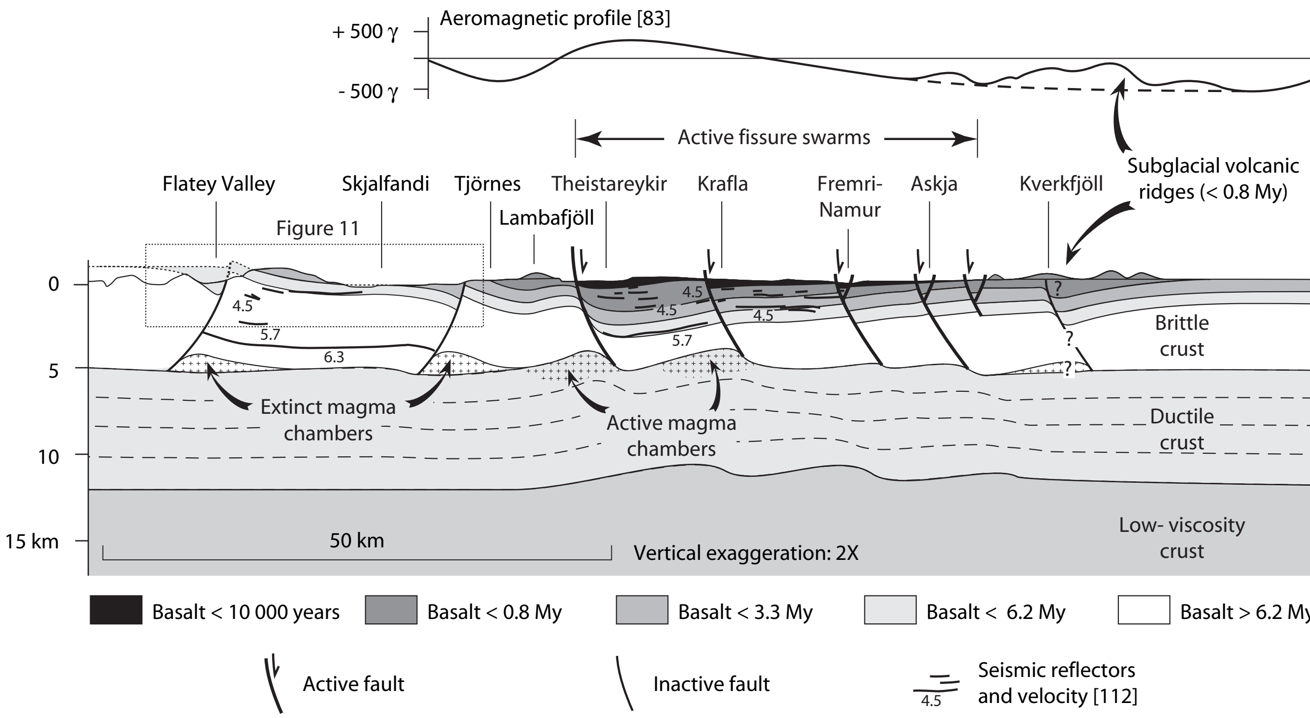

Fig. 8 Interpretative cross-section, Holocene eruptive fissure swarms of NE-Iceland (location indicated in Fig. 7). Active fissure swarms are half-grabens tilted westwards. Older extinct fissure swarms of Flatey and Skjalfandi Valley (Fig. 11) are half-grabens tilted eastwards. Dike swarms connecting magma chambes to lava flows not drawn for clarity. Seismic reflectors and velocities after [112]. Aeromagnetic profile after [83].

Figure 8. Coupe interprétative des faisceaux de fissures éruptives holocènes du NE de l'Islande (localisation indiquée sur la Fig. 7). Les faisceaux de fissures holocènes sont des demi-grabens basculés vers l'Ouest. Les faisceaux de fissures plus anciens et inactifs de Flatey et de Skjalfandi (Fig. 11) sont des demi-grabens basculés vers l'Est. Les faisceaux de dikes qui connectent les chambres magmatiques aux coulées ne sont pas dessinés par souci de lisibilité. Réflecteurs et vitesses sismiques d'après [112]. Profil aéromagnétique d'après [83].

are exposed on the western flank of the valley (Fig. 10). The youngest of these lava flows are 10.6 ± 0.6 to 10.0 ± 0.7 My old [96, 97]. These are cross-cut by a number of subvertical dikes, which are 7.5 ± 0.5 to 5.0 ± 0.7 My old [37, 38, 96]. On the eastern flank of the valley, the lava flows dip 10° east and there is a stratigraphic gap between 9.6 ± 0.5 My old lava flows at the base and 6.2 ± 0.3 to 4.1 ± 0.3 My old lava flows at the top [96]. The contact between the folded and the low-dipping flows on the two sides of the Flatey Valley has classically been interpreted as an unconformity; such an unconformity has not been observed directly however [37, 38, 96]. On the valley floor, between the folded and the low-dipping flows, we observed pervasively fractured basalts and intensely sheared dikes, forming an amalgam of sigmoidal lenses (Fig. 10c). Fractures and dikes dip 65°W, whereas lava flows dip 25°E. We re-interpret this contact as a fault zone and not as an unconformity. This fault zone can be traced northwards along the eastern flank of the Flatey Valley. At the foot of the Kambur summit (Fig. 10a), the fault zone separates acidic lava flows dipping 20° east, from low-dipping basic lava flows. Further north, the fault zone joins the extinct Nattfaravik Central Volcano (Fig. 10a) [40, 96].

Folding of the lava flows to the west of this fault zone can be attributed to the development of a rollover anticline in its hanging wall (Fig. 11). As the 10.6 ± 0.6 to 10.0 ± 0.7 My old lava flows on the hanging wall block were folded but do not thin updip, they are pre-kinematic. Whether syn-kinematic flows are partly preserved beneath the floor of the Flatey Valley, as is postulated in Fig. 11, is uncertain. They might have been removed totally by the carving of the valley.

In the footwall, we interpret the stratigraphic gap between the 9.6 ± 0.5 and younger flows as indicating the contact between pre-kinematic and syn-kinematic flows, with the age of syn-kinematic flows (6.2 ± 0.3 to 4.1 ± 0.3 My) constraining the age of fault activity. This age is similar to that of the Nattfaravik Central Volcano (6.2 ± 0.3 to 4.9 ± 0.2 My [96]) and to that of dikes occurring in the folded hanging wall (7.5 ± 0.5 to 5.0 ± 0.7 My [37, 38, 96]). Thus, activation of Flatey Valley fault, folding of the hanging wall, injection of dikes and emplacement of Nattfaravik Central Volcano were coeval. This volcanic system remained active for 3 My.

