Deep earthquakes (35 – 45 km) in the East African Rift Zone: Frictional properties from shear experiments on rock samples from the Lake Malawi area

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Abstract

Earthquakes in the southern part of the East African Rift occur at depths in excess of 30 km, up to 35 - 45 km, indicating a thick seismogenic crust. This is uncommon, especially in an extensional regime. Typically, earthquakes don't nucleate at depths corresponding to a temperature above 350°C, the temperature at which quartz becomes ductile. In an extensional regime, the geotherm is usually elevated, and earthquake depth more shallow. Several hypotheses have been proposed to explain why this is not the case in the (southern part of the) East African Rift Zone. Mafic material has a high viscous strength and it is therefore expected to deform brittle to deeper depths than quartzofeldspathic material. The presence of weak zones would locally increase the strain rate within a strong lower crust, lowering the brittle-ductile transition. Here we provide clues to the potential of seismic behavior at depth for three different rocks sampled from the Malawi rift zone. Rotary shear experiments are conducted at 100 MPa, 200 MPa and 250 MPa effective normal stress, at temperatures from room temperature up to 600°C. These experiments can say something about the strength of the rocks and the velocity dependence of frictional strength, expressed using the rate and state friction parameter (a-b). This gives clues to seismic behavior from shallow to deeper depths, and possible deformation processes playing a role in this. A calc-silicate sample containing mostly diopside shows a relatively decreasing trend in friction with temperature, with its highest friction at 200°C of 0.82, and 0.57 at 600°C. It shows no trends in (a-b). Its microstructures show areas of smaller grain size compared to the surrounding grains. Grain size reduction thus played a major role in this sample. The felsic sample (58% plagioclase) shows a possible dependence of friction on normal stress, indicating brittle behavior up to high temperatures. The friction coefficient of the mafic sample is relatively unaffected by changes in temperature and normal stress. It has negative (and decreasing) (a-b) values at high temperatures (> 400°C). The strength of the mafic material and its ability to localize shear to planes of weakness at depth, in combination with its velocity weakening properties at high temperatures may provide part of the answer as to why these deep earthquakes take place in an extensional setting and in which rock type the earthquakes nucleate.

1. Introduction

Brace & Kohlstedt (1980) and Kirby (1980) developed a rheological model of the continental crust (figure 1.1). This strength profile of the crust consists of two parts, brittle and ductile, and a transition between these. Sibson (1974) described the brittle crust by the following failure criterion:



Figure 1.1: Rheological model of the crust (Nyblade & Langston, 1995, after Brace & Kohlstedt, 1980 and Kirby, 1980).

$$(\sigma_1 - \sigma_3)(z) = \alpha \rho g z (1 - \lambda) \tag{1}$$

In this criterion, $(\sigma_1 - \sigma_3)$ represents differential stress (maximum normal principle stress – minimum normal principle stress), depending on depth *z*. The density (of the crust) is indicated by ρ , *g* is the gravitational acceleration, and α is a fault parameter. The parameter λ is the pore fluid factor, which is defined as the pore fluid divided by the overburden pressure. The line that this equation creates, is the first part of the strength profile. It increases linearly with depth following the Mohr-Coulomb failure line, with a friction coefficient that fits with Byerlee's Rule (Byerlee, 1978). This behavior dominates the upper crust, making this the brittle or seismic part. The second part, which dominates in the lower crust, is created by the power law creep indicating viscous strength (e.g. Carter & Tsenn, 1987; Kirby & Kronenberg, 1987):

$$(\sigma_1 - \sigma_3)(z) = \left(\frac{\dot{\varepsilon}}{A}\right)^{\frac{1}{n}} \exp\left(\frac{E}{nRT(z)}\right)$$
(2)

where \dot{e} is the strain rate, *R* is the gas constant, *T* is the temperature (in Kelvin) depending on depth *z*, and *A*, *n* and *E* are material properties. This lower part of the crust is generally viscous or aseismic. The transition between these two, called the brittle-ductile transition, can be found by combining the two equations, obtaining the yield strength envelope in figure 1.1. These crustal strength curves show the (shear) strength of the material, indicated by the differential stress, plotted against the depth. If the differential stress for frictional failure is less than for ductile creep, brittle failure will occur by ductile creep. This is because the medium will always go for the easiest way to deform and release stress. It will "fail" at the lowest differential stress because the rock will fail and the stress is not able to build up higher than the lowest possible failure criterion. When the differential stress for frictional failure equals the one for ductile creep, the transition takes place. The depth of the transition is primarily controlled by the temperature (and thus the geotherm), because the power law creep is temperature dependent.

Because of a high geotherm due to lithospheric thinning, the brittle-ductile transition in rift zones is generally shallow, up to 15 km (Jackson & White, 1989), with an effective elastic thickness of 10 - 15 km (Barton & Wood, 1984; Fowler & McKenzie, 1989; Bechtel et. al., 1990). For the East African Rift zone, this transition seems to be a lot deeper. Craig et. al. (2011), Yang & Chen (2010) and Albaric et. al. (2009) all report earthquakes down to 44 km depth, so the effective elastic thickness is estimated to be 20 - 40 km

(Bechtel et. al., 1987; Ebinger et. al., 1989c). Shudofsky et. al. (1987) argues that these deep earthquakes take place in the lower crust rather than the upper mantle, even though the upper mantle is usually stronger due to the olivine rheology. Crustal thickness is estimated to be around 42 km (Camelbeeck & Iranga, 1996; Fagereng, 2013), which is generally greater than the observed focal depths. Yang & Chen (2010) question the assumption that the earthquakes take place in the lower crust, and say that a few deeper events may occur in the mantle. However, they also point out that there is a gradual transition zone between the lower crust and upper mantle.

Yang & Chen (2010) also point out that there are three patterns in focal depths of the earthquakes. (1) Earthquakes only occur to a depth of about 15 km in areas beneath magmatic segments, where the rift is well-developed. (2) Two peaks in seismicity, one near 15 km (\pm 5 km), and one near 35 (\pm 5 km), can be identified for amagmatic but well-developed rift segments. This is also noted by Fagereng (2013). (3) Where rifts are underdeveloped and not yet visible at the surface, seismicity is restricted to the top 15 km of the crust. However, this area shows unusually deep aftershocks down to 35 (\pm 5 km). This pattern seems to follow rift development, which is from north to south (see *Background on Regional Geology*).

Following previous authors (Fagereng, 2013; Albaric et. al., 2009; Craig et. al., 2011; Nyblade & Langston, 1995; Foster & Jackson, 1998; Jackson & Blenkinsop, 1997; Shudofsky et. al.; 1987), I assume here that the reported deep earthquakes occur in the lower crust. This assumption poses some issues that need to be answered. The lower crust generally behaves viscous rather than brittle (Scholtz, 1988; Brace & Kohlstedt, 1980; Kirby, 1980). This means that the lower crust in East Africa must be rheologically different from lower crust elsewhere in the world. Besides earthquake depth, there are several other differences between 'normal' rift zones and the East African Rift Zone. A 'normal' rift zone is bordered by border faults, which are segmented. They show high vertical displacements. At the surface, border faults generally have high angles, and progress into the earth to form low angle detachment faults or shear



Figure 1.2: Different crustal strength profiles for varying composition and strain rate (Fagereng, 2013). a. felsic lower crust; b. mafic lower crust; c. elevated strain rate.

zones at depth. Typically, a rift is only bordered by a border fault on one side, making them asymmetric. These are called half-graben basins. The other side of the half-graben can be bordered by monoclines, outward-tilted step faults (with minor vertical offsets), or a flexure. Fault segment length is usually 20 – 25 km, with dips of about 45°. They extent to about 15 km (Jackson & White, 1989; Wallace, 1989).

However, in the East African Rift zone, border fault segments are larger (~ 100 km) and half grabens are wider (~ 50 km). These differences may all be linked to each other (Jackson & Blenkinsop, 1993).

There are several possible explanations as to why the brittle-ductile transition is deeper in the East African Rift zone. Many previous authors (e.g. Jackson & Blenkinsop, 1993; Shudofsky et. al., 1987; Nyblade & Langston, 1995; Craig et. al., 2011; Jackson et. al., 2004) suggest that old, cold, anhydrous and strong material in the crust as well as the thickness of the crust influences the depth of the transition. Jackson & Blenkinsop (1993) go on to say this also may be a fundamental control on the scale of the structures that form within it.

An additional explanation may be the presence of mafic material in the lower crust, which, with its material properties, can account for brittle behavior to a deeper level than for example felsic material. This is because its viscous strength is greater than for felsic rocks (Mackwell, 1998). Figure 1.2 (Fagereng, 2013) shows a strength profile for felsic (a) and one for mafic material (b), clearly showing that the brittle-ductile transition appears to be a lot deeper for mafic material. This indicates that invoking mafic material at depth may explain deep earthquakes (Fagereng, 2013; Albaric et. al., 2009; Shudofsky et. al., 1987; Nyblade & Langston, 1995). However, by looking at the mafic strength profile in figure 1.2, a mafic layer of at least 15 km thick is required (Fagereng, 2013), and there is no evidence for this.

