The internal structure of fault zones in basaltic sequences

Marc Holland a,⁎, Janos L. Urai a, Stephen Martel b

a Geologie/Endogene Dynamik – RWTH Aachen, Lochnerstr. 4-20, D-52056 Aachen, Germany
b Department of Geology and Geophysics – University of Hawaii, 2525 Correa Rd., Honolulu, HI 96822, United States

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Abstract

Normal faults in basalt along mid-ocean ridges and continental rifts play a major role in the formation of topography, advection of fluids, and the dynamics of biosphere. However, almost nothing is known of the internal structure of volcanic growth faults at depth, where many key processes take place. In this paper we present the results of scaled model experiments validated against outcrop studies of a volcanic growth fault system on Hawai‘i. The analogue model uses cohesive hemihydrate powder, with carefully characterized physical properties. A cohesion of 62 Pa, friction angle of 0.71 and tensile strength of 33 Pa imply a length scaling ratio of 1:5000–50,000 between the model and the upper part of extending oceanic crust with a basalt column above a normal fault. Time lapse imagery, high resolution particle image velocimetry and field observations show ground flexure in an initial elastic stage before subvertical mode I fissures propagate from the surface downwards. Some of the mode I fractures become inactive, while others develop into massively dilatant normal faults. The dominant processes within the faults are brecciation, block rotation and gravitational transport of fragments along the fault, maintaining meter-sized cavities to a depth of several hundred meters. Even when these volcanic growth faults, are resurfaced, a pervasive dilatancy persist within the system. If circulating seawater provides a sufficient oxygen supply, the large cavities in normal faults at mid-ocean ridges could harbour large ocean floor life forms to several hundred meters depth.

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1. Introduction

Basalts are common in the upper crust at large igneous provinces, continental rifts, volcanic passive margins, oceanic plateaus, ocean basins and large seamounts [1]. Unlike most sedimentary rocks that need burial for consolidation, basalt flows solidify at the surface, producing a strong but brittle rock. In tectonically extending regions where normal faults form in basalt, the scarpers typically are nearly vertical and are accompanied by open fractures [2–10]. The mechanics of near-surface faulting in basalts are quite well known [2–5,7–9,11–13] and the growth of a near-surface fault tip in basalts has been explored by several workers [2,8,11,13]. However, because of difficulties in access, almost nothing is known of the internal structure of these fault zones, even though this internal structure controls transport of intrusive, meteoric and hydrothermal fluids [5,6,11,15,16], the mechanical properties of a fault, and the potential to host submarine ecosystems [17–21]. In this paper we present a model for the evolution of the internal structure of normal faults in basaltic sequences based on scaled physical models calibrated against field observations of the Koa‘e fault system on Hawai‘i.

⁎ Corresponding author.
E-mail address: m.holland@ged.rwth-aachen.de (M. Holland).
Fig. 1. (A) Shaded DEM image of the central part of Kīlauea volcano with two rift zones merging at the caldera. The Koaʻe fault system is located S of the caldera and is interpreted to act as a structural link between the rift zones [10]. The N-facing scarps are visible as bright WSW/ENE trending streaks. (Illumination from 315°/45°; The upper insert shows the location of the DEM in respect to the Island of Hawai‘i, the white box shows the location of the SPOT image; UTM coordinates, WGS-84.) (B) Panchromatic SPOT satellite image shows recent lava flows as dark features. The 1974 flow ponds against the scarps of the Koaʻe fault system (arrows) at several locations before it bypasses the SW end of the fault system (UTM coordinates, WGS-84). Box shows location of panel C; figure locations indicated. (C) Detailed map of a fault (location shown in panel B). Dark grey areas are fractures of different widths; arrows show direction of opening vector in the horizontal plane while numbers indicate the vertical offset in centimeters. Note that opening magnitude prevails over vertical offset at this young stage of faulting (UTM coordinates, WGS-84).
Fig. 1 (continued).
2. Field study