Below the Engjafjall and Kinnarfjöll mountains, the lava flows dip gently eastwards (Figs. 10 and 11). Further east they are bent downwards and ultimately reach a dip of 15° east in the Skjalfandi Valley. The 3.3-0.8 My old basalts exposed in this valley are pervasively fractured and cross-cut by intensely sheared dikes dipping 75° to 85° west. The $^{39}Ar/^{40}Ar$ age of a dike of this swarm is 2.3 ± 0.3 My [38]. This

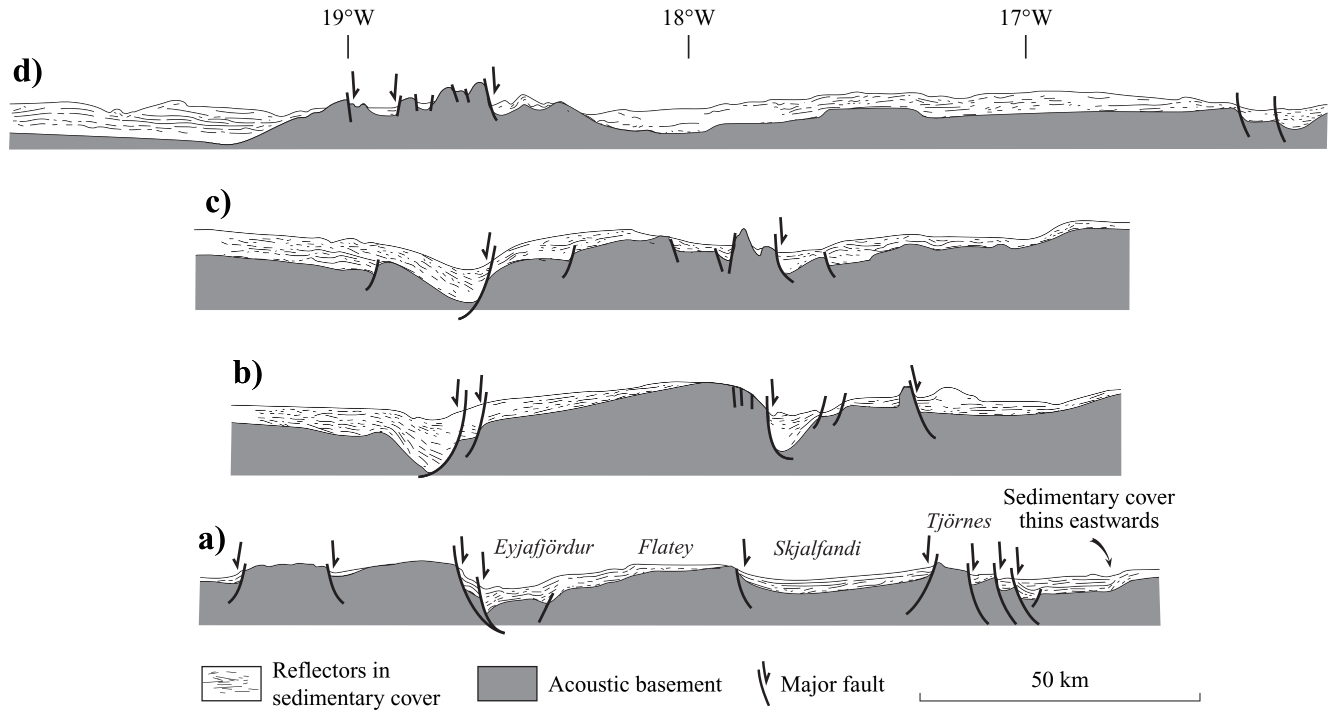

Fig. 9 Seismic profiles across plate boundary, from N-Iceland to Kolbeinsey Ridge (re-interpreted after McMaster *et al.* [81]). Profile locations indicated in Fig. 7. Map location indicated by box in Fig. 3a.

Figure 9. Profils sismiques perpendiculaires à la limite de plaques, depuis le Nord de l'Islande jusqu'à la Ride de Kolbeinsey (réinterprétés d'après McMaster *et al.* [81]). Les profils sont localisés sur la Fig. 7.

suggests the occurrence of another younger growth fault/ rollover system in the Skjalfandi Valley (Fig. 11).

6.2. Breiddalur

Breiddalur is a glacial valley, located in eastern Iceland, that strikes orthogonal to the structural grain (Figs. 3c and 7). A NNE striking dike swarm cuts across the valley head and can be followed south and north across a number of other valleys striking WNW [40, 90-92]. This dike swarm is connected laterally to the extinct Breiddalur central volcano, which is extensively exposed on the southern flank of the Breiddalur Valley [90]. K/Ar measurements yield an age of 8.9 ± 0.8 My for a microgranitoid block contained in one of the latest products of this central volcano [99].

The cross-sectional structure of this old fissure swarm is perfectly exposed on the northern flank of the Breiddalur Valley, near Thorgrimststadir (Fig. 12). Lava flows are sub-horizontal in the far eastern part of the cliff, and dip gently westwards in its central part. This dip increases downwards and westwards, up to a value of 12° in the lower part of the cliff, with lava flows getting thinner updip. A number of minor normal faults and dikes cut across the central part of the section. The west-dipping lava flows abut against a gully in the west. West of the gully, lava flows are horizontal and dikes are scarce.

Walker [90] interpreted the contact between the west-dipping lava flows and the horizontal lava flows as an unconformity between lava flows emplaced on the slopes of the volcano and latter flood basalts covering up the volcano. He admitted, however, that the unconformity could not be observed in the cliff. In the gully, we observed a thick complex of amalgamated dikes, providing evidence for focused and renewed magma injection. The dikes form sigmoid lenses that are covered by slickensides, indicating intense shearing. Also broken pieces of lava flows and brecciaed rocks are abundant. We interpret this contact as a major east-dipping fault zone, into which dikes were injected during or before faulting, rather than as an unconformity. The detailed relative chronology between the beginning of faulting and the injection of magma in the dikes remains unknown.

