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Reverse-slip structures at oceanic diverging plate boundaries and their kinematic origin: data from Tertiary crust of west and south Iceland

Maryam Khodayar*,[1], Páll Einarsson

Science Institut, University of Iceland, Hofsvallagata 53, 107 Reykjavik, Iceland

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Abstract

Normal faulting predominates within rift zones, but reverse fault plane solutions are occasionally found for earthquakes along the Mid-Atlantic Ridge. Furthermore, field observations of striae and in situ measurements in places show off-rift compression and reverse faulting. We analyse eight cases of reverse-slip motion observed in the eroded Tertiary crust of the Borgarfjörður and Hreppar rift-jump blocks, respectively, in west and south Iceland. These structures were the only ones observed where both striae and marker horizons could be used to deduce the sense of reverse-slip motion. The vertical displacements are less than 10 m, and the motions are of local origin. We believe these structures are formed independently from regional compression, as they have variable kinematic origins. Some occur in association with dykes, sills, or cone-sheet, without evidence of horizontal shortening across intrusions; others are due to local bends of regional steeply-dipping normal faults or secondary fractures (Riedel or secondary fractures). Reverse-slip striations are found at the irregularities of steeply-dipping normal faults, or at the tips of steeply-dipping normal faults uplifted by an underlying propagating dyke. Striations, even conspicuous ones, do not necessarily reflect significant amounts of displacement. If horizontal compression acts in Iceland, its magnitude is not sufficient to shorten the rocks and to close the fractures of various tectonic settings. The observed examples are explained as local kinematic adjustments to companion structures, without regional compressional stress field or shortening, and can lead to misinterpretation if not put into the correct context.

Keywords: Reverse-slip structures; Compression; Dyke injections; West Iceland; South Iceland; Oceanic plate boundary

1. Introduction

Oceanic crust is generally formed at diverging plate boundaries under conditions of rifting and crustal extension. Crustal faults and focal mechanisms of earthquakes reflect this. Almost all fault plane solutions from ridge crest areas show normal faulting (e.g. Sykes, 1967; Einarsson, 1986). However, there are notable exceptions. Einarsson (1979) identified two sources of reverse faulting mechanisms along the Mid-Atlantic plate boundary using teleseismic data. One area is in the caldera region of the Bárðarbunga volcano near the centre of the Iceland hotspot, and the other is at the ridge crest south of the Charlie–Gibbs fracture zone. More recently, a large swarm of apparently reverse faulting earthquakes occurred at the Mid-Atlantic Ridge crest near 34°N (USGS Earthquake Files, 1999). An ocean-bottom seismograph study of micro earthquakes at the MAR crest near 26°N also revealed reverse faulting beneath the axial volcano in the rift valley, with normal faulting occurring on adjacent faults (Kong et al., 1992). These anomalous fault plane solutions are attributed to forceful intrusions within central volcanoes or uplift of rift mountains (Einarsson, 1979), faulting on curved planes above a deflating magma chamber (Einarsson, 1991), or faulting on conical planes beneath an inflating magma chamber (Nettles and Ekström, 1998).

The general belief that compression exists in Iceland is mostly based on the observation of reverse-slip striations on fault planes in several Tertiary and Quaternary areas (e.g. Bergerat et al., 1990), as well as on the hydrofracturing stress measurements in few areas off-rift (Haimsson and Rummel, 1982). If compression acts severely in Iceland, its corresponding structures should commonly be observed. However, only few direct geological observations report reverse faulting and compressional structures other than

* Corresponding author.
[1] Now at ÍSOR-Íslenskar orkurannsóknir (Iceland Geosurvey), Orkugardur, Grensasvegur 9, 108 Reykjavik, Iceland.
E-mail address: mak@isor.is

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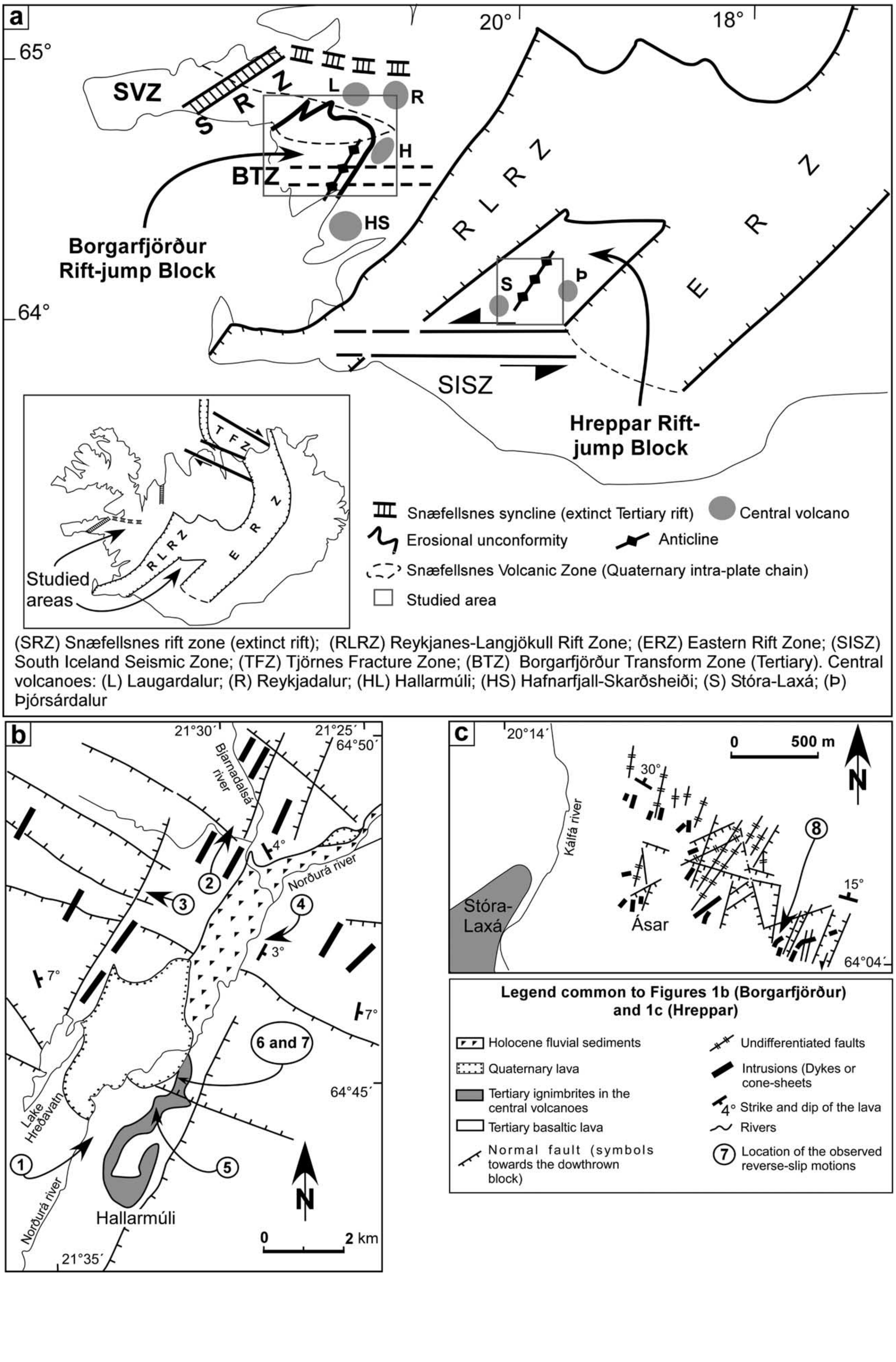
a
20°
18°
65°
64°
SVZ
S R Z
L
R
H
BTZ
HS
Borgarfjörður Rift-jump Block
R L R Z
E R Z
S
Þ
SISZ
Hreppar Rift-jump Block
T F Z
R L R Z
E R Z
Studied areas
Snæfellsnes syncline (extinct Tertiary rift)
Central volcano
Erosional unconformity
Anticline
Snæfellsnes Volcanic Zone (Quaternary intra-plate chain)
Studied area
(SRZ) Snæfellsnes rift zone (extinct rift); (RLRZ) Reykjanes-Langjökull Rift Zone; (ERZ) Eastern Rift Zone; (SISZ) South Iceland Seismic Zone; (TFZ) Tjörnes Fracture Zone; (BTZ) Borgarfjörður Transform Zone (Tertiary). Central volcanoes: (L) Laugardalur; (R) Reykjadalur; (HL) Hallarmúli; (HS) Hafnarfjall-Skarðsheiði; (S) Stóra-Laxá; (Þ) Þjórsárdalur
b
21°30´
21°25´
64°50´
Bjarnadalsá river
4°
Norðurá river
2
3
4
3°
7°
7°
6 and 7
64°45´
Lake Hreðavatn
1
5
Norðurá river
Hallarmúli
N
0
2 km
21°35´
c
20°14´
0
500 m
N
Kálfá river
30°
8
15°
Stóra-Laxá
Ásar
64°04´
Legend common to Figures 1b (Borgarfjörður) and 1c (Hreppar)
Holocene fluvial sediments
Quaternary lava
Tertiary ignimbrites in the central volcanoes
Tertiary basaltic lava
Normal fault (symbols towards the dowthrown block)
Undifferentiated faults
Intrusions (Dykes or cone-sheets
4° Strike and dip of the lava
Rivers
7 Location of the observed reverse-slip motions

striae. Along the active plate boundaries, these structures are metre-scale push-ups associated with strike-slip faults in the South Iceland Seismic Zone (Einarsson and Eiríksson, 1982; Bjarnason et al., 1993), and in the oblique rifting on the Reykjanes Peninsula (Erlendsson and Einarsson, 1996). In the eroded Tertiary crust, although reverse-slip striations are seen on some fault planes, only four reverse faults deduced from offset marker horizons are reported from east and west Iceland. One is a fault caused by a gabbro intrusion (Friðleifsson, 1983) in east Iceland. Another is a low angle fault generated by a sheet injection (Torfason, 1979) also in east Iceland. The two remaining faults were observed in west Iceland, in association with magma injection and changes in the dip direction of secondary fracture planes (Khodayar, 1999).