2.1. Geological setting

We studied the Koa'e fault system located on Hawai‘i [8,10,11]. The youngest sub-aerial volcanic edifice on Hawai‘i is the shield volcano Kīlauea (Fig. 1A) with recent effusive activity at the Pu‘u ‘Ō‘ō vent. The major structures related to the Kīlauea edifice are the summit caldera, with Halema‘uma‘u crater and two rift zones that merge at the caldera [10]. The two rift zones separate Kīlauea into a relatively static block adjacent to Mauna Loa and a mobile southern flank [22]. While the trace of the Southwest Rift Zone is rather linear, the East Rift Zone shows a prominent curvature. The structural link between the rifts is the Koa‘e fault system [10].

2.2. The Koa‘e fault system

The Koa‘e fault system (Fig. 1B, C) contains normal faults with scarps that face predominantly to the north and that offset the ground by as much as 25m. The fault system is active [23]; it is approximately 12 km long and 3 km wide and narrows towards the SW. Many prominent fissures are associated with the faults. The last major volcanic resurfacing in this area event took place approximately 400–750 yrs ago [10,24]. The maximum total throw on single faults is larger than the morphological offset [8,10,12].

2.3. Structures along the trace of the Koa‘e faults

The normal faults are associated with mode I fissures predominantly on the footwall block [11]. Both the fissures and the normal faults that incorporate them show local horizontal openings of several meters (Fig. 1C) [8,10–12]. Fractographic observations show that opening precedes slip. A morphological transect (Fig. 2) through a typical large fault scarp shows [10]:

1. A footwall where fissures striking subparallel to the scarp show prominent openings in a zone up to 100 m wide next to the fault (Fig. 2). Field observations reveal that the fissures are open at least to several meters depth, where accumulations of detached wall fragments prevent the true depth from being observed.

2. Subvertical fault scarps. Their rough surface formed during an initial opening phase prior to subsequent slip (Fig. 2).

3. Slab-like ramp structures connecting footwall and hanging wall (Fig. 2, see also Fig. 9) which can have a convex, linear or concave cross section and are heavily fractured at their termination on both the hanging wall and footwall sides. In some cases an antithetically dipping cavity is found at the lower termination [8]. Duffield [10] and Martel and Langley [8,11] explain the ramp as a remnant of a monocline formed by early surface flexure, with the cavity as a result of failure at the crack tip of an upward propagating subvertical fault [8,11]. (Note that this ramp is not a relay ramp described in many structural geology studies (e.g., [12]).)

4. The hanging wall, with a surface generally having a much lower density of fissures as compared to the footwall (Fig. 2).

2.4. Volcanic growth faults

The understanding of the internal structure and morphology of the Koa‘e faults involves the concept of a volcanic growth fault [25]. As recognized by Langley [8] and Peacock and Parfitt [12] the Koa‘e fault system is periodically covered by lava flows. The most recent example is the Ka‘ū Desert 1974 lava flow that ponds against the fault scarps at several locations before it bypasses the fault system at its SW termination (Fig. 1B). Field observations in the area show that some lava flows are able to pass over pre-existing fissures without filling them (Fig. 3). Due to erratic emplacement of lava flows, the thickness of the unfaulted cover may vary laterally and vertically.

3. Scaled physical models of a normal fault system in basalt

In a properly scaled model of near-surface structures in basalt, the key requirement is the proper understanding and choice of the tensile strength of the model material and its prototype. The formation of vertical walls and scarps requires a cohesion and tensile strength. Specifying the (Mohr–Coulomb) cohesion and friction angle alone [26] is not sufficient to demonstrate correct scaling. Sand (Fig. 4) [27–29], or glass beads which are commonly used are not suited for our purposes, since these materials have neither true cohesion nor tensile strength – although we note that “apparent” values for these parameters are commonly obtained by extrapolating the failure envelope to low normal stresses [29]. Others use wet plaster, flour, starch, damp sand or wet clay [30–34] with a finite but unknown tensile strength and therefore an incomplete material characterization. In
Fig. 3. A section of pāhoehoe of the 1971 Kaʻū Desert surface flow passes two fissures only filling them to a few tens of centimeters. The differential cooling of the lava (against rock vs. air) leads to a rapid cooling of the lava at fissures such as this. We expect therefore that open fractures are not necessarily plugged to large depths by younger overlying surface flows (location indicated in Fig. 1B).