In the hanging wall (east), the lava flows display a wedge-shaped geometry that can be attributed to progressive tilting during their syn-kinematic extrusion, and to the development of a rollover anticline in the hanging wall of the fault. Correspondingly, the age of the fault is constrained by the age of syn-kinematic lava flows. To our knowledge, no age determinations are available for the lava flows of the Breiddalur Valley. K/Ar ages are available, however, for stratigraphically equivalent flows in Hamarsdalur (Fig. 3c) [98]. Along-strike extrapolation of these ages tentatively yields an age of 12.1 ± 1 My for lava flows in the footwall and an age of 10.6 ± 0.5 My for syn-kinematic lava flows in the hanging wall (Fig. 12). An agglomerate emplaced during one of the latest eruptions of the Breiddalur Central Volcano contains blocks of microgranitoid intrusions, the K/Ar age of

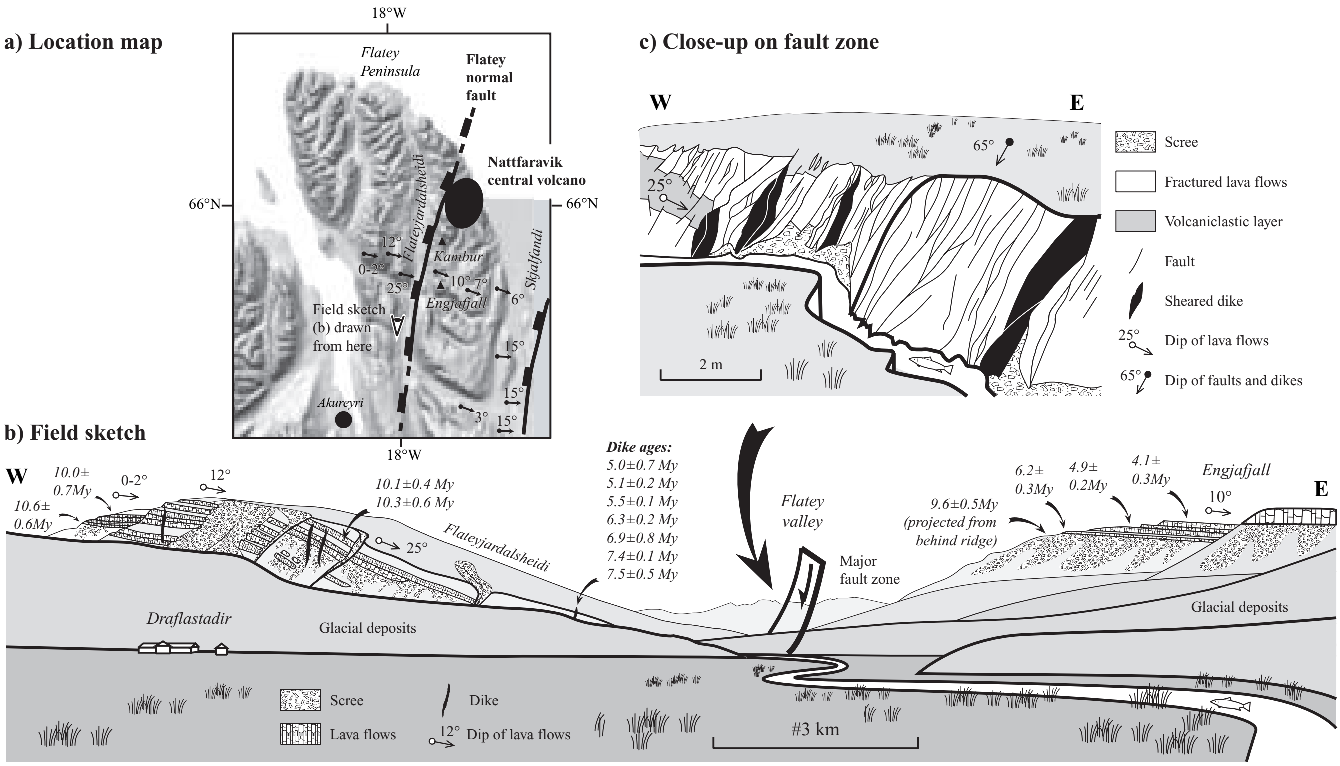

Fig. 10 Field cross-section, Flatey Valley. **a)** Structural sketch map (location indicated by dotted box in Fig. 7). **b)** General view, looking northwards from road 835 near Draflastadir. The valley marks the location of a major fault zone between folded lava flows in the west and low-dipping lava flows in the east. In the hanging wall (Flateyjardalsheidi), lava flows are 10.6 to 10.0 My old; in the footwall (Engjafjall), there is a stratigraphic gap between 9.6 My old and 6.2-4.1 My old lava flows [96, 97]. There are 7.5-5.1 My old dikes in the hanging wall and along the fault zone [38, 96]. **c)** Close-up view of fault zone (stream at intersection of roads 835 and F899) showing intensely fractured lava flows and sigmoidal dikes. Outcrop faces North; sketch has been reversed, so that that its orientation fits orientation of general view.

Figure 10. Coupe de terrain, vallée de Flatey. **a)** Carte structurale schématique (localisation indiquée sur la Fig. 7). **b)** Vue générale, en regardant vers le Nord depuis la route 835 près de Draflastadir. La vallée est située sur une zone de faille majeure qui sépare des coulées de lave pliées et fortement basculées à l'Ouest, de coulées faiblement inclinées à l'Est. Dans le compartiment supérieur (Flateyjardalsheidi), les coulées de lave ont un âge de 10,6 à 10,0 Ma ; dans le compartiment inférieur (Engjafjall), il y a une lacune stratigraphique entre les coulées datant de 9,6 Ma et celles datant de 6,2 à 4,1 Ma [96, 97]. Il y a des dikes datant de 7,5 à 5,1 Ma dans le compartiment supérieur et le long de la zone de faille [38, 96]. **c)** Vue rapprochée de la zone de faille (ruisseau à l'intersection des routes 835 et F899) montrant les coulées de lave intensément fracturées et les dikes sigmoïdaux. L'affleurement fait face au Nord ; le schéma est présenté à l'envers, de telle manière que son orientation corresponde à l'orientation de la vue générale.

which is 8.9 ± 0.8 My [99]. Thus the development of the fault can be linked to the extrusion of volcanic products and to the emplacement of acidic intrusions in the Breiddalur Central Volcano, between 11 and 8 My.