Because compression induces shortening and closes fractures, it has implications for the kinematics of the plate boundaries and the behaviour of fractures in fractured geothermal reservoirs. However, little attention has been paid to the analysis and origin of reverse-slip structures in an extensional context. In this paper, we describe new observations of eight examples of reverse-slip structures found during our field investigations in two rift-jump blocks south of 65° in Iceland (Fig. 1a), and we analyse their kinematic origin. We evaluate the relative importance of fault geometry (dip value and dip direction), the role of intrusions and general tectonic compression. Our analysis shows that the origin of these reverse-slip structures is not due to a regional compressional stress field or shortening.

2. Geological context

The rift-jump blocks of Borgarfjörður and Hreppar (Fig. 1a) are two crustal blocks that were initially formed at the plate boundaries, but were shifted away and eroded down to 1.5 km in Borgarfjörður (Jóhannesson, 1975) and to 0.5–0.7 km in Hreppar (Sæmundsson, 1970). The Borgarfjörður block lodges between the extinct Snæfellsnes Rift Zone (SRZ), which was active from 15 to 5 Ma (Moorbath et al., 1968; Aronson and Sæmundsson, 1975; Jóhannesson, 1975, 1980; Kristjánsson and Jóhannesson, 1999), and the Reykjanes–Langjökull Rift Zone (RLRZ), which has been active from 7–6 Ma until present (McDougall et al., 1977; Jóhannesson, 1980; Kristjánsson and Jónsson, 1998). A possible zone of transform faulting or oblique rifting, as far as 64°40′N, is suggested during Tertiary time (Khodayar and Einarsson, 2002a). The Snæfellsnes Volcanic Zone (SVZ), a Quaternary intra-plate alkali chain (Sigurðsson, 1970; Jakobsson, 1972; Sæmundsson, 1978) cuts the Tertiary crust. This chain is interpreted by some authors as being the result of horizontal shear (Sigurðsson, 1970; Schäfer, 1972) or representing the remnant of a Tertiary transform fault (Jóhannesson, 1975; Sæmundsson, 1978; Jancin et al., 1985). The Hreppar rift-jump block is located between the receding RLRZ and the propagating Eastern Rift Zone. The South Iceland Seismic Zone connects these two rift segments, and is itself believed to be unstable (Einarsson, 1991). Rocks have an age from 3.1 to 0.7 m.y. in Hreppar (Aronson and Sæmundsson, 1975).

In these rift-jump blocks, the Tertiary lava pile consists of tholeiitic flows and sedimentary layers. Acidic rocks are confined to the central volcanoes, namely Hallarmúli, Laugardalur, Reykjadalur (Jóhannesson, 1975), and Hafnarfjall–Skarðsheiði (Franzson, 1978) in Borgarfjörður, and Stóra–Laxá and Þjórsárdalur in Hreppar (Friðleifsson et al., 1980). Hallarmúli and Stóra–Laxá are without apparent collapse caldera. Dykes intruded each of the two rift-jump blocks, but sills and cone-sheets are more frequent in Hreppar than in Borgarfjörður (Khodayar and Einarsson, 2002b).

Regional 'fold-like' structures, with axes parallel to the rift zones, exist in these rift-jump blocks, namely the Snæfellsnes syncline and the Borgarnes anticline in Borgarfjörður, and the Stóra–Laxá anticline in Hreppar. These structures are open with maximum dips of the order of few degrees and do not reflect compression by shortening. The syncline is the surface expression of the extinct SRZ and resulted from loading, extension and lava-tilt towards the active rift zones (Walker, 1963; Sæmundsson, 1967; Pálmason, 1973; Grellet, 1983). The anticlines were formed later, due to local re-tilt of the lava towards the new rift zones (Sæmundsson, 1967; Jóhannesson, 1975). The tectonic pattern of each rift-jump block is dominated by fractures striking N–S, NNE, WNW, ENE, E–W and NNW. These fractures acted alternatively as normal and strike-slip faults, but reverse-slip motion constitutes a small percentage of data sets on fractures (Khodayar, 1999; Khodayar and Einarsson, 2002b).

3. Description of the structures

We made several field campaigns to study large fault and dyke populations in the rift-jump blocks of Borgarfjörður in western Iceland and Hreppar in southern Iceland. We analyse here eight examples of reverse-slip motions that we observed during these campaigns. Two of the examples from Borgarfjörður were briefly analysed for the first time by Khodayar (1999), and further discussed here. The other six examples are new observations that are described for the first time in the literature. Examples 1–7 are from the

Fig. 1. Location of the studied areas, (a) in Iceland (map modified from Jóhannesson and Sæmundsson (1998)), (b) in Borgarfjörður (map modified from Jóhannesson (1994)), and (c) in Hreppar (map modified from Khodayar and Einarsson (2002b)). Circled numbers show the locations of the observed reverse-slip motions in each area.

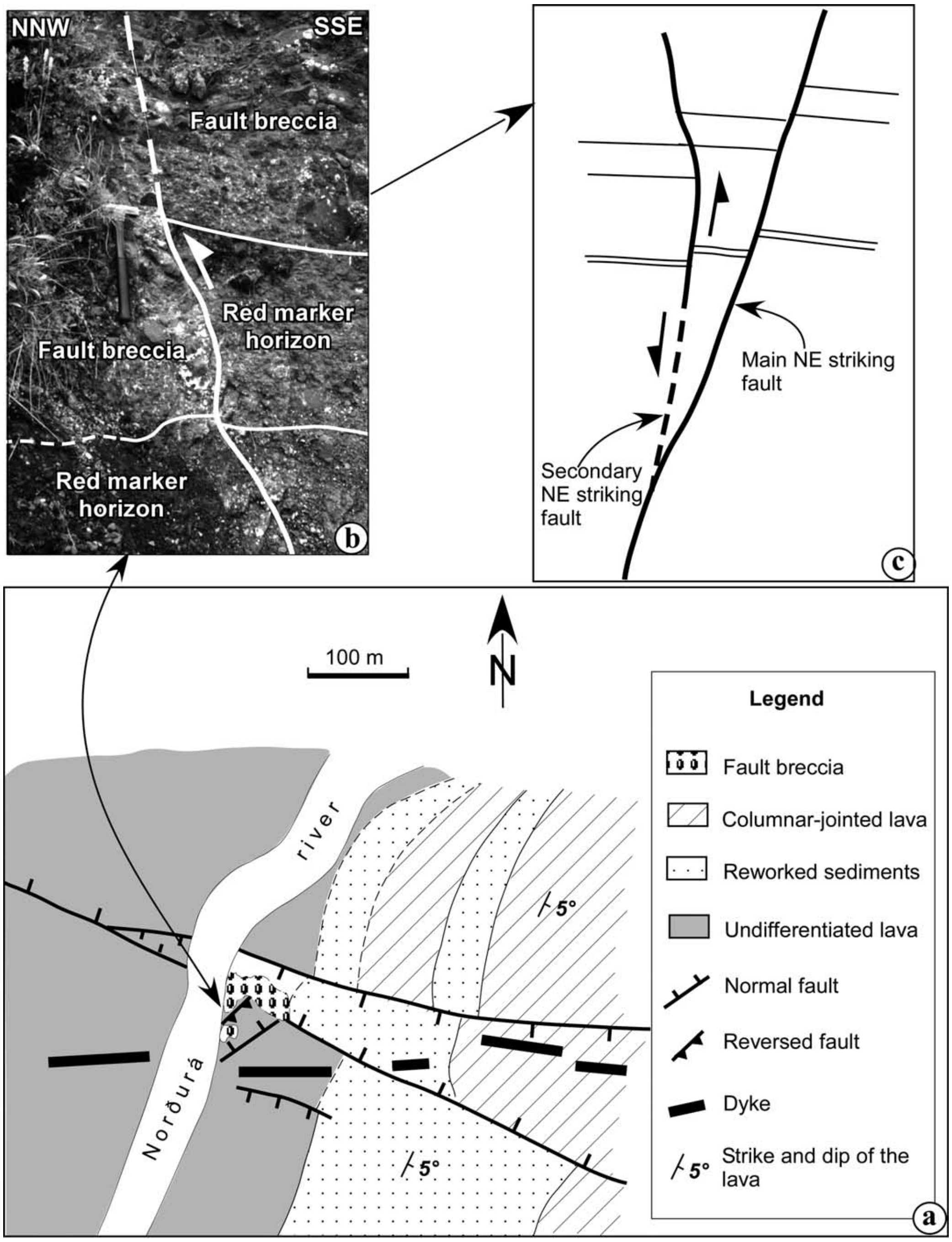

Fig. 2. (*Example 1*). Reversed fault in fault breccia. (a) Simplified geological map (Khodayar, 1996–2003, unpublished data) of the outcrop. (b) Oblique view of the detail of the reverse-slip displacement of the marker horizon in vertical section. (c) Schematic cross-section showing the displacement of the fault with respect to the fault-dips. The fault to the right is the NE striking normal fault, and the fault to the left is the reversed fault. Not to scale.

Borgarfjörður rift-jump block (Fig. 1b) and span both sides of the Borgarnes anticline, and some are in the Hallarmúli central volcano. Example 8 is from the Hreppar rift-jump block, north of the South Iceland Seismic Zone, and it is located southeast of the Stóra–Laxá central volcano (Fig. 1c), on the northeast flank of the Stóra–Laxá anticline. These structures were selected as examples because they represent the only known cases where both striae and marker horizons could be used to determine the sense of reverse-slip motion.

We use the terms 'sill' when the intrusion is nearly horizontal; 'cone-sheet' when the intrusion dips below 50°; 'dyke' when the intrusion dips above 50° (cone-sheets dip steeper if injected into pre-existing steeply-dipping fractures); and 'reversed fault' where inclined dykes show hanging wall uplift, or a local bend of normal fault shows reverse-slip motion.