Fig. 2. Photographs of the fault scarp morphology at central parts of Koaʻe faults. Sub-vertical fault scarps, ramp structures, debris accumulations and fissures on the footwall are common throughout the system (figure locations indicated in Fig. 1B).
our studies we used dry hemihydrate powder with a mean density of $\rho_P = 860 \text{kgm}^{-3}$. Its mechanical properties were measured in shear tests at low normal stresses as well as by measurement of the tensile strength [35,36] with a tensile strength tester developed by Schweiger and Zimmermann [36]. Results show a curved failure envelope (Fig. 4) with a cohesion of $C_P = 62 \pm 5 \text{Pa}$ and a tensile strength of $S_{P,T} = 33 \pm 4 \text{Pa}$.

Shear modulus of the powder is of the order of $E_P = 50,000–80,000 \text{Pa}$.

Typical rock parameters for the basalt prototype are a density around $\rho_B = 2800 \text{kgm}^{-3}$, a friction coefficient of $\Phi_B = 0.66$, and a cohesion in the range of $C_B = 3.5–32 \text{MPa}$ depending on sample size and degree of fracturing [37–39]. Unfortunately the bulk shear modulus of a fractured basalt mass is not well known [38], and the
scaling of elastic deformation in our models is not well constrained. Scaling relationships for inelastic deformation [26] require that the failure envelopes are of a similar shape. To derive the scaling ratio for length, we fit a linear failure envelope to the hemihydrate data and obtain an apparent cohesion of approximately $C_{p_{-Ex}}=200 \text{Pa}$ and a friction coefficient of $\Phi_{B_{-Ex}}=0.71$. The scaling ratio for distances (neglecting time-dependent forces [26]), is 1:5000 to 1:50,000. 1cm in our model corresponds to 50–500m of basalt, depending on the strength of the rock prototype.

The model setup for producing a symmetric graben system has a rigid base with faults that dip at 60°. The semi-2D model is sandwiched between low friction glass panels for observation in profile (Fig. 5). The model setup simulates an unfaulted cover of basalt above a pair of pre-existing blind faults, as an analogue for volcanic growth faults [25]. To enable optical analysis with PIV [40], a small fraction of blue grains ($\ll 5\text{ vol.\%}$) was added to the powder. To simulate different thicknesses of lava cover, powder columns from 7cm to 15cm were used (Table 1). The documentation of the experiment is done with top-view and side-view digital cameras delivering a high-resolution time-lapse image sequence. A list of the experiments performed in this study is given in Table 1.

All the experiments show similar structural evolution, and experiments done under the same conditions are reproducible. Detailed observations of the evolution of fractures in 3D (top and sides of the model) show that the processes seen in profile are not significantly influenced by interaction of the powder and the glass surface. The common structural elements are described below, while a description of the differences between experiments with a different thickness is presented in Section 3.2.

### 3.1. Common structural elements

A PIV analysis [40] of experiment 030804b (Fig. 6) shows how initially continuous deformation is localized within the first 1% of bulk extension. The initial displacement field is similar to the one calculated for an elastic model [8,11,41] and we interpret this initial deformation to be dominantly elastic. This elastic deformation produces a monoclinic flexure of the surface (cf. [8], note the displacement vectors in Fig. 6). The first permanent deformation features are (i)

### Table 1

Overview of the experiments

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<td>Optical resolution</td>
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<td>No. of cameras</td>
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vertical mode I fissures at the surface, nucleating close to
the projection of the basement fault to the surface and
propagating downwards, and (ii) shear fracture nucleat-
ing at the tip of the basement fault and propagating
upwards. With further deformation these two fractures
grow towards each other, producing a curving zone of
localized deformation.