Another factor must thus play a role in the issue. The power law creep is temperature dependent, and the temperature is a function of depth. The equation thus depends on the existing geothermal gradient. Lessening of the geothermal gradient will lower the temperature of the lower crust, which may be necessary to explain deep lower crustal earthquakes. Nyblade & Langston (1995) suggests reducing lower crustal temperatures in two ways: (1) raising thermal conductivities (which will lessen the geothermal gradient); and/or (2) increasing upper crustal heat production, which would lessen the amount of heat that enters the lower crust from above, thus reducing lower crust temperatures. Such high upper crustal



Figure 1.3: Thermal structure for the Malawi rift, as estimated by Fagereng, 2013.

heat production however, is unlikely for Proterozoic rocks (e.g. Jaupert et. al., 1998) and there is no evidence of elevated heat production in Africa's upper crust (Fagereng, 2013). Craig et. al. (2011) suggests that the presence of thick conductive lithosphere will act as a thermal insulator for the crust from the mantle, which lowers the geotherm. This is in agreement with Shudofsky et. al. (1987), who points out that there is no seismicity in the upper mantle, which is generally perceived as stronger (and more seismogenic) than the lower crust. They propose a thermal anomaly in the upper mantle, which heats up the upper mantle, pushing it into the ductile regime. As heat conductivity is slow, and the crust is thick, the lower crust has not yet been affected by this anomaly, keeping temperatures here low, thus keeping it in the brittle regime.

The depth of the earthquakes in combination with the geothermal gradient in the lower crust suggest that it stays seismic for well over 350°C (onset of crystal plasticity for quartz) to 450°C (onset of crystal plasticity for feldspar; Scholz, 1988). However, even with the coldest reasonable thermal structure

calculated, temperature at the depth of the earthquakes (35-42 km) would still be over 700°C, and 450°C occurs at 22 km depth (figure 1.3 - Fagereng, 2013). This means that the combination of mafic rocks in the lower crust with a low thermal gradient still does not entirely explain the seismic behavior of the lower crust.

An elevated strain rate would also explain frictional behavior at lower crustal levels (Shudofsky et. al., 1987; Fagereng, 2013). Nyblade & Langston (1995) and Fagereng (2013) estimate a strain rate of about 10^{-15} s⁻¹, using the 3 – 4 mm/yr spreading rate estimated by Stamps et. al. (2008) distributed over a 50 km wide rift (Ebinger et. al., 1987). There are several ways to increase this strain rate. When material is in the brittle regime, increasing the fluid pressure decreases the friction and therefore increases the strain rate. This could be a result of the presence of magma or other volatiles. This is however unlikely. Not only do dike intrusions accommodate strain, which requires less tectonic stress than normal faulting (Buck, 2004), magma would also greatly increase the temperature at lower crustal levels, pushing the lower crust towards the ductile regime (Fagereng, 2013). Also, by having a thick conductive lithosphere, the lower crust would be somewhat protected from rehydration by percolating metasomatic fluids (Craig et. al., 2011). This makes it seem unlikely that fluid pressures in the lower crust would be elevated.

Another way to increase strain rate is by strain localization. The crust is likely anisotropic: it has a preexisting structural fabric, because of earlier deformation phases described above. These fabrics affect the fault geometry (Versfelt & Rosendahl, 1989). This anisotropy causes development of the border faults along weak planes, because of deformation localization on these planes. This not only affects the lengthscales of brittle fractures (Fagereng, 2011; Fagereng & Sibson, 2010), it also affects the frictional properties (Collettini et. al., 2009). If deformation occurs on a weak plane, then this plane determines the frictional properties and therefore failure conditions of the crust. Deformation localizes on these planes, strongly increasing the strain rate at these points. This allows frictional failure to significantly greater depths, as can be seen in figure 1.2c (Fagereng, 2013). A condition for these weak planes causing seismicity at great depths, is that the surrounding rock must be strong, meaning it should not be accommodating strain. No viscous behavior can thus take place here, it must still be in the frictional regime (but not at failure conditions). This way, enough elastic strain can be stored to create seismic slip on the weak planes. Also, without these weak zones to accommodate strain, a mafic crust would likely behave viscous, due to temperatures being too high. Fagereng (2013) suggests that seismic behavior at great depths can take place on weak planes, where deformation is localized and strain rate is elevated, in an otherwise strong, possibly mafic, lower crust, a hypothesis in agreement with Handy & Brun (2004).

In summary, Yang & Chen (2010) point out that earthquake distribution seems to have a trend with rift development. The rift develops from north to south, and so does the earthquake depth distribution pattern. The deep earthquakes investigated here take place in an area where earthquakes occur at two depth intervals: 10 - 20 km and 35 - 45 km. They are part of the well-developed but amagmatic part of the rift in the south. Shudofsky et. al. (1987) states this is still a pre-rift phase, compared to the further developed magmatic rifts in the north. They also suggest a thermal anomaly in the mantle. This makes sense for a pre-rift phase, which means the area is already being rifted by the southwards spreading mantle plume, but this mantle plume has not yet caused the rift to be magmatic. It also explains why there are no earthquakes in the upper mantle. Temperatures here are raised so the mantle rocks behave ductile. Due to the old, cold, thick crust, and slow conductivity, the lower crust has not yet been affected by this

anomaly. Thinning of the crust and an elevated geothermal gradient have not yet affected the area, also pointing towards the pre-rift phase. The assumed feldspar (rather than quartz) rheology, makes brittle failure possible until about 450°C instead of 350°C (for quartz). Rocks are anisotropic due to previous deformation stages. It has weak zones on which deformation localizes and strongly increases the strain rate. Frictional failure can take place here, if the surrounding rock is strong. This is probably mafic material since this would increase viscous strength and thus elastic strain can be stored.

To generate these deep earthquakes, several conditions need to be met. First, the rocks must be in the brittle regime of the crust, for which several possible explanations are given above. One of these explanations involves strain localization on weak planes. Also, since earthquakes are the result of a frictional instability, frictional properties of the rocks at depth (high temperature and normal stress conditions) have to be in the unstable (or conditionally stable) regime: (a-b) values have to be negative. These are material properties, which will be further explained in the section *Background on Earthquake Mechanics and Rate and State Friction*. Here, I do rotary shear experiments on several rock samples from Malawi. They were tested for trends in temperature and pressure going up to 600°C and 250 MPa. The aim is to determine the frictional properties of these rocks, learn more about the stability regimes and the brittle-ductile transition within these rocks under high pressures and temperatures. This may give clues to the possibilities of earthquakes at depth. Also, since localization on weak planes may play a crucial role in increasing the strain rate and lowering the brittle-ductile transition, microstructures from the shear experiments are studied to learn more about the deformation processes and strain localization.

1.2 Background on regional geology

At a triple junction, three plates meet at one point. There can be different types of triple junctions, depending on the type of movement between plates. At the Afar Triangle in Ethiopia, the Arabian plate meets the African plate, which is rifting apart to form two plates: the Somalian and Nubian plate. The relationship between the plates can be seen in figure 1.4. The Nubian plate moves northwest, while the Somalian plate moves east, and the Arabian plate north-north-east. All three plates are moving away from each other, making this the only ridge-ridge triple junction system above sea level in the world. This creates the East African Rift (EAR) system, a zone of continental extension.

Archaean (4000 - 2500 Ma) (Tanganyika) cratons are composed of greenstone belts in between plutons unrelated to plate tectonics. Wrapped around these thick, stable cratons are Proterozoic mobile belts (2500 - 541 Ma) (Ebinger et. al., 1987; Versfelt & Rosendahl, 1989). Several orogenies created these highly folded mobile belts. The Ubendian orogeny took place between 2250 - 1800 Ma. During this period, a belt of tightly folded rocks, with NW-trending fold axes formed (Fitches, 1970). Following the Ubendian orogeny, the Kiberan orogeny took place from around 1400 Ma until around 1000 Ma. It existed of several orogenic events. One of these events folded basement rocks along NE-SW axes. These are now found on the western central margin of the Malawi rift in the Irumidian belt (Carter & Bennett, 1973). A more recent orogeny that took place is the Mozambican orogeny (700 – 400 Ma), which deformed rocks on the eastern margin of the Malawi rift. The deformation processes involved are complex, but most likely involved plastic deformation and high grade metamorphism (Cannon et. al., 1969). The rocks show N-S orientated



Figure 1.4: African plate tectonics, edited from <u>https://pubs.usgs.gov/gip/dynamic/East_Africa.html</u> and <u>http://africa-arabia-plate.weebly.com/future-tectonics.html</u>.

grains (Cannon et. al., 1969). It overprinted parts of the Ubendian and Kiberan orogenic belts (Carter & Bennett, 1973). Within the mobile belts formed by these orogeny's lie transcontinental dislocation zones (TDZs), which may influence the structure of the rift by changing the polarity of the rift (Versfelt & Rosendahl, 1989). During the Permo-Triassic period (299 – 201 Ma) sediments were deposited and there were active volcanics (Ebinger et. al., 1987). These rocks can be found near or within the boundaries of the Malawi rift system, and possibly extend beneath Lake Malawi. During the Jurassic-Cretaceous period (201 – 66 Ma), a phase of rifting took place, creating alkaline plutons and dikes with NE trending lineament (Wooley & Garson, 1970; Carter & Bennett, 1973).