Concerning the use of steps as a kinematic indicator on the fault planes, our interpretation of the slip-senses of the

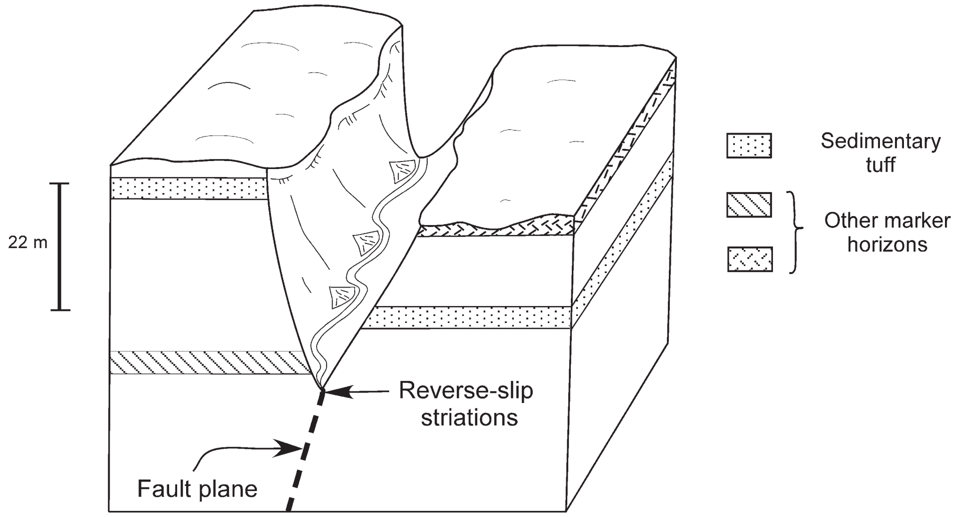

Fig. 3. (*Example 2*). Schematic cross-section of the outcrop showing the location of the reverse-slip striations along the steeply-dipping regional fault.

missing blocks agree with the observations of Vialon et al. (1991) and Doblas et al. (1997b) that steps face in the direction of the missing fault block. These later observations are from brittle but also ductile contexts and therefore analogous to the tectono-magmatic environment of Iceland. However, an opposite view on the use of steps as kinematic indicators has been expressed by Hobbs et al. (1976).

3.1. Reversed fault on secondary normal fault (Site 1, Fig. 1b)

We observed a NNE striking reverse fault in the fault breccia of a regional fault in the Hallarmúli central volcano (Fig. 2a). The regional fault strikes WNW, dips 80° towards the northeast, and has ≥50 m throw based on marker horizons. Whether the WNW fault also has a strike-slip component could not be deduced from the outcrop.

The WNW fault splits into two segments west of the river. The fault zone is 25 m thick east of the river, in the lower part of the WNW fault in the lava basement of the central volcano. The brecciated zone consists of metre-scale blocks of a pre-existing dyke, sub-angular millimetre to centimetre-scale basaltic fragments, and a red marker horizon. No clear evidence can be found on the origin of this thick fault breccia. However, it can be assumed that this fault breccia could either be due to the combination of normal and strike-slip motions along the southern WNW fault segment, or due to fault reactivation in time. Considering the location of the breccia between the two WNW faults, it could also be that this fault breccia is cumulative from both fault segments.

The reverse fault is near the outer edge of the breccia zone, on the footwall of the southern WNW regional normal fault, where it strikes parallel to another NE striking normal fault. Fault breccia of some 7 m thickness, excluding dyke elements, crops out between the NE striking normal and reverse faults. The reverse fault dips 84° towards the southeast in its upper part where it displaces the red marker horizon about 0.75 m (Fig. 2b), but the fault plane is not eroded enough in vertical section to check whether it contains striations. This fault plane changes its dip direction to the northwest in its lower part, and thus becomes a normal fault. Clearly, the normal motion of this fault is reversed in the upper part of the plane, and the reversed fault is due to a local change in the dip direction of the steeply-dipping normal fault plane.

The relation between the two NE fault planes could not be observed due to limited vertical exposure. However, it can be assumed that after changing its dip to the northwest, the ‘reversed fault’ merges with the NE-trending normal fault at greater depth (Fig. 2c), and behaves as an associated fault to this latter fault.

3.2. Reverse-slip striations on regional fault planes (Site 2, Fig. 1b)

Reverse striations were observed in a few cases on the main regional fault planes. In the Tertiary areas such as Borgarfjörður and Hreppar, the main fault planes are commonly poorly exposed and eroded, and when they are located in the axes of valleys, they are covered by rivers and alluvium. However, in places waterfalls have carved the bottom gullies and made a small portion of the fault planes accessible. We describe one such example (Fig. 3).

In the eroded Tertiary lava of the Borgarfjörður, a NW striking regional fault dips 73° toward the northeast, and showed reverse-slip striation (Fig. 3). The kinematic indicators to deduce the sense of the motion are steps on striated planes. The maximum thickness of the fault breccia is 2.5 m, and the fault plane is mostly eroded, especially in the upper part of the valley. The vertical offset between the sedimentary tuff across the valley is about 22 m. While striations on this fault are of reverse type, as discussed in Section 4.1, the vertical slip is likely to be normal.

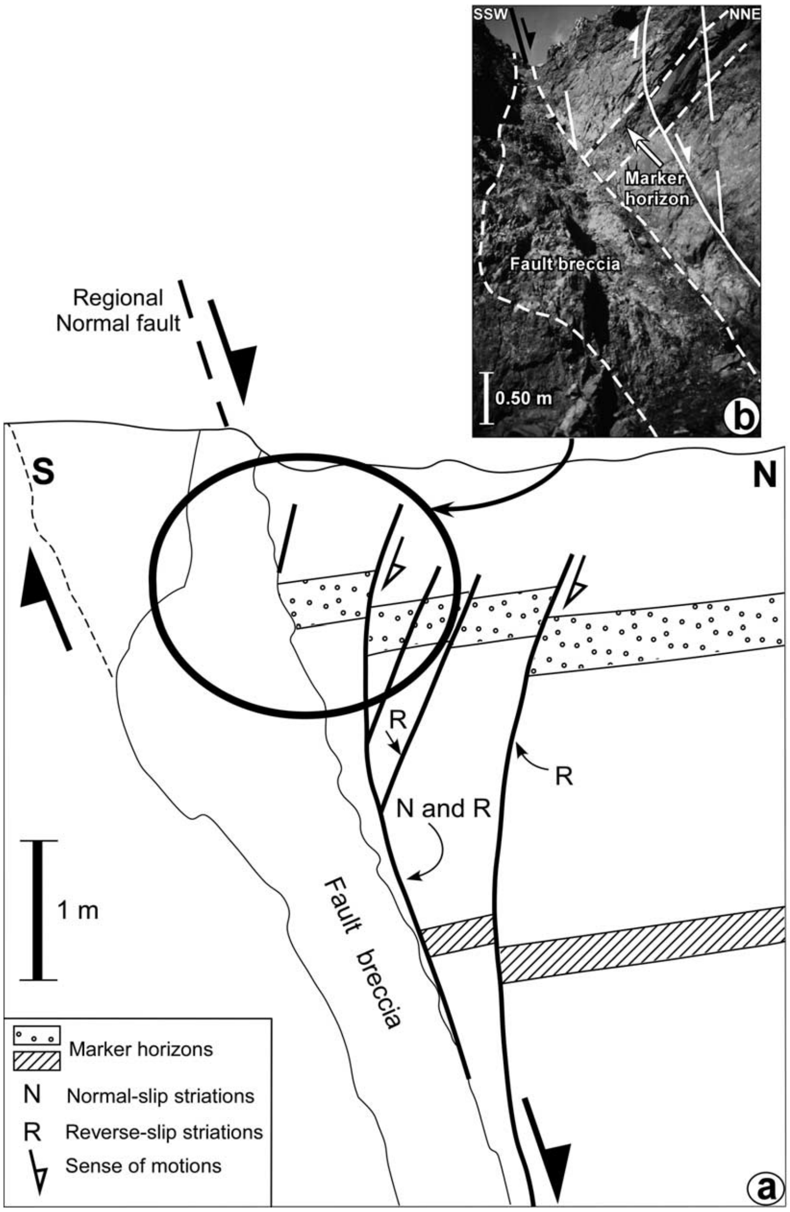

Fig. 4. (*Example 3*). Reverse-slip motion associated with Riedel fractures. (a) Cross-section of the geometry of the Riedel fractures with respect to the main fault. (b) Oblique view of the upper part of the Riedel fractures system.

3.3. Reverse slip on secondary fractures (Site 3, Fig. 1b)

Metre-scale secondary fractures associated with the major faults such as Riedel fractures are usually not well preserved in the Icelandic Tertiary rocks due to the erosion of the fault planes and the crushed fault zones. However, a pattern of Riedel fractures, showing normal and reverse-slip motions (65–88° pitch) was observed in association with a regional normal fault in the younger Tertiary lava in Borgarfjörður (Fig. 4a and b), as originally described by Khodayar (1999, Fig. 6a). The main regional normal fault strikes NW, dips steeply (72°NE), and has 12-m vertical displacement down to the northeast. A maximum of 1.3-m-thick fault breccia is observed in association with the fault.

At least five metre-scale Riedel fractures, with less than 1 m vertical displacement, are located on the hanging wall of the main normal fault. Four of these fractures strike between 22 and 32°, and one of them strikes N50°. Riedel fracture planes dip between 55 and 85°, but their dip direction changes in vertical section. They dip towards the east and the southeast in the lower part of the system, where they tend to merge with the main fault. There, the striae and the displacement of the marker horizons indicate normal-slip. In the upper part of the system, the Riedel fractures dip towards the west and the northwest, and are antithetic to the main fault. Most of the reverse-slip striae, identified based on steps on striated planes, are towards the upper part of the system, consistent with the reverse-slip displacement of marker horizons (Fig. 4b).