6.3. *Breiddalsvik*

Another NNE-striking dike swarm extends along the eastern coast of Iceland and crosses the mouth of the Breiddalur Valley (Figs. 3c and 7) [40, 89-92]. This dike swarm includes several amalgamated dike complexes, one of which is visible on the cliff north of Breiddalsvik (Fig. 13). This complex includes three parallel dikes dipping 80°E. Each dike is composed of vertical *en echelon* sigmoidal lenses that are covered by slickensides, indicating normal dip-slip displacements. Lava flows are vertically offset across this dike complex. The geometry of the dike lenses is consistent with theoretical dilation fractures produced during normal faulting. Thus, we interpret the dike complex as a normal fault zone, into which magma was injected during faulting. The sigmoidal shape of dikes has been perfectly preserved, because the amount of displacement on this fault zone was much less than on the Flatey and Breiddalur fault zones. Unfortunately we are not able to constrain the age of the fault activation, nor that of the dike emplacement. Our observations only show that some deformation took place after the crystallisation of the dike magma and we still do not know how much deformation, if any, occurred before the dike emplacement.

6.4. *Sandfell*

The Sandfell Mountain is a microgranitoid body, a laccolith-shaped intrusion that is connected to the Breiddalsvik dike swarm described above (Figs. 3c and 7) [40]. This mountain has an upper and a lower summit (Fig. 14a). The petrology of the microgranitoid body and its structural rela-

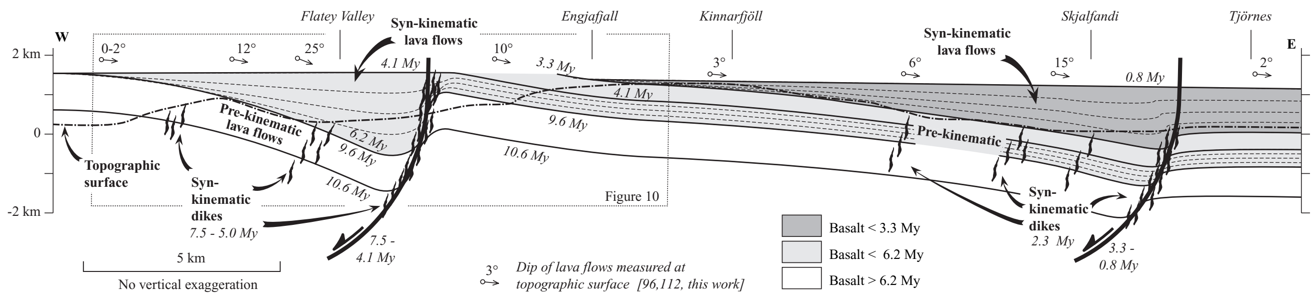

Fig. 11 Interpretative cross-section across Flatey and Skjalfandi Valleys (location indicated in Fig. 7). Overall structure interpreted as two adjacent successive growth fault/rollover systems. Injection of dikes (between 7.5 and 5.1 My in Flatey Valley; around 2.3 My in Skjalfandi Valley) and emplacement of lava flows (between 6.2 and 4.1 My in Flatey Valley; between 3.3 and 0.8 My in Skjalfandi Valley) accompanied development of faults. Pre-kinematic lava flows (> 9.6 My old in Flatey Valley; > 3.3 My old in Skjalfandi Valley) passively bent in response to development of rollover anticlines in hanging walls. Syn-kinematic lava flows tilted towards faults, as they progressively filled hanging walls. Location of field sketch (Fig. 10) indicated by dotted box.

Figure 11. Coupe interprétative à travers les vallées de Flatey et de Skjalfandi (localisation indiquée sur la Fig. 7). L'ensemble de la structure est interprété comme deux systèmes « faille de croissance/anticlinal en roll-over » successifs et adjacents. L'injection des dikes (entre 7,5 et 5,1 Ma dans la vallée de Flatey, autour de 2,3 Ma dans la vallée de Skjalfandi) et la mise en place des coulées de lave (entre 6,2 et 4,1 Ma dans la vallée de Flatey ; entre 3,3 et 0,8 Ma dans la vallée de Skjalfandi) ont accompagné le développement des failles. Les coulées pré-cinématiques (> 9,6 Ma dans la vallée de Flatey, > 3,3 Ma dans la vallée de Skjalfandi) ont été pliées passivement en réponse au développement des anticlinaux en roll-over dans les compartiments supérieurs. Les coulées syn-cinématiques ont été continuellement basculées en direction des failles, au fur et à mesure qu'elles remplissaient les compartiments supérieurs. La localisation de la coupe de terrain (Fig. 10) est indiquée par le rectangle en pointillés.

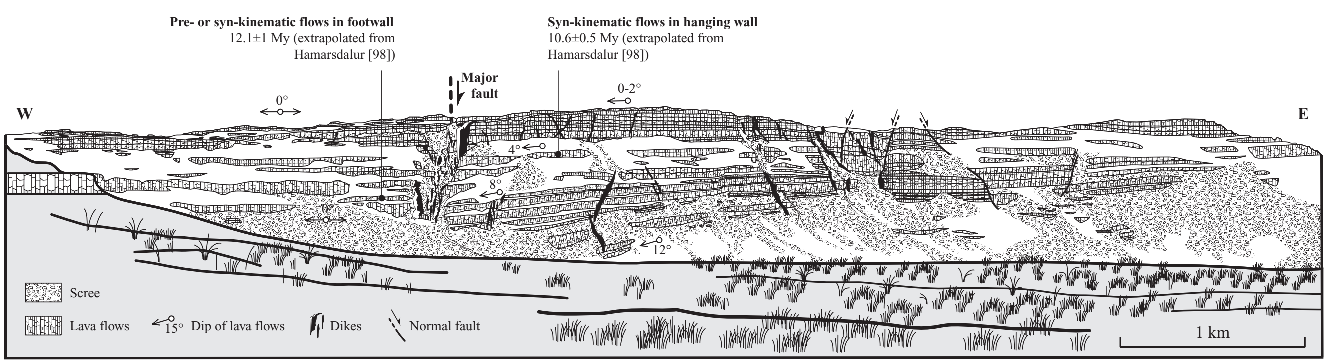

Fig. 12 Field cross-section, northern flank of Breiddalur Valley near Thorgrimsstadir. Western part of section composed of horizontal lava flows, 12.1 ± 1 My in age. Major fault zone associated with complex of amalgamated dikes in gully. Syn-kinematic lava flows, 10.6 ± 0.5 My in age, folded into rollover anticline in hanging wall (eastern part of section). Subsidiary dikes and faults accommodate folding of hanging wall. Lava ages tentatively extrapolated along strike from K/Ar measurements in Hamarsfjördur [98].