Between 45 – 37 Ma, mantle plume activity caused volcanism in southern Ethiopia and northern Kenya (Ebinger et. al., 1993a; Furman et. al., 2006; McDougall & Brown, 2009; Roberts et. al., 2012). The rift system initiated here, with extension and uplift of rift shoulders starting around 45 – 40 Ma (Ebinger et. al., 1989a; Morley, et. al., 1992; McDougall & Brown, 2009; Roberts et. al., 2012). The mantle plume spread around 30 Ma, causing volcanism in central Ethiopia, including the Afar region, and Yemen. This was accompanied by broad thermal uplift (Burke, 1996; Pik et. al., 2008; Roberts et. al., 2012). With time, volcanic activity progressed southwards. Rift extension spread to more regions between 30 – 20 Ma (Ebinger et. al., 1989a; Morley, et. al., 1992; McDougall & Brown, 2009; Roberts et. al., 2012). The East African Rift zone is generally divided into the Eastern and the Western branch. The Eastern branch was well established by 20 Ma (Ebinger et. al., 1989a; Ebinger, 1989b; Wolfenden et. al., 2004; Chorowicz, 2005; Roberts, 2012). However, the age of the Western branch is debated. Ebinger et. al. (1989a), Cohen



et. al. (1993) and Tiercelin & Lezzar (2002) argue and show evidence for the initiation of the Western branch around 12 Ma, making the Western branch considerably younger than the Eastern branch. More recent data by Roberts et. al. (2012) show that the Western branch started 14 Ma earlier than this, around 26 Ma. Spreading rates across the entire rift system are estimated from kinematic models to be < 3 -4 mm/yr in ENE-WSW orientation (DeMets et. al., 1990; Jestin et. al., 1994; Roberts et. al., 2012).

Once extension starts, the development of different rift systems follows a similar pattern (Ebinger et. al., 1989a). First, border faults develop. After this, the formed basin becomes asymmetric as the flank is uplifted and the basin subsides. On the opposite side of the rift, monoclines develop. The basinal asymmetries are enhanced by continued subsidence of the basin and tilting along intrabasinal faults (Ebinger et. al., 1989a).

The Malawi rift is the southernmost part of the Western Branch, and extends over 900 km. It runs from the Rungwe volcanic province in the north to the Urema graben in Mozambique. Volcanism in this region is restricted to alkalic volcanics at the northern end of the lake (Harkin, 1960; Ebinger et. al., 1987). Figure 1.5 shows the regional geology of Lake Malawi (Fagereng, 2013). The rift system is segmented and consists of 100 km long and 50 km wide half grabens. These are bordered by high angle border faults on one side, and the opposite side is characterized by either ENE dipping monoclines, en echelon normal faults with small throws, or relay ramps (Ebinger et. al., 1993a). The flanks of the rift are tilted away from the rift valley (Dixey, 1937; Carter & Bennett, 1973). The border fault segments follow the Proterozioc mobile belts, but do not seem to follow lithological or tectonic contacts (Ebinger et. al., 1987). The vertical offset of a border fault segment is biggest at the central part. Here, orientations tend to be oblique to preexisting structures as well (Ebinger et. al., 1987). Versfelt &

Rosendahl (1989), state that rift zones follow the basic 'grain' of the mobile belts. It seems like border fault systems do not follow lithological or tectonic contacts (Ebinger et. al., 1989a; Versfelt & Rosendahl, 1989). Pre-rift fabrics (and the orientation of the stress field) do seem to influence the way that half grabens are linked (Versfelt & Rosendahl, 1989). The half graben basins created by the border fault segments create accommodation zones that trend oblique to the border fault segments. This supports the tectonic model shown in figure 1.6 (Ebinger et. al., 1987).



1.3 Background on Earthquake mechanics and rate-and-state friction

Earthquakes are (usually) the result of a stick slip frictional instability on an existing weak plane or fault. Faults usually contain frictional wear material as unconsolidated fault gouge between the surface of two planes. This gouge determines the frictional properties of the fault best in accordance with the rate and state friction theory (Dieterich, 1978, 1979; Ruina, 1983; Marone, 1998; Scholz, 1998). Sliding begins

when the ratio of shear stress to normal stress overcomes the static friction (Scholz, 1998):

$$\mu_s = \frac{\tau}{\sigma_n} \tag{3}$$

This static friction, which corresponds to the maximum frictional resistance, increases logarithmically with hold time (Marone, 1998). After this frictional resistance is overcome, dynamic friction μ_d , or sliding resistance, controls the behavior of the fault. Dynamic friction depends on the sliding velocity *V* (Scholz, 1998). If the system is subjected to a sudden change in sliding velocity *V*, friction evolves over a critical slip distance D_c to a new steady state value. This appears to be the distance necessary to renew the surface contacts. The state variable ϑ represents the average contact lifetime, defined by the ratio D_c/V .

$$\mu_{1} = \mu_{0} + a \ln \left(\frac{V}{V_{0}} \right) + b \ln \left(\frac{V_{0}\theta}{D_{c}} \right)$$

$$\frac{d\theta}{dt} = 1 + \left(\frac{V\theta}{D_{c}} \right)$$
(5)

In this law (Dieterich "slowness" Law/state evolution law), *a* and *b* are material properties for describing friction, in which *a* is seen as the direct effect and *b* as the evolutionary effect. Rewriting this equation (assuming steady state friction: $\frac{d\theta}{dt} = 0$) gives:

$$(a-b) = \frac{d\mu_{ss}}{\ln(V_1/V_0)}$$
(6)

In this equation, $d\mu_{ss}$ is the difference between steady state friction before and after the sudden increase in velocity (velocity step), V_0 is the sliding velocity before the step, and V_1 is the sliding velocity after the step. In rate and state friction, the stability of the system is independent of the frictional strength of the system. It depends on effective normal stress $(\bar{\sigma}_n)$, shear stress τ , k (stiffness of the system), a, b, (a-b)and D_c . The velocity dependence of steady state friction, which determines the stability of sliding, results from an interplay between a and b (Scholz, 2002). There are three stability regimes. If (a-b) > 0, the material is velocity strengthening, i.e. frictional strength is higher after the velocity step than before. This will always lead to stable sliding. If $(a-b) \le 0$, there is a Hopf bifurcation to consider, which occurs at a critical value of effective normal stress $((\bar{\sigma}_n)_c)$:

$$(\bar{\sigma}_n)_c = \frac{kD_c}{-(a-b)}$$
(7)

Sliding is unstable under quasistatic loading if $\bar{\sigma}_n > (\bar{\sigma}_n)_c$. If $\bar{\sigma}_n < (\bar{\sigma}_n)_c$, it is in the conditionally stable regime, meaning sliding is stable under quasistatic loading, but unstable under dynamic loading if subjected to a jump in velocity exceeding ΔV (Scholz, 1998). Earthquakes can nucleate only in the unstable regime and may propagate into conditionally stable regions. If they propagate into a stable region, a large energy sink from a negative stress drop will stop propagation of the earthquake.

2. Methods

2.1 Sample Preparation

The samples were crushed with a pestle and mortar and sieved until all particles were smaller than 0.5 mm. To obtain samples with a narrow grain size distribution, a Retsch XRD-Mill McCrone was used. This XRD-Mill gently crushes a sample to a smaller grain size. This leads to a narrow particle size distribution while the crystal lattice is preserved. This is useful both for the XRD analysis and the ring shear experiments. An explanation of the XRD procedure is given in the next section. A narrow particle size distribution is necessary so that large particles of a specific mineralogy don't block the signal of the smaller grains with another mineralogy, so that all particles are detected. A larger particle size distribution might lead to not recognizing certain phases in a sample, or inaccurately estimating the percentage of a certain phase in the sample: the accuracy of measurement is increased with a narrower particle size distribution. For shear experiments, a finer starting grain size and more narrow grain size distribution reduces the amount of displacement necessary to reach steady state.

No.	Sample	Material	Intensity	Time	Grain Size	Grain Size 2
1	M01	felsic	4	10	4.88 - 5.96 - 6.63	
2	M01	felsic	4	5	6.63 - 7.72	
3	M17	mafic	4	10	2.65 - 3.09 - 3.60	
4	M17	mafic	4	5	4.88 - 5.69	
5	M01	felsic	4	15	2.65 - 3.09	
6	BM 5-1	calc-silica	4	5	4.88 - 5.69 - 6.63	0.31 - 0.42
7	BM 5-1	calc-silica	4	10	3.09 - 3.60	0.31 - 0.36
8	BM 5-1	calc-silica	4	15	1.95 - 3.09	0.36 - 0.42
9	M01	felsic	4	12	3.60 - 4.2	
10	BM 5-1	calc-silica	4	6	5.69 - 6.63	0.42 - 0.46
11	M01	felsic	4	11	4.2-4.89	
12	BM 5-1	calc-silica	4	7	4.2-4.89	

Table 2.1: Results from particle size analysis after use of the XRD Mill. The results in red are used for experiments. The grain size indicates in between which sizes the size of most of the grains in the sample are, or the peak in distribution. Sometimes there are three numbers because different analyses gave different results.

The three samples used in this research are described in the materials section. For these different samples, different milling times were used to obtain a similar grain size. The target grain size was 5 μ m. Grain size analysis was done with a laser diffraction particle size analyzer. The grains are brought into a suspension with water. Then, a laser beam is directed at the sample. The diffraction pattern of this beam as it passes

through the sample says something about the grain size: smaller particles scatter the beam at higher angels, and larger particles at smaller angles. Table 2.1 shows the results from the grain size analyses for the three different samples at different grinding times. The times in red were used to make a large batch used in the experiments. For M01 (the felsic sample), a grinding time of 10 minutes came closest to the target grain size of 5 µm. Figure 2.1a shows the particle size distribution for this sample. For M17 (the mafic sample), the distribution can be seen in figure 2.1b. The resulting time to obtain a mean grain size of 5 µm is 5 minutes. Different times are needed for different samples because different minerals have a different hardness, making it easier or harder to crush samples and thus taking either a shorter or longer time to obtain the same grain size for different samples. For sample BM 5-1 (a calc-silicate), this problem seems to occur within the same sample. Figure 2.1c shows that there are two distribution peaks in the image of the particle size analyses for BM 5-1. The higher (and bigger) one is used as an indicator for the 5 µm aim. This result was achieved in a grinding time of 5 minutes. The results in table 2.1 shows that the crushing is not very precise. The grain size changes per batch: a specific time does not guarantee a specific grain size for the same sample. For example, sample BM 5-1 after 5 minutes of grinding (no. 6) shows a smaller grain size than after 6 minutes of grinding (no. 10).