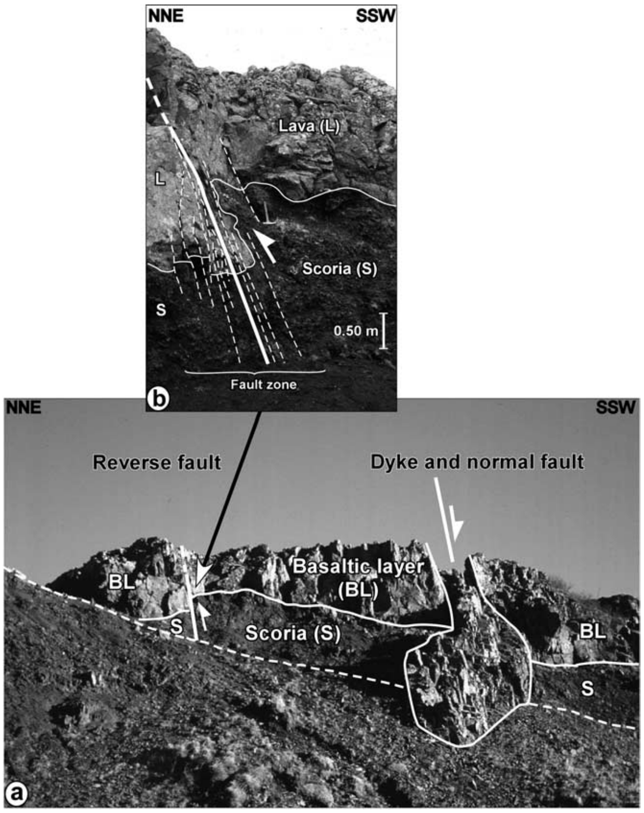

Fig. 5. (*Example 4*). Reverse fault parallel to a dyke. (a) View of the reverse fault, the dyke and the normal fault. The reverse fault and the dyke are 5 m apart. Outcrop height is 15 m from the basis of the dyke to the top of the lava. (b) Details of the reverse fault and the fault zone. Lava postdate the scoria, and the scoria is not infilling a topography.

3.4. Reverse faulting parallel to a dyke (Site 4, Fig. 1b)

A reverse fault parallel to a dyke was identified in the younger Tertiary basalt plateau, north of the Hallarmúli central volcano (Fig. 5a and b), as originally described by Khodayar (1999, Fig. 6b). These structures crop out 5 m from each other, in a lava containing numerous steeply-dipping joints. The fault and the dyke strike NW, and dip 76° and 72° towards the southwest, respectively. The thickness of the dyke is 2.20–2.50 m, and that of the fault zone is about 1.50 m consisting of regularly spaced southwestwards dipping fractures.

The contact of the lava and the scoria serves as a marker showing a reverse-slip displacement of 1.60 m along the fault (Fig. 5b). Centimetre-scale Riedel fractures in the fault zone also indicate reverse-slip motion consistent with the observed vertical displacement. On both sides of the dyke, however, the marker horizon presents a vertical displacement of 1.20 m down towards the southwest, implying existence of a normal fault at the site of emplacement of the dyke (Fig. 5a). Fault breccia and striae are lacking on the dyke margins, but a few flow lineations, with 80° pitch, indicate vertical magma flow in the dyke.

Flow lineations are dense, flat, and linear structures imprinted on the dyke's margins where they often occupy a large surface. Flow lineations range from horizontal to vertical indicating the latest direction of magma injection. Unlike tectonic striations, flow lineations lack kinematic indicators to determine a slip sense.

3.5. Reverse-slip indicators at the tip of a normal fault (Site 5, Fig. 1b)

A NNE striking normal fault (Fault A, Fig. 6) cuts a basaltic intrusive body in the Hallarmúli central volcano. The southern tip of this fault presents kinematic indicators of both normal and reverse types (Fig. 7a–e). The intrusive body consists mostly of a NNE striking sill, with 370-m-length, 15-m-thickness, and flat contacts. The sill ends horizontally and bluntly to the north in vertical section. However, to the south, at the intersection with a NW striking fault (Fault B, Fig. 6), the sill bends upwards, is only 7 m thick, and becomes a NW striking dyke dipping towards the northeast (Dyke A, Fig. 6). Columnar joints perpendicular to the sill and the dyke's edges follow the upward bend. The sill is cut by three faults striking NNE, N–S and ENE with 11–21 m vertical separations (respectively, faults A, C and D, Fig. 6) and by one ENE striking porphyritic dyke with 5.5–3.5 m thickness (Dyke B, Fig. 6).

To the north, the NNE fault (Fault A) dips 84° towards the northwest and displaces the sill 16 m down to the northwest (Fig. 7b). No striae were found along this portion of the fault. A few tens of metres farther south, the fault dips 75° towards the northwest in the lower tuffs. At this location, which may be at or very near the fault's tip, a portion of the hanging wall is connected to the footwall, and the contact of the sill and the lower tuffs is displaced only by 0.40 m along the NNE fault or Fault A (Fig. 7c). Despite this small vertical displacement, metre-scale slickensided plane crops out in the lower tuffs. The plane is shiny and polished. It contains regular dip-slip striae and kinematic indicators of both normal and reverse-slip types (Fig. 7d and e). Sharp steps indicate a reverse motion (Fig. 7d), while few detachments, drop-shaped structures (Doblas et al., 1997a), and striated elements show normal-slip (Fig. 7d and e), consistent with the displacements of marker horizons. The normal-slip indicators cut the striated surface and are younger than the reverse motion.

3.6. Reverse faulting at the tip of an intrusion (Site 6, Fig. 1b)

An outcrop of a sill, dykes, and reverse fault occurs within the ignimbrite of the Hallarmúli central volcano (Fig. 8). The local maximum thickness of the ignimbrite is 105 m, including a spatter layer at the base, a sill in the middle part, and an andesitic layer on top. The sequence is covered by alternating tholeiitic lavas and scoria, and dips 5–8° towards the southeast. Normal and reverse faults, dykes, and the sill cross-cut each other (Fig. 8a). The sill, interbedded in the ignimbrite, bends upwards laterally and turns into dykes at

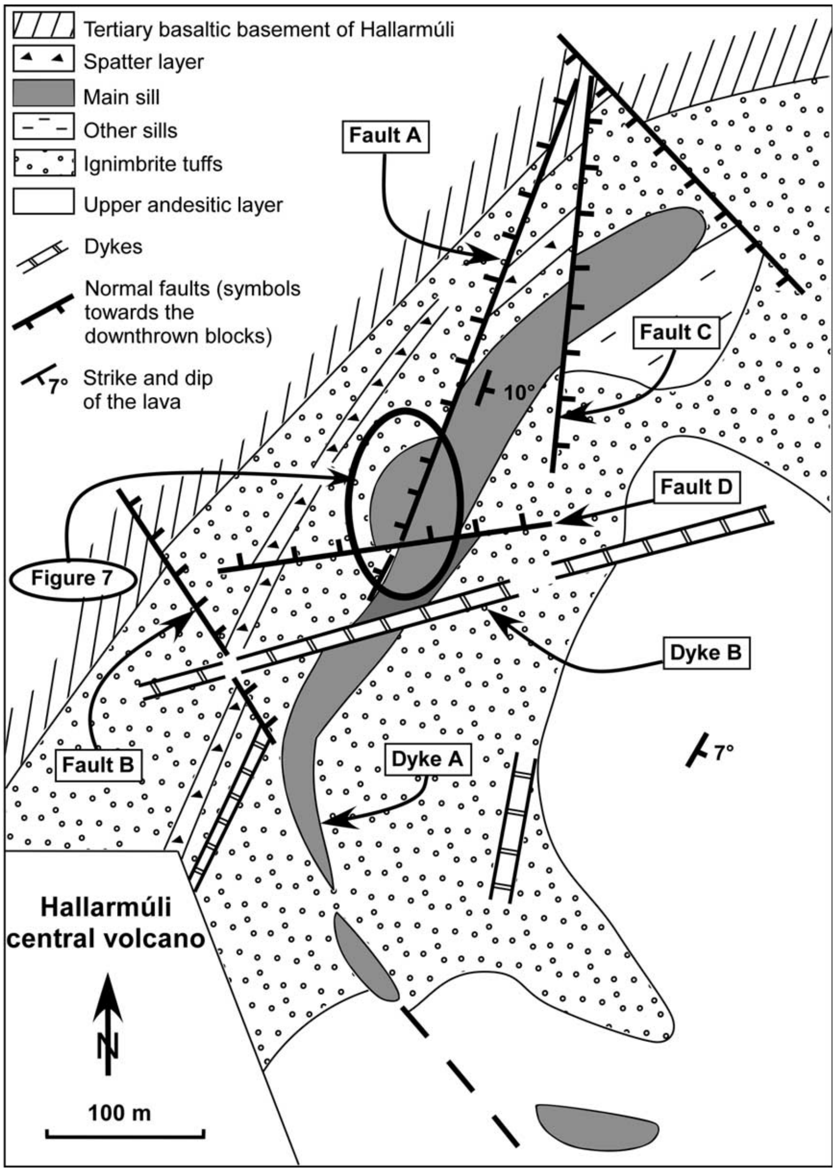

Fig. 6. (*Example 5*). Simplified geological map of the observed faults, dykes and a sill on the western flank of the Hallarmúli central volcano. The thickness of the dykes is exaggerated.

both tips (Fig. 8b) generating a semi-saucer shape. Columnar joints in the intrusive body turn from vertical in the sill to horizontal in the dykes of the Hallarmúli. The sill is 15 m thick over 280 m length, but 8.5 and 3.5 m thick in the northern and southern dykes, respectively (Fig. 8b). The northern dyke (3b in Fig. 8b) strikes NE and dips 65° towards the southeast. No shear displacement occurs across this dyke. To the south, the sill bends sharply upwards and becomes a N–S striking dyke (3c in Fig. 8b) dipping 85° towards the west in the ignimbrite and 65° towards the east in the upper andesitic layer (Fig. 9a). The andesitic layer is displaced vertically by 6.50 m across the southern dyke (3c), with the hanging wall uplifted (Fig. 9b). The dyke dips towards the uplifted block and thus resembles a reverse fault.