Fig. 7 shows ten stages of experiment 111403. After
the first vertical mode I fissures form at the
surface (Fig. 7B, K) a buried antithetic fracture is
formed. New subvertical fissures open progressively
closer to the centre of the graben with time, but they
have smaller apertures and extend to shallower depth
than the first fissure. With progressive throw on the
basement fault, the fissures propagate further down-
wards with a decrease in dip and connect to the
nondilatant fault zone which in turn connects to the
basement fault. Some of the surface fractures start to
slip (Fig. 7C, K). In most models the initial fissures
in the footwall block become inactive but remain
open (Fig. 7C; left side). With further deformation a
large block (or slab) separates from the substrate
along the shallow antithetic fracture. As this block
then rotates, it forms a ramp, and a cavity opens
beneath it (Fig. 7D–F). The deformation underneath
the antithetic cavity is obscured by the ramp (Figs.
7G and 8).

Fig. 8 shows that the rotation of the ramp increases
the compression at its base. This leads to the buckling
and break-up of the lower termination of the ramp. The
ramp rotation leads to large openings along the top of
the scarp at the upper termination of the ramp. During
the faulting process unstable fragments of the wall fall
into the open fault system and are reworked with further
fault slip (Fig. 7I, L). In all experiments three zones (Fig.
7F) can be distinguished showing (1) open mode or
mode I fracturing at shallow depth, (2) mixed mode at
intermediate levels and (3) mode II movement at depth
[4]. This can be correlated with the curved shape of the
failure envelope (Fig. 7M).
Fig. 7. (A–J) Image sequence of a hemihydrate powder experiment 111403. Prominent structural elements are indicated. The curved failure envelope (M) of the powder divides the experiment in: (1) a shallow mode I section, (2) a deeper mixed mode section and (3) a deep section with mode II deformation (Frame numbers in brackets). (K) Detailed views the same experiment showing rubble accumulation and (L) change of deformation modes. Arrows indicate crack tip positions. The approximate location of K and L is shown in panel E; frame numbers are given.
3.2. Differences caused by different basalt thickness

The structural evolution in the experiments with different thicknesses of the powder is different in two aspects. Firstly, in experiments with an extremely thin layer of powder, the mode II fractures at depth are absent. We interpret this to be a result of the stress state that is only of type 1, shown in Fig. 7M. Secondly, a thin layer of powder produces a less localized zone of fracturing and more complex fracture pattern including arching effects. Due to the small stresses mechanical heterogeneities become more significant, and the thin layer causes the initial vertical displacement gradient to be higher.

4. Discussion

In this study we focused on normal faults in basalt which are not conduits for upward flow of magma. Scaled physical models of normal faults (Fig. 7) in basalt can be constructed using cohesive powders [33,34,42]. Accurate characterization of tensile strength and low-stress shear behaviour is necessary for proper scaling (Fig. 4). The close correspondence of the near surface features of our models and the structures observed of the field gives us confidence in the utility of our model and in its predictive ability. The basic morphological elements observed in the field, like fissures, dilatant faults, ramp structures and buckles, are all reproduced by the analogue models (Fig. 10). One parameter for which the scaling is not known is the elastic modulus. This is a common problem in scaled physical models. A consequence of this is that the magnitude of elastic deformations in the field and in our experiments may not scale correctly [11]. Nonetheless, the good correspondence of the non-elastic deformation we observe at the surface in the model with that in the field suggests to us that our model reflects the non-elastic deformation in the deeper parts of the fault zone as well.