2.2 XRD & ICP

Different techniques have been used to analyze the composition of the sample. The results and conclusions of these can be found in the Materials section. In X-Ray Diffraction (XRD) Analysis, the d-spacing of a material is measured by letting x-rays interact with the sample and satisfying Bragg's Law: $n\lambda = 2dsin\theta$. In this equation, *n* is a positive integer, λ is the angle of the incident wave, ϑ is the scattering angle and *d* is the d-spacing or interplanar distance. When an x-ray is shot at a sample, the x-rays are scattered in all directions by the atoms they collide with. In most of these directions, waves cancel each other out by destructive



Figure 2.1: Particle size distribution after crushing for a. the felsic sample; b. the mafic sample; c. the calcsilicate sample.

interference. But in a few specific directions, Bragg's Law is satisfied and constructive interference gives an interference pattern. Which specific directions these are and what the interference pattern looks like depends on the d-spacing of the crystal lattice: the distance between two adjacent crystal planes. Each mineral has a specific set of d-spacings, so qualitative analysis to determine the composition of a sample can be done. However, mistakes in interpreting an XRD spectrum can easily be made, as spectra may look similar for some minerals. For quantitative analysis, XRD is less suited. To make the quantitative analysis more accurate, 10% corundum (Al₂O₃) is added to the sample. While analyzing the spectra, one can now keep in mind that the amount of corundum in the sample should be about 10%, giving more accurate percentages for the other minerals in the sample as well.

The second method used to calculate the mineral composition of the samples is Inductively Coupled Plasma Mass Spectrometry (ICP or ICP-MS). This method gives no indication of minerals in the sample, but gives a quantitative analysis of the elements in the sample. The sample is dissolved in a strong acid so that the elements can be ionized by the plasma. They then go through a mass spectrometer to be separated by their mass to charge ratio. This ratio is specific for each element, so that the exact amount of an element in the sample can be measured.



Figure 2.2: section of the sample used for microstructural study: normal to the shear plane, and subparallel to the shear direction (Verberne et. al., 2015).

2.3 SEM & EDX

A tabletop Scanning Electron Microscope (SEM) was used for two reasons. (1) Pictures were made to study microstructures. After the experiment, the samples were dried and put in epoxy. They were then cut so that a thin section could be made from which the inside of the sample could be studied. Figure 2.2 (Verberne et. al., 2015) shows what part of the sample was studied. The sample was cut in half so the inside, normal to the shear plane and subparallel to the shear direction, could be looked at (Verberne et. al., 2015). After the thin sections were made, they were coated with a Pt/Pd layer to conduct the charge from the SEM. The SEM shoots an electron beam at the sample, of which the electrons interact with the atoms in the sample. Several detectable signals are produced, including

secondary electrons and backscattered electrons (BSE). Secondary electrons are electrons that are kicked out of an atoms electron shell (usually the k-shell) by interacting with an electron from the electron beam moving through the sample. They have low energy and therefore won't go further than the first few nanometers of the sample, i.e. they only interact with the surface. Secondary electron imaging is therefore used to study the surface of a sample. In this case, the surface of the sample has been polished, so information may be altered or lost. To create our images, backscattered electrons are used. These are electrons from the electron beam which have been elastically scattered. The electrons scatter when interacting with the atoms in the sample. The heavier the atom, the more backscatter it produces. Therefore, a contrast appears between atoms based on atomic mass. We use this method to look at the samples, because it provides more information on the location, orientation, grain size and shape of different minerals and porosity. (2) Thin sections were also made of pieces of the original hand samples. To further analyze the composition of the samples, EDX (Energy Dispersive X-ray spectroscopy) was used to create element maps. When the SEM's electron beam hits the sample, electrons get kicked out of their shells. Electrons from higher energy shells then fill the created vacancies in the lower energy shells. To balance the energy difference between the shells, the "left-over" energy is emitted in the form of x-rays.

These set of x-rays are specific for each element, and can therefore say something about the elements in the sample. The element maps that were created were then overlaid in ImageJ to find correspondence

between elements and possibly minerals within the sample.

2.4 Pressing Rings

Three test experiments were performed in which a lot of material was lost. Therefore, pre-pressed compacted rings were made to minimalize material loss. Also, as porosity affects the friction coefficient in various ways, it is convenient to press rings to control the starting porosity of the sample. Sample thickness can also be controlled by pre-pressing them into compact rings. For these experiments, I used a porosity of 15%, and a starting sample thickness of 1.25 mm. The inner diameter

of the ring is 22 mm and the outer diameter is 28 mm, so that it matches the piston set of the ring shear. The volume of the ring would therefore be:

$$V = \pi \cdot r_{out}^2 \cdot h - \pi \cdot r_{in}^2 \cdot h = \pi \cdot 1.4^2 \cdot 0.125 - \pi \cdot 1.1^2 \cdot 0.125 = 0.2945 \, cm^3$$

If the porosity should be 15%, then 15% of this volume is pore space, and 85% is sample. Assuming the average density of the crust (2.8 g/cm³) as the density of the sample, the mass of the sample that should be pressed into the ring is:

$$mass_{sample} = \rho \cdot V \cdot 85\% = 2.8 \cdot 0.2945 \cdot 0.85 = 0.70097 \ g$$

Water is added so that the pore space in filled with water. This gives the sample the cohesion it needs to not fall apart after pressing. Since the targeted porosity is 15%, 15% of the volume should be water. Distilled water is used because this is less likely to react chemically with the sample. The mass of the distilled water (density = 1 g/cm^3) to be added to the sample is:

 $mass_{water} = \rho \cdot V \cdot 15\% = 1 \cdot 0.2945 \cdot 0.15 = 0.04418 \, g$

The water is mixed as evenly as possible with the sample and the mixture is then deposited into a ring-shaped die (Figure 2.3) with an outer diameter of 28 mm and an inner diameter of 22 mm (the dimensions of the pistons of the ring shear). It is put under a hydraulic press for 20 minutes at 208 MPa (5 ton).

After pressing, the ring is carefully removed from the die and put in a 50°C oven overnight. This evaporates the water from the sample, leaving the sample with 15% porosity. Figure 2.4 shows an example of the resulting ring. In table 2.2, the details of all the pressed rings are summarized. Number 4 and 8 in gray are rings that were not used, because the experiment failed or because the ring was incohesive. The thickness of the ring is consistently larger than the aimed thickness. This could be due to the density of the sample material being different from the average density of the crust. A better explanation may be that the porosity of the sample is larger than the targeted 15%. This would mean that not all the pores were filled





Figure 2.3: ring shaped die of dimensions equal to the ring shear pistons: outer diameter or 28 mm and inner diameter of 22 mm.

No.	Sample	Experiment	Mass	Mass	Thickness	Mass	Thickness	Thickness
			sample	water	from caliper		before	after
1	M17	u556	0.70109	0.04485	1.5	0.64753	1.45	0.8
2	M17	u557	0.70107	0.04493	1.6	0.66342	1.59	0.98
3	M17	u558	0.70103	0.045	1.6	0.65631	1.6	1
4	M01		0.70098	0.04469	1.8	0.66599		
5	M01	u559	0.70117	0.04462	1.9	0.67108	1.76	1.46
6	M01	u560	0.70088	0.04464	1.85	0.66994	1.5	0.79
7	M01	u561	0.70123	0.04447	1.6	0.6612	1.64	
8	BM 5-1		0.70101	0.04468	1.55			
9	BM 5-1	u571	0.70105	0.04475	1.6	0.6489	1.44	1.07
10	BM 5-1	u575	0.70149	0.04505	1.6	0.65976	1.33	
11	BM 5-1	u576	0.70095	0.04472	2.15	0.6547		0.93
12	BM 5-1	u578	0.70092	0.04485	1.6	0.63361		0.43
13	M17	u579	0.70122	0.0452	1.6	0.66083	1.58	1.19
14	M17	u580	0.70034	0.0444	1.6	0.66523	1.56	1.06
15	M01	u581	0.70113	0.04452	1.7	0.66673	1.68	1.38
16	M01	u582	0.70117	0.04451	1.6	0.65402	1.56	1.17
17	BM 5-1	u583	0.70107	0.04479	1.6	0.59368	1.58	0.9
18	BM 5-1	u584	0.70067	0.0449	1.5	0.64695	1.43	0.89
19	M17	u598	0.70113	0.045	1.6	0.66276	1.48	
20	M01	u599	0.70104	0.04487	1.15	0.67054	1.73	0.87
21	BM 5-1	u601	0.70086	0.04491	1.5	0.62726	1.22	0.58
22	M17	u608	0.70107	0.04447	1.8	0.6656	1.35	0.83
23	M01	u609	0.70122	0.04464	1.85	0.6863	1.68	1.01
24	M01	u610	0.7012	0.04486	1.8	0.684	1.69	0.96
	•		•		-			

with water. While the thickness as measured from the caliper is generally in accordance with the thickness as measured from the piston, the thickness measured from the caliper is the more reliable one.

Table 2.2: Pre-pressed rings details.