3.7. Reverse-slip striations along undulating dyke margins (Site 7, Fig. 1b)

Reverse-slip striations were observed in association with an ENE en échelon dyke system that branches from the N–S dyke described in Section 3.6 above (Figs. 8 and 9). These striations are different from flow lineations produced during magma injection (Khodayar and Einarsson, 2002a), and are formed on a hard and cold material, i.e. after the cooling of the dyke.

The en échelon dykes are left stepping, with two segments in the ignimbrite and one in the upper andesitic layer and the lapilli horizon. They are 50 cm thick with five injections separated by chilled margins in the ignrimbrite, but only 25 cm thick and with a single injection at the contact of the N–S dyke and in the andesitic layer and lapilli horizon (Fig. 9c–e). The dykes strike N80°, dip 83° towards

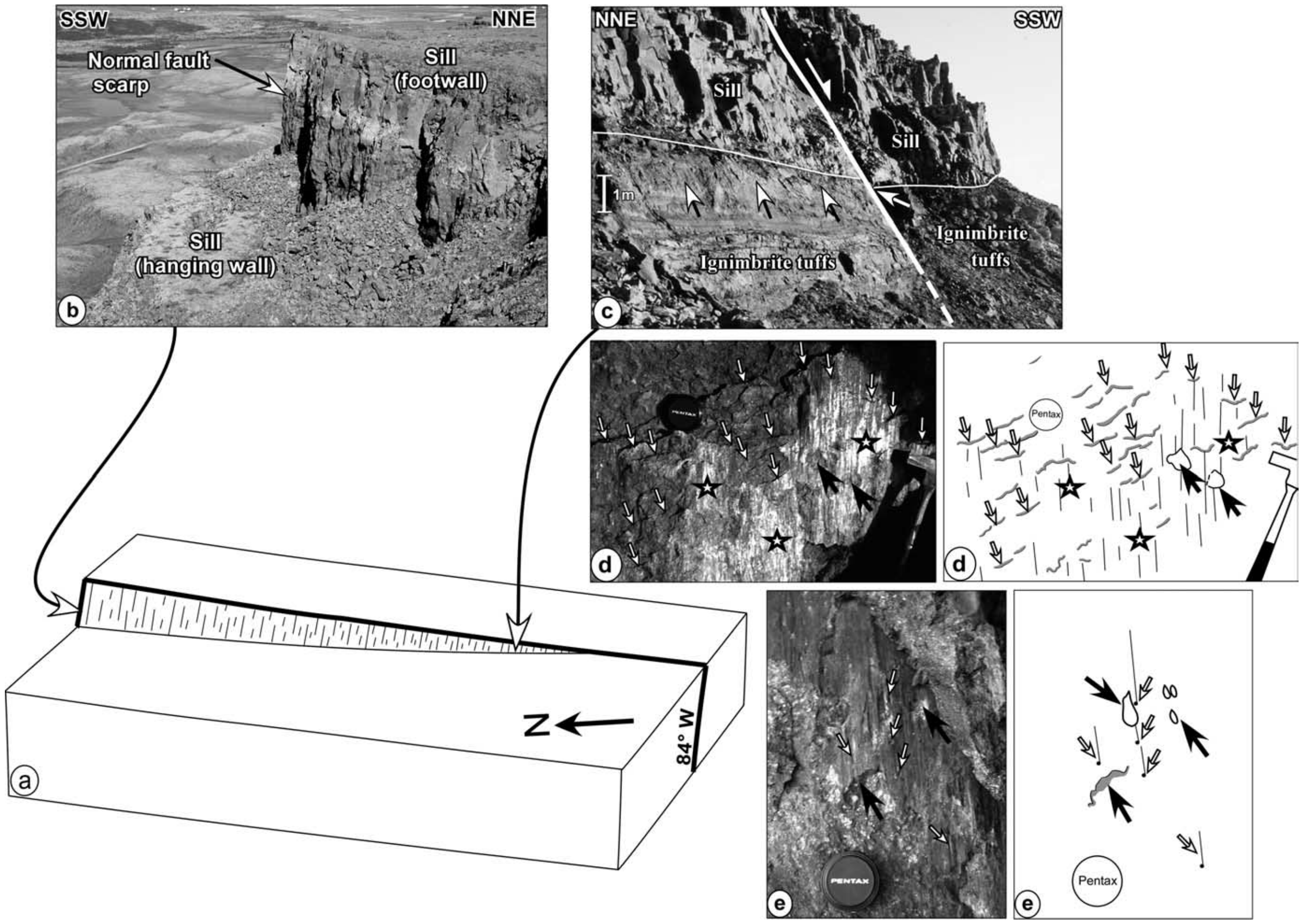

Fig. 7. (*Example 5*). Deformation along a steeply-dipping NNE normal fault. (a) Block diagram of the fault geometry and the location of the deformation. (b) The fault scarp in the sill, where the vertical displacement is 16 m. (c) The fault tip, where the vertical displacement is only 0.40 m. The face of the tuff in the left part of the photograph is also the fault plane. White arrows with black tails point to the slickensided plane in the lower tuffs. (d) Photograph and a simplified sketch showing the striations on the slickensided plane at the fault tip. The plane is polished and contains numerous regular sub-parallel striae (stars). The steps indicate a reverse-slip motion (white arrows), and the drop-shaped detachments on the striae show a late normal-slip motion (black arrows). (e) Photograph and a simplified sketch showing the detachments (black arrows) and the striated elements (white arrows) on the same slickensided plane indicating a normal-slip motion.

the northwest, but their margins undulate towards the northwest or the southeast (Fig. 9e).

Five slickensided planes were observed on, or very near, the margins of the thickest dyke segment. Dip-slip striae (80–58°) are located on planes that dip either towards the northwest or towards the northeast. Striae are very fine, numerous, and regularly spaced. Kinematic indicators (irregular steps) associated with one plane on the dyke's margin show reverse-slip motion. Despite the presence of dip-slip striae on the dykes' margins, no net vertical displacement seems to have occurred in the upper andesitic and lapilli marker horizons across the dyke (Fig. 9e).

3.8. Reverse-slip motion along a steeply-dipping cone-sheet (Site 8, Fig. 1c)

A reverse fault was observed in the eroded Matuyama rocks of the Hreppar rift-jump block in association with one of the cone-sheets of the Stóra–Laxá central volcano (Fig. 10a). The cone-sheet crops out in an olivine porphyritic brecciated lava that dips towards the northeast. The intrusion strikes NNE, dips 68° towards the northwest, is 22 cm thick, and ends in vertical section. Locally, it intrudes a pillow lava layer and displaces this marker horizon by 35 cm towards the southeast (Fig. 10b). The cone-sheet dips towards the uplifted block, and the displacement is clearly of reverse type. No striae were observed on the margins of the cone-sheet, however.

4. Interpretation

We suggest three kinematic origins for the observed reverse-slip structures: local bends of steeply-dipping fractures, magma intrusions, or a combination of both processes.

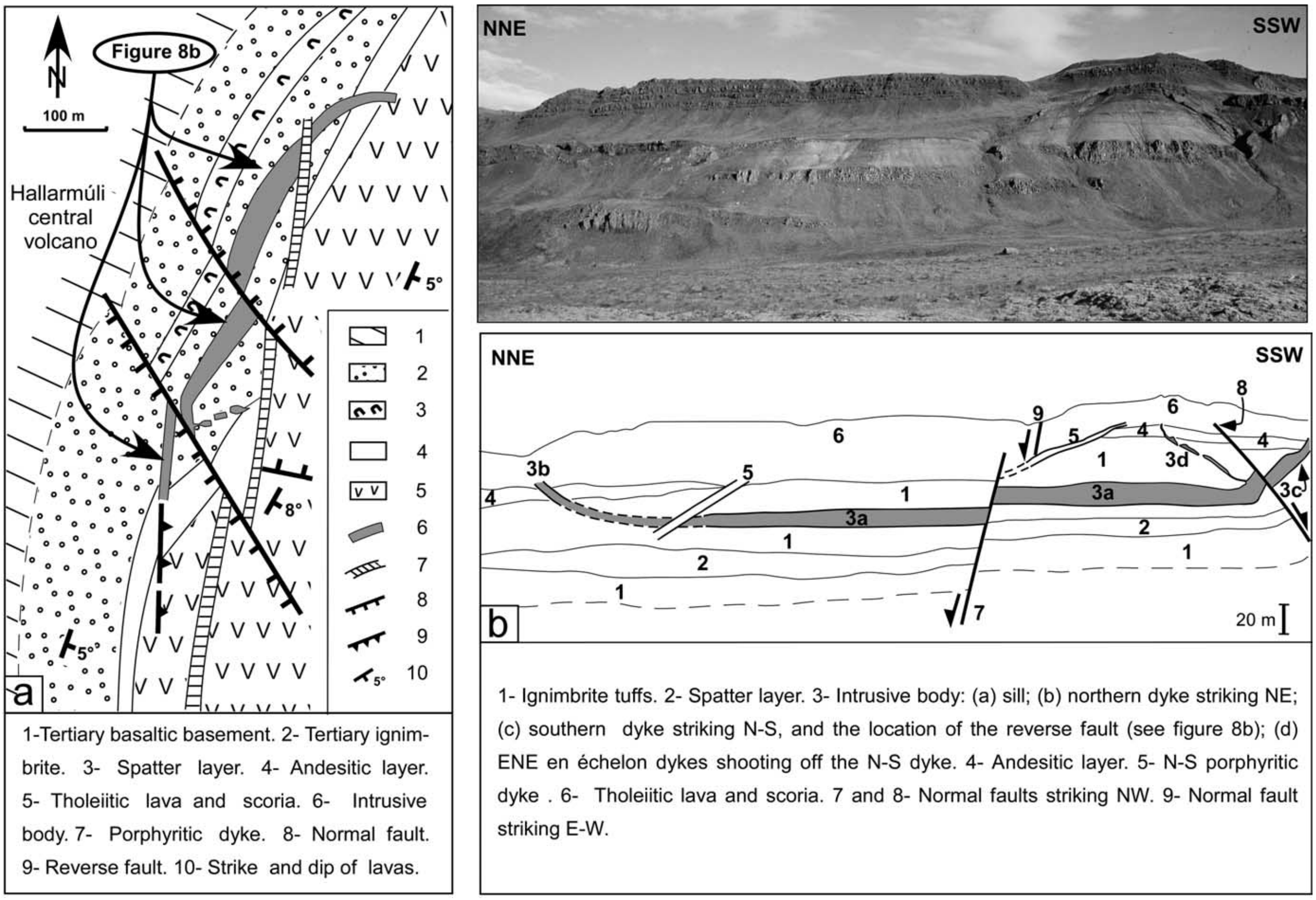

Fig. 8. (*Example 6*). Outcrop of a sill, dykes and the faults on the western flank of the Hallarmúli central volcano. (a) Simplified geological map of the outcrop. Dyke thicknesses are slightly exaggerated. (b) Cross-section of the outcrop. Outcrop height is about 150 m.