Figure 12. Coupe de terrain, versant nord de la vallée de Breiddalur près de Thorgrimsstadir. La partie ouest de la coupe est composée de coulées horizontales datant de 12,1 ± 1 Ma. Dans la partie centrale, une faille majeure associée à un complexe de dikes amalgamés se trouve dans un ravin. Les coulées syn-cinématiques, datant de 10,6 ± 0,5 Ma, forment un anticlinal en roll-over dans le compartiment supérieur de la faille (partie est de la coupe). Des dikes et des failles secondaires accommodent le plissement du compartiment supérieur. Les ages des laves sont extrapolés à titre indicatif, à partir de mesures K/Ar effectuées dans la vallée de Hamarsfjördur [98].

tionships with the surrounding lava flows were described in details by Hawkes and Hawkes [101] and by Gibson *et al.* [109]. East and north of the Lower Sandfell, the lava flows dip gently westwards. At the contact with the microgranitoid body, they are abruptly turned-up and dip 80°NE. Here, the microgranitoid body and the host lava flows display a penetrative foliation parallel to the contact. Following Hawkes and Hawkes [101] we interpret this contact as a shear zone, corresponding to an east-dipping normal fault.

The Upper and Lower Sandfell are capped by horizontal lava flows. The difference in elevation of the capping lava flows indicates the presence of another east-dipping normal fault between the two summits. On the western and southern flanks of the Upper Sandfell, the lava flows capping the intrusive body display dips of up to 45°SW before they flatten out west, forming a syncline, and ultimately recover a horizontal attitude. On the western and southern slopes of the Sandfell, there is no evidence of deformation

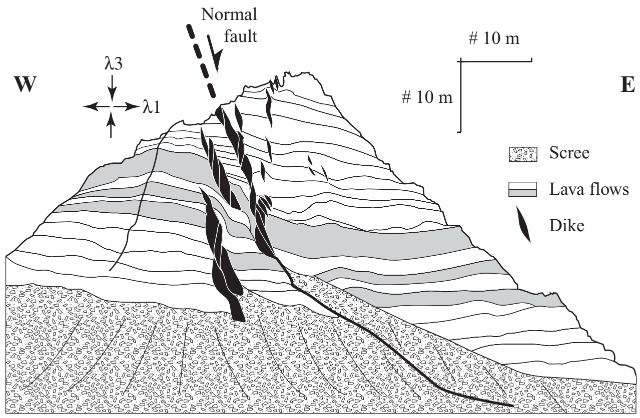

Fig. 13 Field sketch, complex of amalgamated dikes associated with normal fault (road 96 near Breiddalsvik, N64°27.899', W13°53.586'). Complex includes three parallel strips composed of vertical lenses disposed *en echelon*. Attitude of lenses consistent with theoretical attitude of dilation fractures produced during normal faulting (principal stretching axis λ1 horizontal, principal shortening axis λ3 vertical). Vertical offset across fault-dike complex emphasized by three shaded lava flows.

Figure 13. Schéma de terrain, complexe de dikes amalgamés associé à une faille normale (route 96 près de Breiddalsvik, N64°27.899', W13°53.586'). Le complexe comprend trois bandes parallèles composées de lentilles verticales disposées en échelon. La disposition des lentilles est compatible avec la disposition théorique de fentes de tension produites lors du fonctionnement normal d'une faille (axe principal d'allongement λ1 horizontal, axe principal de raccourcissement λ3 vertical). Le décalage vertical de part et d'autre du complexe faille/dike est souligné par les trois coulées grisées.

along the contact between the intrusive body and the capping lava flows.

These observations indicate that the emplacement of the Sandfell intrusion was linked with the development of east-dipping normal faults along its eastern margin (Fig. 14b). The hanging wall (east) has been gently tilted towards the fault zone, whereas the footwall (west) was bent upwards to accommodate the emplacement of the intrusive body into the fault zone.

7. Tectonic model

7.1. Structure and development of a fissure swarm

We assume that Holocene fissure swarms and extinct dike swarms represent the shallow and deeper parts of a volcanic system, respectively. We also assume that the field outcrops described above correspond to cross-sections at various crustal levels of an ideal fissure swarm (Fig. 15). Our model fissure swarm involves a growth fault/rollover system. The growth of a rollover anticline requires that motion along the controlling fault is coeval with infilling of the hanging wall accommodation space and with uplift of the footwall, entailing an uplift of the brittle/ductile transition zone [102]. Thus the development of such an Icelandic volcanic system involves the following mechanisms, which are essentially coeval on a time scale of a few thousand years. Points (a) to (h) refer to letters in Fig. 15a.