2.5 Ring shear apparatus

The ring shear apparatus as described by Niemeijer et. al. (2008) and Den Hartog et. al. (2012) was used to perform rotary shear experiments. The apparatus is shown in figure 2.5 (Den Hartog et. al., 2012). The pre-pressed ring is put in between an upper and a lower internal piston, and is kept in place by two confining rings coated with a MoS anti-friction spray. The position of the sample (gouge) in the pistons is shown in figure 2.5c (and figure 2.5b). An insulating mullite tube, which prevents convection of the pore fluid, is placed over the upper piston. The pistons are attached to the piston head, and the piston head is protected against corrosion from the pore fluid by a teflon ring in between the mullite tube and the piston head. This set is placed into the pressure vessel, shown in figure 2.5b. The vessel is closed by the upper closure nut. Pore fluid pressure and temperature inside the vessel can be controlled. The pore fluid



pressure is regulated by a hand pump outside the vessel. The vessel is internally heated by a furnace element. There are two sensors that measure the temperature. The furnace temperature is measured directly on the furnace element using a K-type thermocouple and regulates the heating of the furnace. A K-type thermocouple directly adjacent to the sample (Figure 2.5c) measures the sample temperature. The vessel is cooled by cooling water. Seals inside the vessel and the upper closure nut prevent the leaking of both pore fluid and cooling water.

After putting the pistons inside the vessel, closing the upper closure nut, and connecting the cooling water and pore fluid, the entire vessel is lifted into an Instron loading frame (figure 2.5a). The Instron loading ram applies the normal stress onto the sample. The lower forcing block on which the vessel is placed, is coupled to a motor. This motor drives the rotation (at several rotational velocities) of the lower forcing block and therefore the vessel. On top of the vessel, the upper forcing block is kept stationary. Because this is kept stationary, the upper internal piston does not move, while the lower internal piston is moved

No.	Sample	Normal Load (MPa)	Fluid Pressure (MPa)	Temperature (°C)	Run-In (μm/s)	Velocity steps (µm/s - s)
u553	M01	100	100	roomT	1 (5000s)	3 (600) - 10 (300) - 30 (90) - 100 (9)
u554	M01	100	100	300 - 400	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u555	M01	100	100	roomT - 100 - 200	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u556	M17	100	100	300 - 400	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u557	M17	100	100	500 - 600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u558	M17	100	100	roomT - 100 - 200	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u559	M01	100	100	300 - 400	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u560	M01	100	100	roomT - 100 - 200	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u561	M01	100	100	500 – 600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u571	BM 5-1	100	100	300 – 400	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u575	BM 5-1	100	100	roomT	1 (5000s)	
u576	BM 5-1	100	100	roomT - 100 - 200	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u578	BM 5-1	100	100	500 - 600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u579	M17	100	100	500	1 (5000s)	
u580	M17	100	100	300	1 (5000s)	
u581	M01	100	100	500	1 (5000s)	
u582	M01	100	100	400	1 (5000s)	
u583	BM 5-1	100	100	200	1 (5000s)	
u584	BM 5-1	100	100	100	1 (5000s)	
u598	M17	200	100	500-600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u599	M01	200	100	500-600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u601	BM 5-1	200	100	500-600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)
u608	M17	100	100	roomT	1 (5000s)	
u609	M01	100	100	roomT	1 (5000s)	
u610	M01	250	100	500-600	1 (3000s)	3 (300) - 10 (90) - 30 (30) - 100 (9)

Table 2.3: Performed Rotary shear experiments. The gray ones were test experiments and the data is not used. The experiments indicated in blue were performed to create and study microstructures.

in a clockwise rotation. This creates a dextral shear on the sample. Two load cells in the upper forcing block measure the torque applied to the system by the rotational movement which is then converted into shear stress.

The performed experiments are summarized in table 2.3. Several types of experiments were done. The experiments in gray were test experiments, after which I decided to press rings. Their data has not been used. Velocity stepping experiments are indicated in black. They were performed at varying temperature and normal stress conditions, to test the dependence of friction on these variables. The used normal stresses were 100 MPa, 200 MPa and 250 MPa. The tested temperatures are room temperature, 100°C, 200°C, 300°C, 400°C, 500°C and 600°C. The pore fluid pressure is the same in all the experiments: 100 MPa (1000 bar). To achieve steady state, a run-in was performed at 1 µm/s for 3000 seconds, creating a



displacement of 3 mm. After this, four velocity steps were performed to 3 µm/s, 10 µm/s, 30 µm/s and 100 um/s. to test the velocity dependence of friction and measure the material properties a and b from the rate and state friction theory (see "Background on Earthquake mechanics & Rate-and-State Friction"). The total displacement after one set of velocity steps is 6.6 mm. The experiments in blue are the experiments done to create microstructures. The samples were sheared at 1 µm/s for 5000 seconds, creating 5 mm displacement. The lighter blue experiments were not used because they showed the same behavior (instabilities or stick slips) as one of the previous experiments. The six dark blue experiments are two experiments from each sample, one which shows stable sliding, and one which shows stick slips. Thin sections were made to study the (possible) differences in microstructure between the different sliding behaviors. They were studied both in the SEM and the light microscope.

The friction coefficient of the samples in these experiments is calculated with: $\mu = \tau / \sigma_n^{eff}$

The effective normal stress is controlled by the Instron loading frame. The shear stress τ is measured in the ring shear by the torque on the upper forcing block. Two types of load cells are used for this. The two load cells of 1.2 kN (full scale, individual range), used for the 100 MPa experiments, have a range of 60 MPa, and a resolution of 0.0001%. The resolution is 0.00012 kN or 0.006 MPa, which is 1.41 N on the shear stress. The two load cells of 20 kN (full scale, individual range), used for the 200 MPa and 250 MPa experiments, have a 997 MPa range, and a resolution of 0.0997 MPa, which is 23.5 N on the shear stress. In both cases this analogue resolution is bigger than the digital resolution caused by the 16 bit A/D converter, which collects the data. Digital resolution is 0.018 N and 0.31 N respectively. This means that the noise on the shear stress data should be smaller than the analogue resolution. Figure 2.6 shows the time plotted against the shear stress, zoomed in on a section where the shear stress does not change. The noise seen is 0.0003 MPa for the 1.2 kN load cells, and 0.02 MPa for the 20 kN load cells, both smaller than their analogue resolutions.

This shear stress resolution can be extrapolated to an error on the friction data. For the 100 MPa experiments, this error would be 0.00006. For the 200 MPa experiments, this is 0.0005, and for the 250 MPa it's 0.0004. However, this is only the error calculated due to the shear stress, and not the normal stress. This means the actual error on the friction data might be slightly different.

The data is logged using LabView. Analysis of the data is done using Xlook, a modelling program designed for rate and state friction data. In this program, the data is corrected. The normal and shear stress needs to be corrected for the seal friction within the pressure-compensated piston head. The displacement is corrected for the stiffness of the machine. After all the data has been corrected, values for a, b, (a-b), etc. were obtained by performing inverse modelling of the velocity steps. The velocity steps could only be modelled for stable sliding behavior. The (a-b) values for velocity steps that involved stick slips were calculated by hand using the following equation:

$$(a-b) = \frac{d\mu_{ss}}{\ln(V_1/V_0)}$$

This equation is explained in the rate and state friction part in the introduction.

3. Materials

3.1 Chosen samples

There were a lot of samples from the location available for this research. Three samples were chosen. As a lot of research says the lower crust must be composed of mafic material, the most mafic sample was picked to look at. This is sample M17 (figure 3.1b). It is not found very often at the surface, but may be more common at depth. It is described as a retrogressed basement gneiss.

As a comparison to the mafic sample, a felsic sample was chosen as well. M01 (figure 3.1a) is a commonly found felsic sample, described as a quartzo-feldspathic felsic basement gneiss. It was retrieved from the Mugesse fault zone, at the margin of a retrogressed zone.



The third sample is BM 5-1 (figure 3.1c), which is a calc-silicate, recognized as mostly diopside from the hand sample. In the field, this unit seems mostly undeformed, but is surrounded by tightly folded felsic gneiss. It was sampled from the footwall of the Bilila-Mtakataka fault. The fault bends around the calc-silicate and localizes within the surrounding gneiss.

The locations of these three samples are indicated in figure 1.5 (from the introduction; Fagereng, 2013). The composition of these samples were analyzed and the results are discussed in this chapter to find a fitting mineral composition.



Figure 3.2: Element maps of sample M01 made with EDX.

3.2 Microscope, SEM and EDX Element maps

For sample M01, figure 3.2 shows the element map creating by EDX. The backframe is the SEM backscattered picture. From this picture, at least 4 different phases can be recognized. Figure 3.3 shows two overlap pictures made with ImageJ. In the first one (left), potassium (K) is indicated in red, sodium (Na) in green, iron (Fe) in blue and titanium (Ti) in pink. This shows at least 5 different phases. As it shows more 'holes' than the backframe, silicon (Si) in yellow is added in the second picture (right). This fills up most of the 'holes'. The picture shows that there are at least 6 phases present in the M01 sample. Table 3.1 shows an overview of the observations that can be made from the element maps. The color is the color the phase has in figure 3.3b. The definite element is the element that it used in figure 3.3 to make

the phase recognizable. Other elements are elements that are recognized to be present in the phase from the element maps in figure 3.2. The magnesium (Mg) is put in brackets because the element map seems a bit unclear. The amount is the estimated percentage

Phase	Color	Definite element	Other elements	Amount					
1	Yellow	Si	(Mg)	10%					
2	Green	Na	Al, (Mg)	60%					
3	Red/Orange	К	Al, (Mg)	20%					
4	Blue	Fe	(Mg)	3%					
5	Pink	Ti	(Mg)	2%					
6 Purple Fe + Ti Ca, (Mg) 5%									
Table 3.	1: Phases recog	nized from EDX maps i	n sample M01.						

of the phase present in the sample. Later these results will be discussed and taken into account when concluding the final mineral composition.