4.1. Reverse-slips due to local bends of steeply-dipping fractures: examples 1–3

The reverse-slip structures in examples 1–3 are either on secondary fractures, or on the main fault plane. In examples 1 and 3, the reverse-slip structures are observed in the damaged zones of normal faults whose type were deduced from dip-slip striae, and their magnitudes from marker horizons in the field. The preserved fracture geometry and marker horizons in the damaged fault zones suggest that these secondary fractures were normal faults that behaved as reversed faults in their upper parts due to the local bends of steeply-dipping fracture planes (Figs. 2a and 4a).

Fig. 3 shows the marker horizons and in this example 2 as well, the reverse-slip striations on the main fault planes are likely due to local bends and irregularities of the steeply-dipping normal fault. The reasons why this main NW striking fault is not interpreted as an initial reverse fault are threefold: (a) this fault plane is far away from the nearest central volcano where potential reverse faulting could occur due to magma inflation/deflation; (b) all faults surrounding this NW striking fault are extensional faults; and (c) no dyke was found around this outcrop to analyse the possible relation of the reverse-slip striations with an intrusive body.

Local bends of the steeply-dipping normal faults are by far the most plausible explanations for the three examples. Icelandic faults are unusually steeply dipping compared with faults acting in thicker and colder continental crusts. Generally, Icelandic faults dip above 70°, and in most cases above 80°. Very often they do present irregularities in vertical sections, at least in the upper 1.5 km in the crust (Khodayar, 1999). A few degrees change in the dip of these steeply-dipping fault planes is enough to induce the irregularities and local reversed faulting. Such irregularities are coherent with the general normal slip of the faults, without requiring a particular mechanism. Furthermore, our observation of a larger population of non-eroded irregular fault planes in Iceland indicate that irregular geometry of normal fault planes in the slip direction do not necessarily induce greater deformation in the host rock than fault planes with straight geometry.

4.2. Reverse-slips caused by intrusions: examples 4 and 5

We explain examples 4 and 5 as being caused by dyke intrusions. In example 4, synchronous reverse faulting, dyke injection, and the vertical displacement across the dyke cannot be explained in terms of regional stress field. A

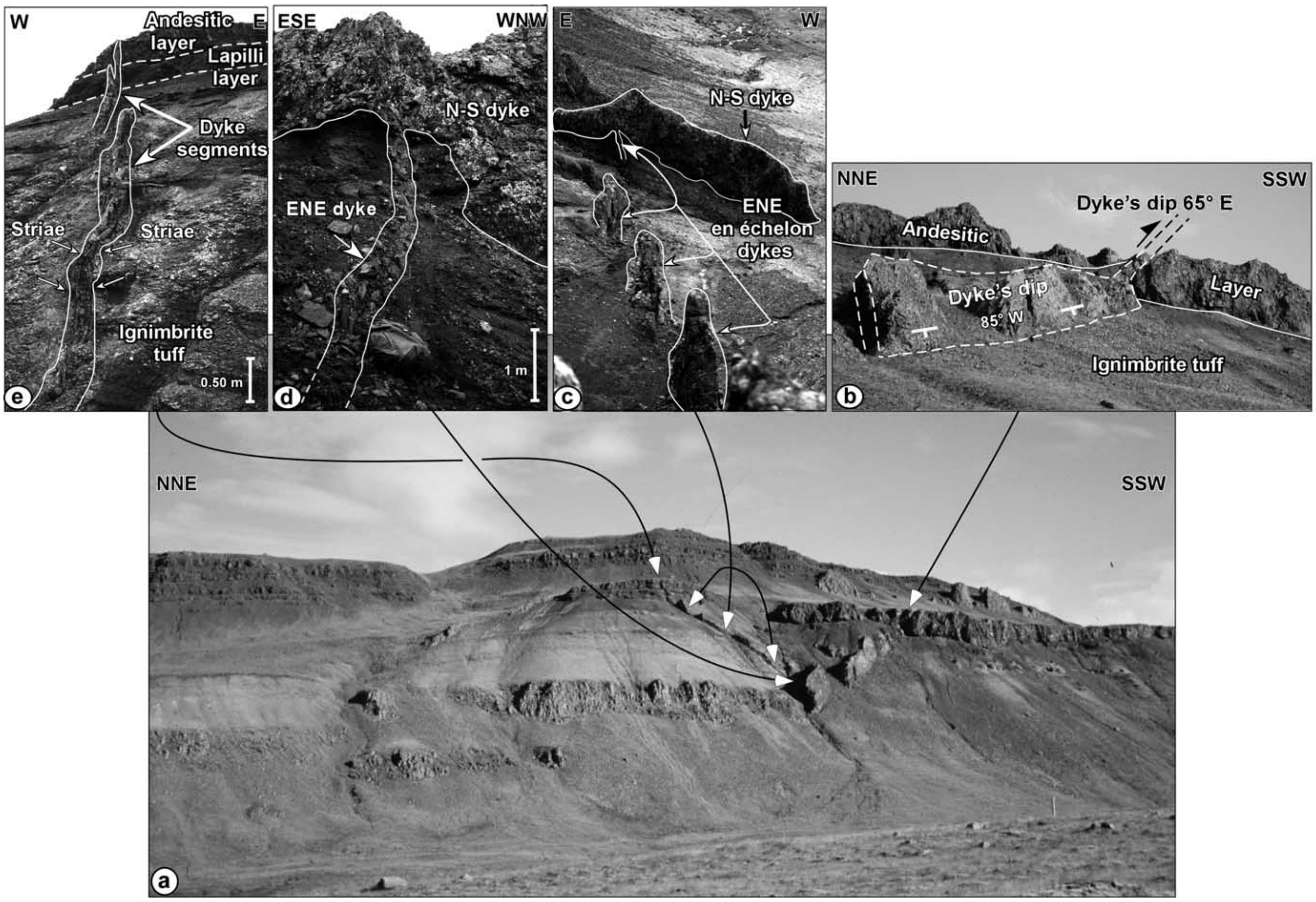

Fig. 9. View of the southern part of the intrusive body (same as Fig. 8). (*Example 6*) (a) To the right of the bend from the sill to the dyke, the N–S dyke is faulted and displaced towards the southwest. (b) View of the N–S dyke and the reverse fault. (*Example 7*) (c) The N–S dyke and the three ENE en échelon segments. (d) The en échelon system thins out and merges with the N–S dyke without a chilled margin, implying that the N–S dyke is the feeder of the en échelon system. (e) Undulating ENE dyke segments containing dip-slip striae.

minimum compressive stress (σ_3) oblique to the dyke's edges would explain the opening of the dyke, but only the fault motion along one of the fractures (Fig. 11a and b). The example can be explained kinematically, in terms of local tilt of pre-existing joints during dyke emplacement, as follows. Flow lineations on the dyke's edges indicate that magma vertically intruded one or more pre-existing joint(s) (Fig. 12a and b). An opening, oblique to the dyke's margins, induced a tilt of the joint to the northeast of the dyke, and the block between the joint and the dyke moved upwards, acting as the hanging wall of the reverse fault. The magnitude of reverse-slip is in the same range as the dyke thickness. Due to this oblique opening, the block to the southwest of the dyke moved downwards, producing a normal-slip motion (Fig. 12c).

Reverse fault plane solutions along the Mid-Atlantic Ridge (Einarsson, 1979), in Hawaii (e.g. Thurber and Gripp, 1988), or in the root of the intra-plate Eyjaföll volcano in south Iceland (Dahm and Brandsdóttir, 1997) are similarly interpreted as the result of forcible dyke intrusions. During the activation of such a process, the magnitude of ground deformation and reverse faulting is small compared with the rifting (Pollard et al., 1983; Khodayar, 1992). Our example 4, as well as a similar case from France (Khodayar, 1992) suggest that: (a) for a dyke 1–2 m thick, the area where reverse faulting occurs due to the joint-tilt lies within 10 m of the dyke; (b) when reverse faulting results from forceful dyke injection, the slip magnitude is roughly equal to the dyke thickness. However, more examples are needed to establish this relation as a general rule.

In example 5, the reverse faulting can be attributed to an underlying propagating dyke. This dyke may be one of the NNE dykes cropping out farther south (Fig. 6), though the depth, thickness, and the mode of injection cannot be deduced from the outcrop. Rubin and Pollard (1988) show that dyke-induced uplift is a common feature in both rift graben and the flanks of rift zones, and that slip occurs on faults that intersect the dyke near its top, that is in the zone of dyke-induced tensile stress. In our example 5, kinematic indicators at the tip of the fault indicate initial reverse faulting. This motion can be attributed to a slight uplift of the hanging wall induced by an underlying propagating dyke. The second generation of kinematic indicators and the fault displacement indicate a subsequent normal faulting, likely due to subsidence and extension after dyke emplacement.