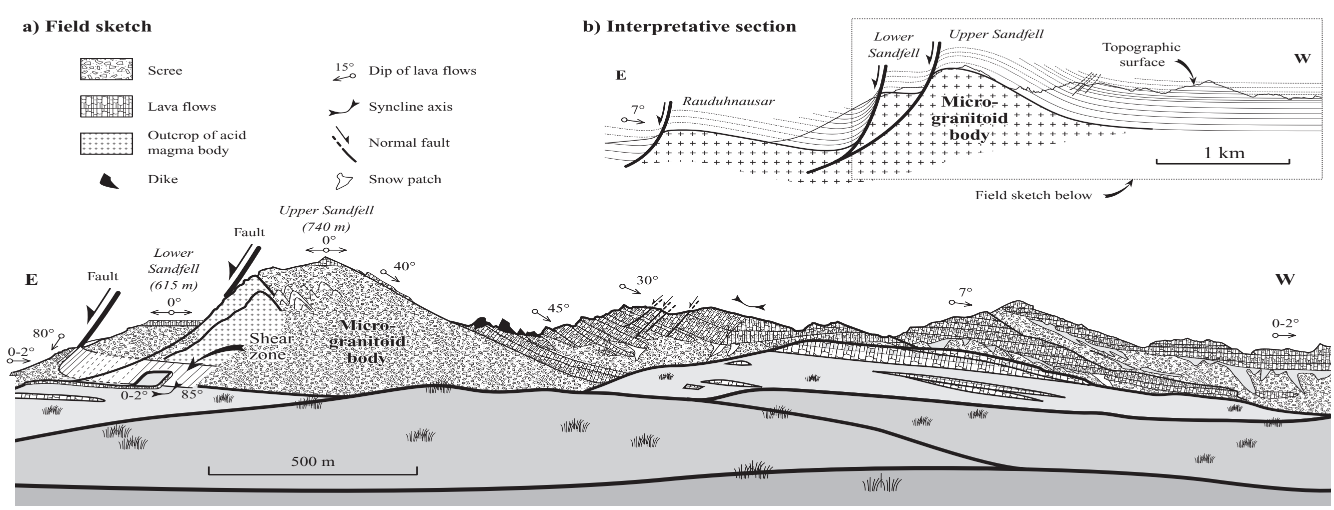

Fig. 14 Field cross-section, Sandfell micrograntoid body (N64°52.94', W13°54.34'). **a)** Field sketch, looking from NW. Note abrupt folding of lava flows and contact-parallel foliation of magma body on northern slope of Upper Sandfell and on eastern slope of Lower Sandfell. **b)** Interpretative section. Western part (box) interpreted after our own observations. Eastern part interpreted after field descriptions and maps of Hawkes and Hawkes [101] and Gibson *et al.* [109].

Figure 14. Coupe de terrain, intrusion de microgranitoïde du Sandfell (N64°52.94', W13°54.34'). **a)** Schéma de terrain, vu du NW. Noter le plissement abrupt des coulées et la foliation de l'intrusion sur le versant nord du Sandfell supérieur et sur le versant est du Sandfell inférieur. **b)** Coupe interprétative. La partie ouest (rectangle) est interprétée d'après nos propres observations. La partie est est interprétée d'après les descriptions d'affleurements et les cartes de Hawkes et Hawkes [101] et de Gibson [109].

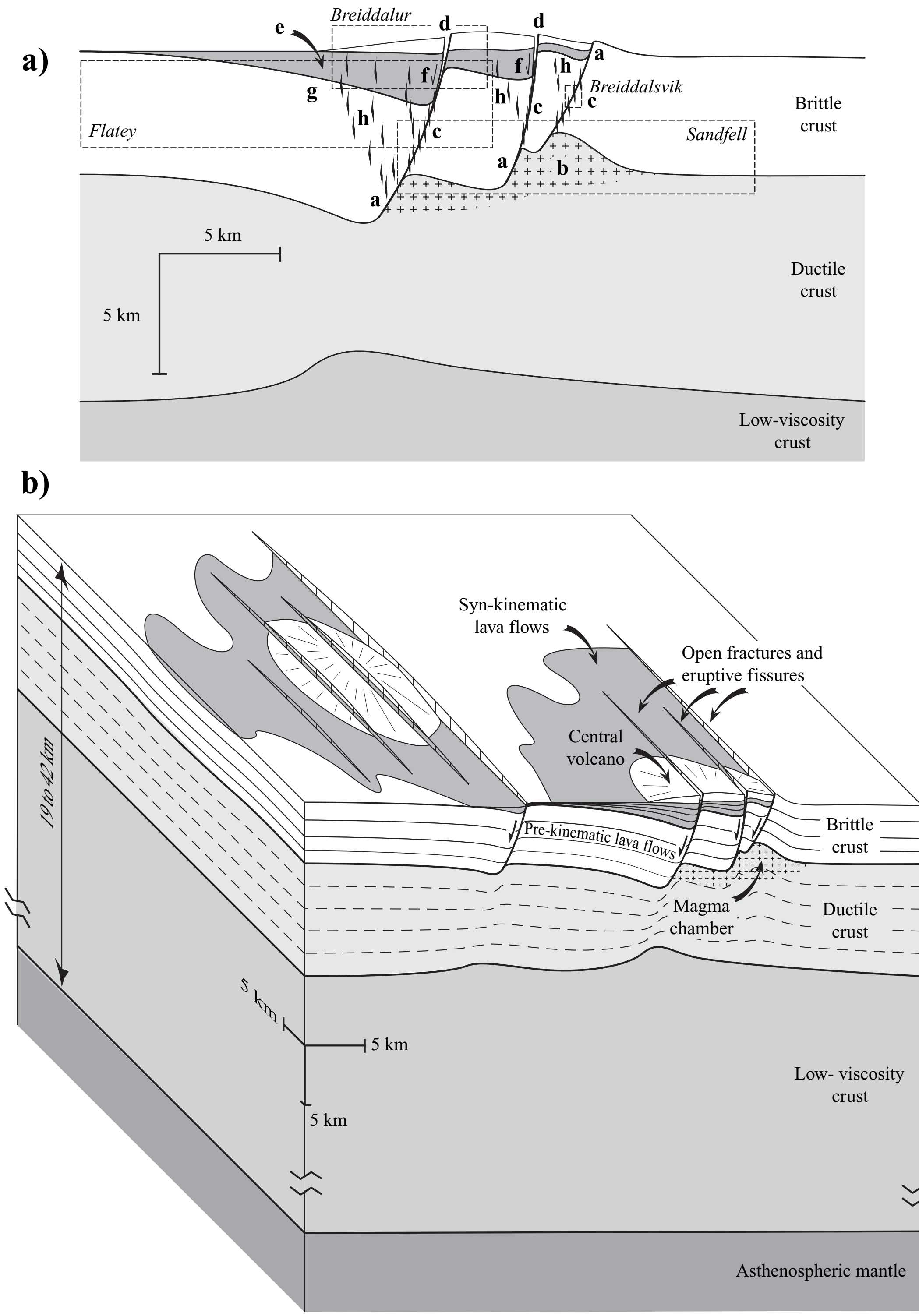

Fig. 15 Tectonic model. **a)** cross-section of ideal fissure swarm. Field outcrops described in text (indicated by dashed boxes) correspond to cross-sections at various crustal levels of growth fault/rollover system. (a) growth faults; (b) magma chamber; (c) amalgamated dike complex; (d) eruptive fissures and central volcano; (e) syn-kinematic lava flows; (f) subsidence of hanging wall along fault; (g) development of rollover anticline by folding of hanging wall; (h) subsidiary dikes and fractures in hanging wall. **b)** 3D-view of two coeval *en echelon* fissure swarms. Dikes not drawn for clarity.