Phase	Mineral	Formula	Amount
1	Hornblende	Ca2(Mg, Fe, Al)5(Al, Si)8O22(OH)2	80%
2	Plagioclase	$NaAlSi_3O_8 - CaAl_2Si_2O_8$	10%
3	Quartz	SiO ₂	5%
4	Ilmenite	FeTiO ₃	5%
5	Sphene/Titanite	CaTiSiO₅	<1%

Table 3.2: Mineral composition according to light microscope analysis by (unpublished source).

Sample M17 was not analyzed in the SEM/EDX. A PPL picture from the light microscope is shown in figure 3.4. A couple of phases were recognized. The sample consists mostly of hornblende (80%). The composition as defined through the light microscope is summarized in table 3.2. Cheung (2017) describes the sample as having a distinct fabric defined by the alignment of hornblende grains and the orientation of quartz-plagioclase aggregates.



Figure 3.3: EDX map overlap of sample M01, showing 5 different phases present.



Figure 3.4: Light microscope picture in PPL of sample M17.

Figure 3.5 shows the element maps of sample BM 5-1. From the backframe it is clear that two phases are present in this sample. The colored picture in figure 3.6 shows a very small third phase. Aluminum (AI) is in red, magnesium (Mg) is in blue, and Ca in yellow. Table 3.3 shows the summary of the observations from these element maps. The amount presented in the table does not seem to fit with the picture. Figure 3.7 shows two zoomed out SEM (backscattered) pictures that show a more representative distribution of the phases in the sample. This is where the phase amount in the table is based on.

Phase	Color	Definite element	Other elements	Amount	
1	Red	Al	Na, Ca	90-95%	Table 3.3: Phases recognized from EDX
2	Purple/yellow	Mg, Ca	Si, minor Fe	5-10%	element maps in sample BM 5-1
3	Yellow	Са	-	<1%	



Figure 3.5: Element maps of sample BM 5-1 made with EDX.



Figure 3.6: EDX map overlap of sample BM 5-1, showing 3 different phases.



3.3 XRD Results

Figure 3.8 shows the XRD spectra for the three samples. Table 3.4 shows the mineral compound of sample M01 according to the XRD. It seems like the sample mostly consist of Albite (Na endmember of plagioclase feldspar). What is striking in this felsic sample is that there is not as much quartz in the sample as one would expect for a felsic sample. However, it is assumed that the part of the crust in East Africa where the deep earthquakes occur consists mostly of feldspar. This is not inconsistent with the XRD result.

Mineral	Formula	Percentage	Percentage
			without Al ₂ O ₃
Quartz – low	SiO ₂	7.83 %	8.72 %
Albite	AlNaSi ₃ O ₈	56.55 %	62.95 %
Muscovite – 2M1	$AI_{2.68}F_{0.04}Fe_{0.3}K_{0.96}Na_{0.04}Si_{3.08}O_{11.96}$	25.46 %	29.456 %
Corundum	Al ₂ O ₃	10.17 %	
Table 3.4: Mineral	composition of sample M01 accordin	ig to XRD	

In table 3.5, the XRD results of the mafic sample (M17) are presented. Ferrotschermakite is part of the hornblende group. These results thus agree with the microscope observation that the sample consists mostly of hornblende. However, there is a pretty

big difference between the 80% estimate from the microscope, and the 58% the XRD finds.

For sample BM 5-1, the XRD only finds diopside, as can be seen in table 3.6. From the SEM pictures and

EDX maps, we know this is not the case. Since the particle size analysis gave two peaks for this sample after crushing it with the XRD Mill, the

Mineral	Formula	Percentage	Percentage	
			without Al ₂ O ₃	
Quartz – low	SiO ₂	14.47 %	16.07 %	
Ferrotschermakite	$AI_{3.3}Ca_{1.86}Fe_{2.4}H_{1.98}K_{0.14}Mg_{1.22}Mn_{0.02}Na_{0.23}O_{24}Si_{6}Ti_{0.1}$	51.8 %	57.54 %	
Oligoclase	$AI_{1.277}Ca_{0.277}Na_{0.723}O_8Si_{2.823}$	23.76 %	26.39 %	
Corundum	Al ₂ O ₃	9.97 %		
Table 3.5: Mineral o	composition of sample M17 according to XRD			

smaller particles may not have been recognized by the XRD. These smaller particles are most likely a softer mineral, as they got crushed more easily.

Mineral	Formula	Percentage	Percentage
			without Al ₂ O ₃
Diopside	$AI_{0.16}Ca_{0.96}Fe_{0.05}H_{0.06}Mg_{0.86}Na_{0.04}O_{6}Si_{1.9}Ti_{0.02}$	88.9 %	100 %
Corundum	Al ₂ O ₃	11.1 %	
Table 3.6: N	Aineral composition of sample BM 5-1 accordi	ng to XRD	

M01 (including Al2O3)

20



40

2Theta (Coupled TwoTheta/Theta) WL=1,78897

60

Figure 3.8: XRD spectra for sample M01, M17 and BM 5-1.

3.4 ICP results & Conclusion

	M01			M17			BM 5-1		
Element	ppm (mg/kg)	Molar mass	mol	ppm	Molar mass	mol	ppm	Molar mass	mol
Al	108003.04	26.9815385	4.002849578	61246.64	26.9815385	2.269946171	36153.03	26.9815385	1.339917292
Ва	935.97	137.327	0.00681563	116.7	137.327	0.000849796	33.68	137.327	0.000245254
Ве	0	9.0121831	0	0	9.0121831	0	0	9.0121831	0
Са	10555.84	40.078	0.263382404	60595.53	40.078	1.511939967	106248.77	40.078	2.651049703
Со	0	58.933194	0	0	58.933194	0	0	58.933194	0
Cr	0	51.9961	0	0	51.9961	0	0	51.9961	0
Cu	0	63.546	0	0	63.546	0	0	63.546	0
Fe	31719.09	55.845	0.567984421	104708.43	55.845	1.874983078	18177.89	55.845	0.325506133
К	31169.05	39.0983	0.797197065	103978.47	39.0983	2.659411535	0	39.0983	0
Mg	20124.28	24.305	0.827989303	3985.48	24.305	0.163977782	83053.34	24.305	3.417129809
Mn	1364.41	54.938044	0.024835431	28824.33	54.938044	0.524669753	434.77	54.938044	0.007913824
Mo	0	95.95	0	0	95.95	0	0	95.95	0
Na	1265.5	22.98976928	0.055046224	1680.08	22.98976928	0.073079463	7064.84	22.98976928	0.307303649
Ni	48546.93	58.6934	0.827127582	15917.14	58.6934	0.271191309	0	58.6934	0
Р	276.46	30.973762	0.008925619	1339.59	30.973762	0.043249186	0	30.973762	0
Pb	0	207.2	0	0	207.2	0	0	207.2	0
S	0	32.06	0	0	32.06	0	0	32.06	0
Sc	0	44.955908	0	0	44.955908	0	0	44.955908	0
Si	292190.71	28.085	10.40379954	237585.58	28.085	8.459518604	233894.645	28.085	8.328098451
Sr	280.3	87.62	0.003199041	243.44	87.62	0.002778361	117.27	87.62	0.001338393
Ti	1984.35	47.867	0.041455491	20844.358	47.867	0.435464057	1772.72	47.867	0.037034282
V	0	50.9415	0	0	50.9415	0	0	50.9415	0
Y	0	88.90584	0	0	88.90584	0	0	88.90584	0
Zn	0	65.38	0	0	65.38	0	0	65.38	0
Zr	0	91.224	0	0	91.224	0	0	91.224	0

Table 3.7: ICP results for all three samples

Table 3.7 shows the ICP results. Things that can be noted from these are for sample M01 the large amount of Al and Si. However, other aluminum-silicate building elements seem not as abundant. Especially Na, which is a necessary ingredient for the albite seen in both the SEM and XRD is hardly measured.

For sample M17 there is a relatively large amount of Ti. Illmenite and sphene recognized in the thin section both contain titanium. There is also a significant amount of K, but there were no potassium-bearing minerals recognized in the thin section or by the XRD.

BM 5-1 has significantly more Ca than the other two samples according to the ICP, which is not at odds with the EDX maps and XRD results for the sample consisting mostly of diopside (a calc-silicate). There is also a relatively large amount of aluminum, while there is no aluminum in diopside.

3.5 Concluding mineral composition

Knowing what we know from the XRD results and SEM/microscope images, I calculated an estimate for the mineral composition of the samples with the ICP result.

Sample M01 is supposed to contain mostly albite (63% from XRD measurements; 60% phase 2 from SEM). However, the ICP results lack the necessary amount of sodium to create this amount of albite. I therefore decided that it may not be just albite present, but the plagioclase solid solution series in general. This however, still does not make a significant amount of plagioclase abundant in the sample (9%). The muscovite measured by the XRD (29%) could be represented by the red color of potassium in the EDX map. Here an amount of 20% was estimated. From the ICP measurements including the plagioclase, muscovite abundance is 32%, which is similar. Quartz is both in the SEM (phase 1 (yellow) – 10%) and the XRD (9%) estimated low in abundance, but in the ICP calculations, a large amount of Si was left, making the quartz percentage 45%. This does not seem likely. The three small phases (4, 5 and 6) recognized in

Mineral co Calculateo	mposition from ICP	M01 Cor Mineral Co	ncluded mposition		
Mineral	Percentage	Mineral Percentage			
Plagioclase	9%	Plagioclase	58%		
Muscovite	32%	Muscovite	25%		
Quartz	45%	Quartz	10%		
Rest	14%	Rest	7%		

Table 3.8: left: mineral composition as calculated from ICP results. Right: conclusions on mineral composition based on SEM, XRD and ICP results.

the SEM are likely similar to sphene and ilmenite, but since these make up only a small portion of the sample, this is not further considered. This 'rest' portion is estimated to be 10% from the SEM. The ICP calculations find a 'rest' percentage of 14%. Table 3.8 shows the calculations made from the ICP results. Considering the amount of plagioclase measured in both the SEM and XRD, I altered this calculation so the percentages are more in agreement with these measurements.