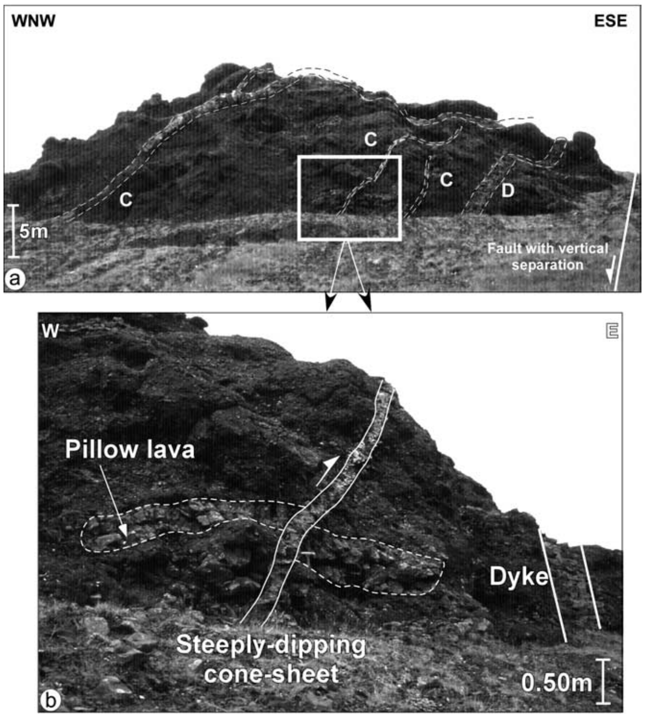

Fig. 10. (*Example 8*). View of a cone-sheet and a reverse fault ((a) and (b)). (C) Cone-sheet; (D) dyke.

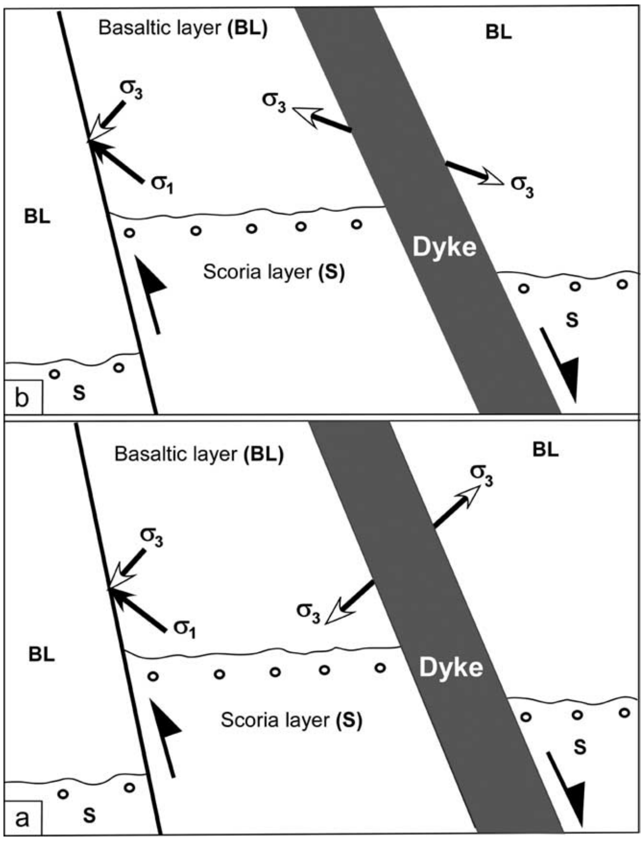

Fig. 11. Attempts to explain the reverse faulting parallel to the dyke and the normal faulting in terms of stress field (*Example 4*). (a) Situation with a minimum compressive stress (σ_3) which fits oblique opening of the dyke and the reverse faulting, but not normal faulting. (b) Situation with a σ_3 that fits oblique opening of the dyke and normal faulting, but not reverse faulting.

4.3. Reverse-slips caused by intrusions and bends of fracture planes: examples 6–8

The intrusive body of example 6 may have been initially a lenticular magma sheet that intruded a pre-faulted series (Fig. 13a–d). The northern tip of the intrusion intruded the host rock in the absence of a fault, hence a gentle bend from the sill to the NE striking dyke and a lack of vertical displacement across the dyke. By contrast, magma injected a pre-existing N–S fault to the south, hence the sharp bend from the sill to the N–S dyke (Fig. 13a–c).

Existence of a pre-existing N–S fault can be suggested from the transgressive bend of the sill to the dyke (c), as well as from the sudden interruption and disappearance of the spatter layer in the ignimbrite. Due to fracture cross-cuttings, it is difficult to confirm the initial geometry, length, and the type of the pre-existing N–S fault. However, N–S faults in Hallarmúli central volcano have normal-slip (Khodayar, unpublished data, 1996–2000), and generally dyke injections into pre-existing reverse faults are not observed in Iceland. Therefore, it can be suggested that the N–S fault at the emplacement of dyke (c) was initially a normal fault, which was intruded by the magmatic body and fractured later on.

The dyke (c) dips towards the west in the softer ignimbrite but towards the east in the andesitic layer. It is likely the reverse fault resulted from this local bend of the dyke plane as the dyke intruded the harder andesitic layer where an opening, oblique to the dyke's margins, uplifted the hanging wall (Fig. 13d). This is an example of a reversed fault by intrusion.

Example 7 shows the following. (1) The slickensided planes dip both towards the northwest and the southeast. If the motion is reverse on the southeast-dipping plane, it becomes normal when the dyke margin dips towards the northwest. In common with other examples, dip-slip motions can change from normal to reverse (and vice versa) as the dip directions of an undulating steeply-dipping plane change. (2) Despite the presence of dip-slip striae, the marker horizons do not show vertical displacement across the dyke (Fig. 9e). (3) With the lack of vertical displacement, we assume that the dip-slip striae resulted from the friction between the dyke's margins and the host rock caused by successive magmatic pulses within the dyke or by adjacent dyke injections. Hallarmúli contains a dense network of dykes. These striae may also be related to adjustments in the sill roof, but there is no direct evidence for this.

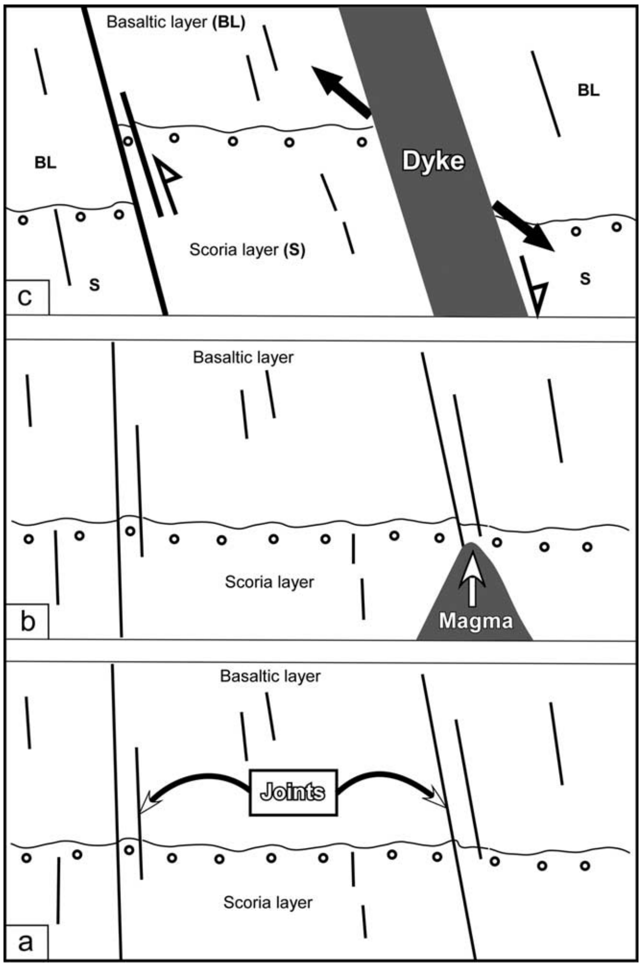

Fig. 12. Schematic cross-sections to explain the reverse faulting parallel to the dyke and the normal faulting kinematically (*Example 4*). (a) Lava contains steeply-dipping joints; (b) magma vertically intrudes one or more of these fractures; (c) the opening of the dyke induces a local tilt of the joint to the left, an uplift of the hanging wall of the reverse fault, and a normal-slip on the block to the right.

Finally, because the cone-sheet in example 8 has a dip direction orthogonal to the host rock, the opening of the intrusion induced an uplift to the hanging wall, causing reversed faulting.

5. Summary and concluding remarks

Because compression induces shortening in the rocks and closes the fractures, it has implications for the kinematics of the plate boundary and the fluid circulation along the fractures. This latter aspect is of importance in Iceland where permeability in geothermal reservoirs is mostly fracture-related, and the geothermal activity is influenced by stress changes during earthquakes (Björnsson et al., 2001). Therefore, results of tectonic studies from the older and deeply eroded crust are important contributions for the interpretation of active deformation that presently takes place at inaccessible depths. The detailed analysis of accessible fractures in the eroded outcrops is of great value to deduce whether or not fractures show rigorous evidence of regional compressional stress field or shortening.

If regional compression acts in Iceland (e.g. Haimsson and Rummel, 1982; Bergerat et al., 1990), its corresponding structures should commonly be observed. However, with the exception of striations on fault planes, compressional structures and reverse-faulting are reported rarely from the plate boundaries (e.g. Einarsson and Eiríksson, 1982; Erlendsson and Einarsson, 1996) and from the eroded Tertiary crust (e.g. Khodayar, 1999).

We analysed eight reverse-slip structures observed during our tectonic investigations of the Borgarfjörður and Hreppar rift-jump blocks in west and south Iceland, respectively. The observed structures constitute less than 3% of fractures we found in each rift-jump block. These structures were observed in Tertiary crust, in basaltic flows, in fault zones, and around two central volcanoes that lack apparent collapse caldera. They are characterised by reverse-slip striations and kinematic indicators on the fracture planes, and reverse faulting deduced from marker horizons across dykes, sills, and cone-sheets. Reverse-slip displacements are between 0.35 and 6.50 m. These structures cannot be easily explained in terms of regional or single stress fields.