Figure 15. Modèle tectonique. **a)** Coupe d'un faisceau de fissures idéal. Les affleurements de terrain décrits dans le texte (indiqués par les rectangles en pointillés) correspondent à des coupes à différentes profondeurs d'un système faille de croissance / anticlinal en roll-over. (a) failles de croissance; (b) chambre magmatique ; (c) complexe de dikes amalgamés ; (d) fissures éruptives et volcan central ; (e) coulées de lave syn-cinématiques ; (f) subsidence du compartiment supérieur le long de la faille ; (g) développement de l'anticlinal en roll-over par plissement du compartiment supérieur ; (h) dikes et fractures secondaires dans le compartiment supérieur. **b)** Bloc-diagramme montrant deux faisceaux de fissures synchrones et disposés en échelon. Les dikes ne sont pas dessinés par souci de lisibilité.

a. Lithospheric stretching causes the development of listric fault zones in the brittle crust; the faults root in the ductile crust.
b. A magma chamber develops at the base of the brittle crust below the fault zone. With increasing strain along the fault zone, the ductile crust and the magma chamber rise in the footwall.
c. Magma injection from the magma chamber towards the surface is focused along the faults and leads to the formation of amalgamated dike complexes.
d. Magma reaching the topographic surface erupts in a fissure swarm and in a central volcano, both located above the fault zone.
e. Lavas extruded via the fissure swarm and the central volcano successively cover the hanging wall.
f. The hanging wall subsides progressively, in response to the combined effects of lithospheric stretching and its loading by lava flows.
g. Lava flows of the hanging wall are passively bent into a rollover anticline that reflects the listric geometry of the controlling fault.
h. Bending of the hanging wall gives rise to the development of secondary faults and fractures, along which subsidiary dikes can be injected.

7.2. Development of successive fissure swarms

Ages of syn-kinematic lava flows and dikes in the field sections described above indicate that a given growth fault/rollover system remains active during a few My, and then becomes inactive. Shifting of tectonic and magmatic activity throughout the island during the last few My suggests that, when a given system becomes inactive, another system located within the deformation strip becomes active.

The distribution of ages of extinct volcanic systems and the dips of lava flows observed in Iceland have been classically attributed to a complex history of lateral shifts of the rift axis [36-38, 45, 46, 53, 107]. Rift relocations were thought to be the effects of westwards migration of the Mid-Atlantic Ridge over the axis of the plume. Alternatively in our model, the age and dip distribution of lava flows are controlled by the order of activation and polarity of successive growth fault/rollover systems within a long-term deformation strip (Fig. 16). Shifting of activity is intrinsic to the deformation process.

The mechanism that controls the abandonment of an active growth fault/rollover system at the expense of the development of an adjacent new one remains uncertain. We suggest two possible explanations. (1) As it drifts laterally away from the plate boundary, an active growth fault/rollover system cools down, hardens and might ultimately lock when it gets out of the long-term deformation strip. Then a new system will develop in the hotter and weaker crust of the deformation strip. This explanation might hold for systems labelled 1, 2, 4, 6, 7, 10 and 11 in Fig. 16. (2) In our model, extension-induced development of lateral pressure gradients controls the accumulation of lower-crust derived partial melts at the base of the brittle upper crust, from where they ascent to the surface along fault systems. Once the partial melt reservoirs in the lower crust are depleted, and the lower crust is uplifted and decompressed below the footwall of a growth fault/rollover system, this system hardens and ultimately locks. As such a system is probably rheologically stronger than the adjacent un-stretched lithospheric segment, persisting extensional stresses will control the activation of a new growth fault/rollover system that is laterally offset from the older one. This explanation might hold for systems labelled 3, 5, 8, 9 and 12 in Fig. 16.

Lava flows emplaced from new systems may partially overlap previous systems, thus creating unconformities in the lava pile. In that sense, the suggested model is consistent with other models that assume shifting of activity [45, 46, 64, 65]. In our model however, several systems within the deformation strip are active at the same time (Fig. 16). The process of rifting that occurs in the anomalously thick and hot oceanic crust of Iceland is somewhat analogous to the wide-rift mode that develops in orogenically thickened and thermally weakened continental crust as, for example, in the Basin and Range Province and in the Aegean Sea [49, 50, 104, 105].

7.3. Implications of model for building of the plateau

Crustal accretion is not focused in a narrow active zone pinned at the spreading axis, but shifts from fissure swarm to fissure swarm within a 200 km wide diffuse plate boundary. Hence, unlike other slow-spreading ridges (Fig. 1), an axial valley bounded by large fault scarps cannot develop (Fig. 16).

In the suggested model, the ductile crust and the magma chamber rise passively beneath the footwall, until the growth fault/rollover system becomes inactive (Fig. 15). Subsequently, the top of the magma chamber may be exposed by erosion. Rising of the ductile crust is analogous to the ascent of salt rollers in the footwalls of sedimentary growth faults that form above salt decollement layers [102,103]. It is analogous also to the ascent of metamorphic core complexes that develop in orogenically thickened continental crust [49,50,104,105]. Rising of the ductile crust in Iceland is probably favoured by its high temperature and by the presence of partial melts, which decreases the viscosity of rocks in magma chambers [103,106].