The microscope estimates for sample M17 say this sample consists mostly of hornblende (80%). The XRD results indeed show the presence of ferrotschermakite, which is a type of hornblende, but in lower abundance of 58%. The XRD also shows oligoclase (26%), which is a plagioclase which is also recognized in the microscope at 10%. Quartz is also present (16% from XRD, 5% estimated from microscope. Some smaller phases containing titanium (sphene, ilmenite) are recognized in the microscope, and this is in agreement with the titanium measured in the ICP. The amount of potassium measured by the ICP gives a problem in this mineral composition. There is no evidence for a K-bearing mineral. However, hornblende may contain K 'impurities' in its crystal lattice. The ICP measurements are calculated to a mineral composition in several ways, and a conclusion is drawn from these results. These can be found in table 3.9. In the second calculation, I generalized ferrotschermakite to hornblende, but this gave a lower abundance for the hornblende, and a larger 'rest' percentage. Also, the ilmenite decreased to 0.01% and the sphene increased to 9%, which is an overestimation. More than half of the 'rest' percentage is made up of the potassium. For the third calculation, I therefore included some of this potassium in the hornblende. This increases the abundance of hornblende and decreases the 'rest'. In all three of these calculations, I think quartz is overestimated, while oligoclase is clearly underestimated.

Sample composition for BM 5-1 is relatively easy. The main mineral is diopside. The XRD finds 100% diopside for BM 5-1. From the SEM images and the EDX map, we can see that this is not true. There is at

Mineral Compo ICP - 1	sition from 1		Mineral Composition from ICP - 2		Mineral Composition from ICP - 2		mposition CP - 2		Mineral Composition M17 from ICP - 3 Mineral		M17 Co Mineral Co	oncluded omposition
Mineral	Percentage		Mineral	Percentage		Mineral	Percentage		Mineral	Percentage		
Ferrotschermakite	52%		Hornblende	40%		Hornblende	48%		Hornblende	58%		
Oligoclase	3%		Oligoclase	4%		Oligoclase	4%		Oligoclase	13%		
Quartz	25%		Quartz	29%		Quartz	26%		Quartz	12%		
Ilmenite	3%		Ilmenite	<1%		Ilmenite	3%		Ilmenite	4%		
Sphene	4%		Sphene	9%		Sphene	4%		Sphene	3%		
Rest	13%		Rest	18%		Rest	15%		Rest	10%		

Table 3.9: left three: mineral composition for sample M17 as calculated from ICP results. Right: conclusions on mineral composition for sample M17 based on SEM, XRD and ICP results.

least one other phase present. There is also a third phase that can be seen in the EDX maps, containing calcium, most likely calcite. As this phase is a minor component, it is included in the 'rest'. As there were two peaks in the particle size analysis after the XRD mill, the second mineral is probably considerably softer (easier to crush) than diopside. According to the element maps, the phase should contain Al, Na, Ca and Si. A soft (1.5 vs. 5.5 on Mohs hardness scale) mineral that would fit the elements from the ICP is vermiculite. Vermiculite is linked to diopside by chrysotile. From the element map it seems pretty obvious that the second phase is a plagioclase solid solution series. However, plagioclase has a hardness similar to the hardness of diopside. The ICP calculations and conclusion with regards to the mineral composition are presented in table 3.10. Plagioclase is chosen as the second phase, as the diopside-plagioclase (especially diopside-anorthite) combination is commonly found.

Mineral Composition ICP - 1		Mineral Composition ICP - 2		BM 5-1 Concluded Mineral Composition	
Mineral	Percentage	Mineral	Percentage	Mineral	Percentage
Diopside	69%	Diopside	73%	Diopside	80%
Vermiculite	19%	Plagioclase	10%	Plagioclase	10%
Rest	12%	Rest	17%	Rest	10%

Table 3.10: left two: mineral composition for sample BM 5-1 as calculated from ICP results. Right: conclusions on mineral composition for sample BM 5-1 based on SEM, XRD and ICP results.

One needs to keep in mind that the mineral compositions as presented here are an estimate based on assumptions. These assumptions are based on actual data, but may the right not be conclusion for certain

samples. Especially for the second phase of the calc-silicate there is strong evidence for the presence of vermiculite (a phyllosilicate) rather than plagioclase.

4. Results

4.1 Displacement vs. Friction

The displacement vs. friction plots (figure 4.1 & 4.2) show the evolution of friction of the samples with displacement. The data from the three samples at the same temperatures are plotted in the same graphs.



Figure 4.1: Displacement vs. Friction plots at 100 MPa, between room temperature and 600°C, for all three samples: the felsic sample in red, the mafic sample in blue and the calc-silicate in green.

Figure 4.1 shows the different samples at all temperatures at 100 MPa. Figure 4.1a is the low temperature graph, showing the room temperature, 100° and 200°C experiments. The felsic sample is indicated in red, the mafic in blue and the calc-silicate sample in green. This graph shows that the calc-silicate is the stronger sample at low temperatures. All three samples show stick slips at 200°C at low velocities which fade out at higher velocities.

At 300°C & 400°C (figure 4.1b), the mafic and calcsilicate samples show almost the same value of friction. The felsic sample is weaker at 300°C, but increases in strength noticeably at 400°C, so that it comes very close to the mafic and calc-silicate. For the mafic sample, there are stick slips at 300°C at high velocities. The felsic sample has some instabilities at 10 μ m/s at 300°C, which include a sudden drop in friction. At 400°C, the felsic sample shows stick slips at high velocities. The calc-silicate seems stable at both temperatures, but shows a few isolated stick slips at 400°C at 1 μ m/s.

At 500°C & 600°C (figure 4.1c), the mafic sample is the strongest sample, and the felsic is clearly weaker. All three samples fail during the load up at 1 μ m/s at 500°C, but seem stable at higher velocities. Friction drops enormously at 600°C for the calc-silicate, but seems to increase again at higher velocities (while remaining stable). The felsic sample seems relatively unstable at 1 μ m/s at 600°C, but doesn't show regular stick slips.

For 200 MPa the graph at 500°C and 600°C is plotted in figure 4.2a. The felsic sample is the strongest sample, and is stable at all (measured) velocities at both 500°C and 600°C. The mafic

sample is again very steady at all velocities at 500°C, and values are around the same 0.7 as for 100 MPa runs. At 600°C, there is a sudden increase in this friction coefficient to 0.76. The sample is stable at both temperatures and all velocities. The calc-silicate shows the same surprising drop in friction at 600°C as could also be seen at 600°C at 100 MPa. The sample is noticeably stronger than at 100 MPa, but stable at all temperatures and velocities.

Figure 4.2b shows the result at 250 MPa for the felsic sample. It is almost identical to the graph at 200 MPa, only the friction coefficient is slightly lower.

The load up at 100 MPa has continuously increasing initial friction for all three samples in all the experiments. At 300°C, the felsic and the mafic sample both fail once during the load up. At 500°C, all three samples fail multiple times during the load up. The initial friction of the mafic sample increases faster at 300°C and 500°C than the other two, while at room temperature they increase about the same.

Re-loading at 100°C gives a small peak in friction, followed by a decay to a new steady state for all samples. At 200°C, initial friction of the mafic sample increases until it reaches a certain value after which it fails and stick slips form after that. For the felsic sample, initial friction increases gradually to a steady state value, and the stick slips start when this steady state is reached. The initial friction of the calc-silicate increases very fast, but does not reach a peak. It bends to a steady state value. When this is reached, stick slips start taking place. At 400°C, the mafic sample and the



MPa and 250 MPa, at 500°C and 600°C. The felsic sample is in red, the mafic in blue and the calc-silicate in green.

calc-silicate reach a peak in friction, which then decreases. The felsic sample seems to reach peak friction, but does not decrease after reaching this value. It seems like this value is already about steady state. For all samples, a strong peak friction with an afterwards decreasing friction is present at 600°C.

At 200 MPa, initial friction increases faster for the felsic sample, while the calc-silicate and mafic sample are about the same. At 500°C, friction increases to a steady state value for all three samples. Also for all three samples, the re-loading at 600°C shows a strong peak friction followed by a decrease to steady state.

The felsic sample at 250 MPa fails a couple of times during the load up at 500°C. Friction increases gradually to a steady state value. A peak friction with decay to steady state value is present at 600°C reload.

For all experiments, it seems like steady state friction was not yet reached after 3 mm of displacement at initial loading at room temperature, 300°C and 500°C.





Left: Figure 4.3: Displacement vs. Layer Thickness for the 100 MPa experiments. Up: Figure 4.4: Displacement vs. Layer Thickness for the 200 MPa and 250 MPa experiments.