Normal faults in active volcanic regions can be periodically covered with lava [8,12], resulting in volcanic growth faults (Fig. 9) [25]. In resurfacing events the lava cover does not necessarily plug surficial fractures to a large depth. Some surface flows are able support themselves while passing over fractures with Fig. 7 (continued).
more than several tens of centimeters opening (Fig. 3). Depending on the fissure width, the flow temperature, viscosity and emplacement velocity, the front of a surface flow is able to support its own weight so it does not flow into the open fracture. Perhaps the air in the fissure promotes sufficient cooling of the propagating flow front to develop a tough skin at the base of the flow that allows the flow to bridge the fissure (Fig. 3). The new unfaulted basalt can be expected to be stiffer than the immediately underlying faulted basalt. Weathering products, like the broken glass surface, aeolian sediments or gas-rich zones within the lava units (e.g. shelly pāhoehoe) produce an additional mechanical stratigraphy.

Based on the analogue model results we infer that the early stages of normal faulting of a basalt unit on top of a buried fault involves prominent subvertical mode I fissures propagating downwards (Fig. 7, cf. [11]). Some footwall fissures become deactivated, while others become incorporated into the fault at deeper levels. The listric character of the system is furthermore responsible for the large openings at the surface of the model.

We interpret the ramp structures in the field to be the result of a shallow antithetic fracture after a stage of a monoclinal surface flexure (Figs. 7, 8 and 10; [11]). This fracture produces a surface ramp on top of a heavily fractured zone with large openings (Fig. 7G). The surface termination of the antithetic fracture is a zone of buckling (Figs. 2, 8 and 10). Field observations show that cavities beneath or adjacent to the ramp can have openings of more than a meter, with the distance between features associated with pronounced buckling and dilatancy being only a few meters.

Depending on the thickness of a basalt cover formed during intratectonic resurfacing events, further faulting leads to different morphological profiles as well as internal fault structures (Fig. 9). Immediately above the fault, a thin cover of basalt will disintegrate along closely-spaced pre-existing weakness planes (bedding, cooling cracks, intercalated weathering products or sediments) (Fig. 9B). This deformation leads to arching effects due to the low overburden stresses whereas a thicker cover of basalt shows more localized faulting (Fig. 9C).
The brittleness of the rocks combined with localized tensile stresses that accompany the faulting lead to an abundance of open fractures and cavities not only at the surface but also at deeper levels. Since hydraulic transmissivity depends on the cube of the channel aperture [6] the flow rates in cavities with apertures of a meter can be millions of times greater than for a cavity with an aperture of a millimeter; the connected voids in a normal fault system can be expected to have a profound effect on the hydraulic behaviour of a fault. In mid-ocean ridges or other volcanically active areas hydrothermal flux will be focused on these fracture systems together with the associated ore forming processes. In addition we note that the habitat of large life forms around black smoker systems at mid-ocean ridges [17,18,20,21] could extend into the meter-scale cavities and open fractures of the normal faults in the subsurface. The apertures of these fractures allow large fluid fluxes and fluid mixing, enabling the transport of oxygen necessary for large life forms (Wirsen, personal communication). The extension of microbial, anoxic life forms in the subsurface below the mid-ocean ridges has already been discussed [21]. We propose that the habitat of large life forms could as well extend several hundred meters below the ocean floor.

5. Conclusions

Normal fault systems within basaltic sequences are common features. Large voids in these normal fault zones can remain open to several hundreds meters of depth [4]. This has an important impact on hydrothermal systems associated with tectonically and volcanically active areas. The observations of the field work on the Koa’e fault system and the insights from the analogue model might not only apply to outcrops on Iceland [4,7,30,43,44] or other rift systems [2,3,9,15], but also to normal fault systems found at mid-ocean ridges [45] or other submarine systems [17,46–50]. The pervasive opening of fractures of normal fault systems in basaltic sequences affects mineral deposition [51] in associated hydrothermal systems (e.g., black smokers). The open fractures may form a previously unknown habitat for micro- and macroscopic life forms that are associated
Fig. 10. Diagram summarizing the features seen in the field and in the experiments. The system is heavily influenced by dilatant fractures on and beneath the surface. Open deactivated fissures exist on the footwall, fault scarps with large fissures terminate at prominent ramp structures, and buried cavities. All might host large live forms in the subsurface (not to scale).

with such systems [20,21], depending on the supply of nutrients and oxygen (Fig. 10).

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