This model allows pre-kinematic lava flows in the hanging wall to subside and to melt in the vicinity of the magma chamber (Fig. 15). This mechanism possibly provides an explanation for the production of acidic magmas that are extruded in Icelandic central volcanoes. In contrast, the footwall does not subside. Hence lava flows of the footwall remain close to the surface and shallow metamorphic minerals, such as zeolites, can grow in their pores. As successive growth fault/rollover systems cannot form along a pinned axis, lava flows do not pile up at the same place. Our model

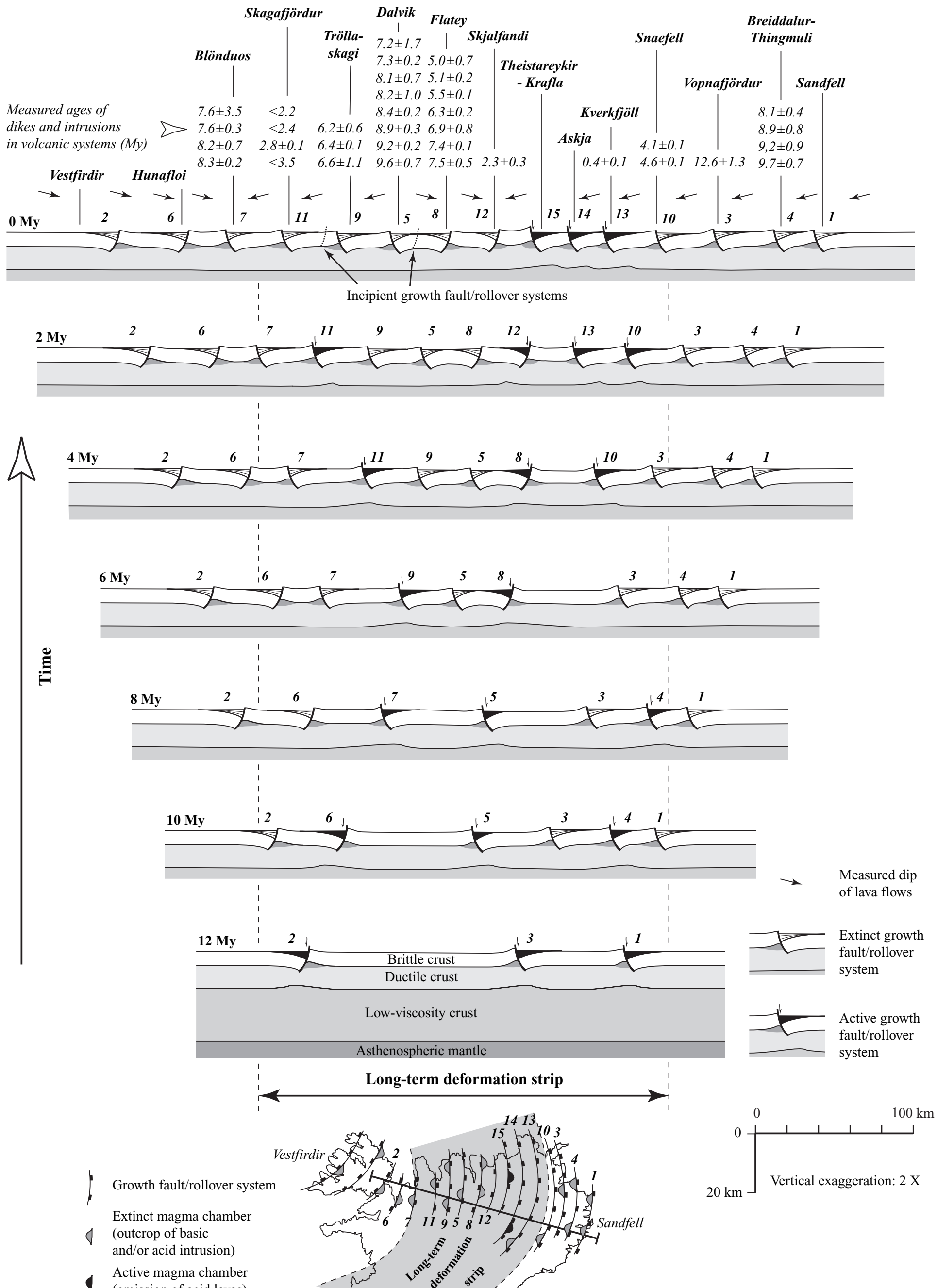

Fig. 16 Development of the Iceland plateau at 2 My time intervals along cross-section A-A'. Volcanic systems comprise a growth fault, a rollover anticline and a magma chamber (dike swarms connecting magma chambers to lava flows not drawn for clarity). Volcanic systems develop successively, they drift as they grow¸ and they become inactive; at any given time, crustal spreading is accommodated by several systems distributed within a 200 km wide deformation strip. The complex distribution of ages [38, 96, 99, 100] and of dips [40] observed in Iceland is controlled by order of appearance, polarity and place of birth of successive volcanic systems. Names above last cross-section indicate present-day locations of drawn volcanic systems. Dotted faults indicate location of current non-eruptive fault swarms interpreted as incipient growth fault/rollover systems. Systems numbered in order of appearance.

Figure 16. Développement du plateau islandais à intervalles de 2 Ma, le long de la coupe A-A'. Les systèmes volcaniques sont composés d'une faille de croissance, d'un anticlinal en roll-over et d'une chambre magmatique (les faisceaux de dikes qui connectent les chambres magmatiques aux coulées ne sont pas dessinés par souci de lisibilité). Les systèmes volcaniques apparaissent successivement, ils se déplacent latéralement pendant qu'ils croissent, puis ils meurent. À tout moment, l'accrétion crustale est accommodée sur plusieurs systèmes distribués dans une bande de déformation de 200 km de large. La distribution des ages [38, 96, 99, 100] et les pendages [40] observés en Islande sont déterminés par l'ordre d'apparition, la polarité et le lieu de formation des systèmes volcaniques successifs. Les noms au-dessus de la dernière coupe indiquent l'emplacement actuel des systèmes volcaniques dessinés. Les lignes pointillées indiquent la localisation des faisceaux de failles non éruptives actuels, qui sont interprétés comme des systèmes faille de croissance / anticlinal en roll-over naissants. Les systèmes sont numérotés selon leur ordre d'apparition.

does not predict deep burial of extinct volcanic systems (Fig. 16).

8. Conclusion

The cross-sectional structure of the Iceland plateau differs from the symmetric pattern of outwards-tilted blocks generally observed at slow-spreading ridges. Field relations between lava flows, dikes, magma bodies and faults provide evidence that the plateau has developed by activation, growth and decay of individual growth fault/rollover systems underlain by shallow magma chambers. These systems formed successively at different places within a 200 km wide diffuse plate boundary. This remarkable process of rifting disappears as the distance to the mantle plume increases. The plume may possibly alter the rifting process in two fashions. (1) The plume increases the temperature of the nascent lithosphere; hence it decreases its strength. (2) The plume promotes magma supply from the mantle to the crust; hence addition of material to the crust is not balanced by its stretching as in other slow-spreading ridges. Modelling would help to assess if these mechanisms are actually responsible for this deformation process, which is specific to Iceland.

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