Field evidence does not allow us to interpret these structures as a result of regional compressional stress fields. The results of our analyses, however, indicate that the observed structures have variable kinematic origins (Fig. 14a–h), and may occur due to:

1. Local bends and irregularities of steeply-dipping extensional fractures such as regional normal faults, secondary normal faults and Riedel fractures, dykes, or cone-sheets (Fig. 14a–c and g). The motions can change from normal to reverse as the dip directions of a steeply-dipping undulating plane change.
2. Local tilt of steeply-dipping joints due to forcible dyke injection (Fig. 14d). In this case, the fault and dyke have parallel strikes, and the magnitude of reverse faulting is roughly equal to the dyke thickness.
3. Uplift of the hanging wall of the steeply-dipping normal fault above an underlying propagating dyke (Fig. 14e).
4. Opening of a dyke or cone-sheet along planes that dip roughly orthogonal to the host-rocks, thus uplifting the hanging wall (Fig. 14f and h).
5. Friction between the dyke wall and the host rock during multiple magma injections (Fig. 14g).

In all intrusion-related cases discussed here, the lack of tectonic striae on the walls or the dyke margins support that the reversed faults are intrusion generated.

The rift-jump blocks are two crustal blocks that were formed at the plate boundaries, but were shifted away and

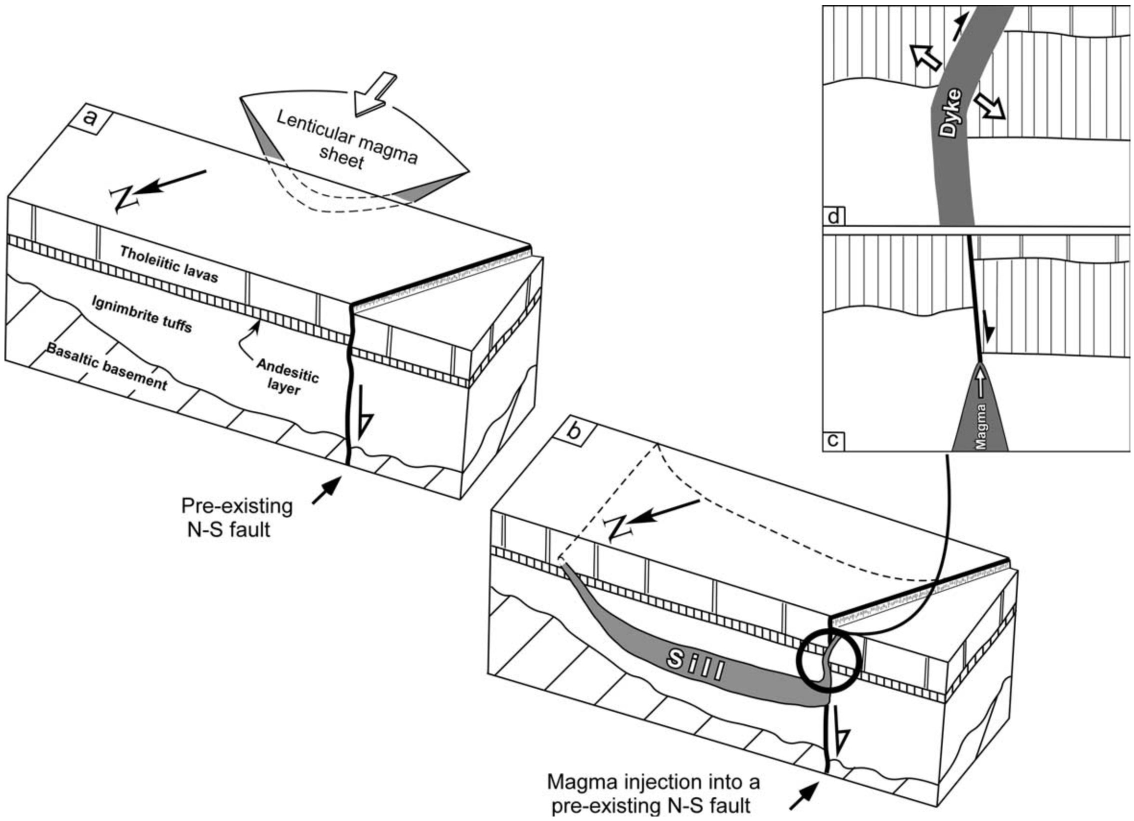

Fig. 13. Schematic figures suggesting the sequence of magma injection and reversed faulting (*Example 6*). (a) Intrusion of a lenticular magma sheet. (b) Formation of the sill, dykes, and reversed faulting along a pre-existing N–S normal fault. (c) Cross-section showing the assumption of magma intrusion into a pre-existing N–S normal fault. (d) Change in the dip direction of the dyke. The opening of the dyke uplifts the hanging wall and induces the reverse faulting.

eroded down to a maximum of 1.5 km into the crust. The geological context of these crustal blocks presents similarities with the presently active plate boundaries in Iceland (Khodayar and Einarsson, 2002a,b). Therefore, our new field observations and results from these two eroded blocks are relevant to the fracturing and the stress field in Iceland in general, and to the earthquake fractures and the fractured geothermal reservoirs at depth in particular. Our data show that:

1. Reverse-slip structures in the context of steeply-dipping fractures are not necessarily indicators of horizontal compression by shortening. We demonstrated cases of inclined intrusions where hanging wall uplifts potentially generate (a) horizontal extension, and (b) no net movement in the horizontal plane. However, cases of horizontal shortening across dykes were not demonstrated.
2. Dip-slip striations, even where conspicuous on metre-scale fault planes, do not always reflect vertical displacement.
3. If horizontal compression acts in Iceland, its magnitude is not sufficient to shorten the rocks and to close the fractures of various tectonic settings. Evidence of shortening by horizontal compression such as folding, common reverse faulting, horizontal shortening across intrusions, and coherence in the strikes of the observed reverse structures is not found.
4. Except possibly the intrusion in example 6, which could give small reverse-slip earthquakes in front of the dyke tip, none of our observed reverse-slip structures is a good candidate for a reverse-slip focal mechanism, as these structures were initially normal faults.
5. Reverse-slip motions may be only local adjustments to companion structures, and can lead to misinterpretation if not put into the correct context.

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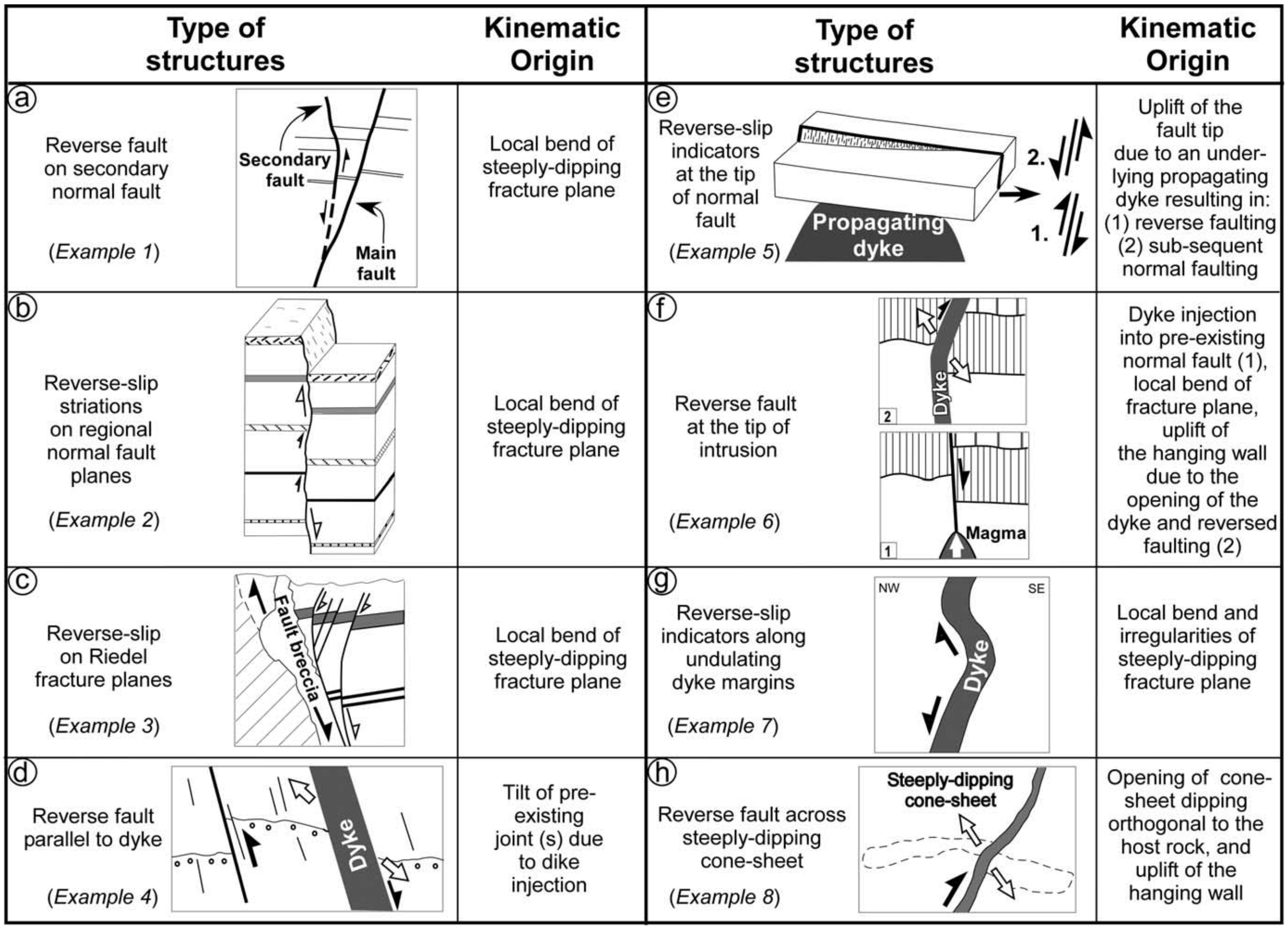

Fig. 14. Summary of the reverse-slip structures and their origin.

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