4.2 Displacement vs. layer thickness

The evolution of the layer thickness with displacement during the experiments can be seen in figure 4.3 for the 100 MPa experiments and figure 4.4 for the 200 MPa and 250 MPa experiments. The instron position, which indicates the layer thickness, is normalized by dividing it by the instron position at the start of shearing. Now, layer thickness starts at 1 for all experiments, and layer thickness evolution can be compared. In figure 4.3 it shows that layer thickness decreases a lot more for the low temperature experiments. This is partially due to that the low temperature experiments have three runs. But even after two runs, all three experiments have decreased a lot more than the higher temperature experiments. For sample M17 (mafic), it looks like the layer thickness measurements are not accurate. For the first and second run, the layer thickness becomes a horizontal straight line at a certain thickness. The measurement may have been out of scale here. This is also the case for the calc-silicate at 600°C (second run) at both 100 MPa and 200 MPa. The amount of decrease in layer thickness increases still at first. In all experiments, most of the decrease happens in the first run, then the layer thickness increases because of the expansion of both samples and the pistons due to the temperature change (always increasing temperature). There does not seem to be a difference in evolution of layer thickness between samples.



4.3 Steady state friction vs. Temperature

Values of the steady state friction for the different samples at different temperatures and normal stresses are presented in table 4.1. The steady state friction values were picked after about 3 mm displacement at 1 μ m/s. The values in red are measured at peaks of stick slips, and they are noticeably higher than the rest. Experiments at 200 MPa and 250 MPa were only done at 500°C and 600°C because the depth at which these normal stresses are reached will not have lower temperatures. The 250 MPa experiment is only done for sample M01 (felsic), because there seemed to be a trend for this material to become stronger at higher stress. Because of the sensitivity of the sensors measuring the torque in the ring shear, these values have an estimated error of ± 0.0006 for the 100 MPa experiments, ± 0.0005 for the 200 MPa experiment.

T (°C)	100 MPa			200 MPa	200 MPa			
	M01	M17	BM 5-1	M01	M17	BM 5-1	M01	
RoomT	0.558	0.667	0.676					
100	0.585	0.705	0.765					
200	0.703	0.763	0.824					
300	0.607	0.714	0.699					
400	0.702	0.720	0.728					
500	0.505	0.723	0.644	0.780	0.707	0.727	0.722	
600	0.549	0.716	0.565	0.809	0.755	0.619	0.718	

Table 4.1: Steady state friction values

In figure 4.5, friction is plotted vs. temperature. At 500°C, three values are very close making the data points in the plot hard to see. This is M17 at 100 MPa, BM 5-1 at 200 MPa, and M01 at 250 MPa. At 600°C, M17 at 100 MPa and M01 at 250 MPa are again very close. For M17 (mafic), the friction coefficient seems to be relatively independent of pressure and temperature and is around 0.7 for all conditions. BM 5-1 (calc-silicate) friction increases rapidly at low temperatures from 0.68 to 0.82 between room temperature and 200°C, but then slowly decreases again to 0.56 at 600°C. At 200 MPa, the friction coefficient seems to be slightly higher than the same values at 500°C and 600°C at 100 MPa, but they still follow the same decreasing trend. Sample M01 (felsic) has a friction of around 0.55 for low temperatures, then increases to 0.6-0.7 between 200°C and 400°C, after which it decreases again. At 200 MPa, friction is considerably higher, up to 0.81 at 600°C. Therefore, this sample was also tested at 250 MPa to possibly find a trend with increasing normal stress. This however does not seem to be the case. The friction is still considerably higher than at 100 MPa, but not as high as the values for 200 MPa.

As mentioned before, the calc-silicate is the stronger sample at low temperatures. This gets taken over by the mafic sample from 300°C onwards. Both these samples seem to be relatively unaffected by higher stresses. The felsic sample on the other hand, is clearly the weaker sample at 100 MPa, but is much stronger at 200 MPa, both than itself at 100 MPa, and than the mafic and calc-silicate at both 100 MPa and 200 MPa (at 500°C - 600°C).

4.4 Velocity vs. (a-b) & (a-b) vs. Temperature

The values for a, b1, b2, b and (a-b) for the samples at different normal stresses and temperatures at all velocity steps are shown in the tables in Appendix A. The velocity that is indicated in the tables and graphs is the upstep velocity from the velocity step. Some velocity steps don't have values for a, b1, b2 and b. The values of (*a*-*b*) for these steps were calculated from the steady state solution ($(a - b) = d\mu_{ss}/\ln\left(\frac{v_1}{v_0}\right)$), because the data could not be fit by the model. The numbers in blue don't seem to fit with the rest of the data. For M01 (felsic) at 100 MPa, 200°C, the first velocity step, the sample shows stick slips, both before and after the velocity step. Calculations are done by hand, but with using the peak strength before and after, the obtained value seems less reliable because friction is velocity dependent and the exact velocity is unclear in a stick slip system. At 300° C, the second velocity step doesn't have a value for (a-b). This is because stick slips after the velocity step are very irregular and there is a very pronounced drop in friction here. Both modelling and calculation using the steady state equation don't work, because a steady state is not reached and/or clear evolution of friction is absent. The (a-b) values at 400°C are positive, while the material shows stick slips here. After the fourth velocity step at 500°C, sliding is unstable and drops, after which a state of stable sliding with a seemingly steady state value is reached. For sample M17 (mafic) at 100 MPa, 300°C, the last two velocity steps again show positive numbers where stick slips take place. For sample BM 5-1 at 100 MPa, 600°C, the first velocity step suddenly shows a negative value for b, making the (a-b) value much larger than the rest. What is interesting is that the same thing can be seen at 600°C at 200 MPa for this sample (calc-silicate).

The felsic sample (M01) shows an increase in *a* with velocity, temperature and normal stress. The mafic sample also shows an increase in *a* with temperature, while the calc-silicate does not. With increasing normal stress, *a* values for both the mafic and the calc-silicate samples don't show any particular trend. The mafic and calc-silicate also don't show trends with velocity, however it can be noted that the fourth velocity step $(30 - 100 \ \mu m/s)$ seems to generally have lower *a* values, especially at higher temperatures.

For the *b* values, there is no trend with velocity. An increase in *b* with normal stress can be seen for the felsic sample. For the temperature dependence, both the felsic and the mafic show an increase in *b* with increasing temperature.

There are two types of graphs. Figure 4.6 shows velocity vs. (*a-b*) plots for the seven different temperatures. The velocity plotted here is the upstep velocity of the velocity step. The velocity is plotted on a log scale to possibly find a (log-linear) trend in the (*a-b*) values. Figure 4.7 shows (*a-b*) plotted vs. temperature (which can be considered to be equivalent to depth, ignoring changes in stress).

The felsic sample at 100 MPa shows a clear decreasing trend in (*a-b*) with decreasing velocity at 100°C (figure 4.6), a trend can also be seen at 200°C and 500°C, but at these temperatures, (*a-b*) values are already negative. At room temperature, all values are positive, and at 600°C, they are all around zero. No trends can be recognized for the other temperatures. For all velocity steps, a parabolic trend is seen with depth/temperature, with a minimum at 200°C or 300°C. At 200 MPa, the felsic sample has a decreasing trend with decreasing velocity at 500°C. At 250 MPa, all (*a-b*) values are positive for the felsic sample, and they don't decrease with decreasing velocity. Generally (at all normal stresses), (*a-b*) values are more positive at 600°C than at 500°C. At higher temperatures (400°C - 500°C - 600°C), (*a-b*) for the first two velocity steps are lower (negative) than (*a-b*) at the last two velocity steps.





The mafic sample at 100 MPa shows a decreasing trend with decreasing velocity at 400°C, 500°C and 600°C, and possibly at 100°C as well (if you leave out the second velocity step). However, again most values are already negative. At 600°C, the decrease is not as rapid and values are around zero, like for the felsic sample. At room temperature, all values are again positive. Somewhat parabolic shaped trends can be seen with depth, this time with a very clear minimum at 200°C. What can be noted here is that at 300° C, (a-b) is generally bigger than at 400°C, making the shape not completely parabolic. At 200 MPa, (a-b) shows a decreasing trend with decreasing velocity for both temperatures: (a-b) values are more negative at the first two velocity steps.



Left: Figure 4.8: Displacement vs. Friction plots for the microstructural experiments. Red shows the stable sliding, and blue shows the stick slip experiment for each sample. Right: Figure 4.9: Displacement vs. Layer Thickness for the same microstructural experiments.

For the calc-silicate, only one possible trend can be recognized in the velocity vs. (*a-b*) plots (figure). (*a-b*) values decrease with decreasing velocity at 100°C, but values are already negative. At room temperature all (*a-b*) values are positive, while at 200°C and at 400°C, they are all negative. In the temperature vs. (*a-b*) plots (figure 4.7), the same trend can be seen as for the mafic sample: the (*a-b*) values are somewhat parabolic with a minimum at 200°C or 400°C. For the third velocity step, the minimum is at 300°C and the shape is perfectly parabolic. At 200 MPa, (*a-b*) values at 600°C are all positive, and at 500°C they are about zero. The first two velocity steps show more positive (*a-b*) values than the last two, opposite to the felsic and the mafic sample.

4.5 Microstructures

Several experiments were done to obtain microstructures for different sliding behaviors and study these under the light microscope and SEM. The results from these experiments are plotted as displacement vs. friction plots (figure 4.8) and displacement vs. layer thickness plots (figure 4.9). For each sample, one experiment shows stable sliding (in red), while the other one shows stick slips (in blue). For the felsic and the mafic sample, layer thickness gradually decreases in the stable sliding experiment, while for the stick slip experiments, layer thickness is more irregular and doesn't decrease as much. For the calc-silicate, layer thickness for both experiments gradually decreases.

Experiment	Sample	T (°C)	Behavior	Figures
U609	M01	roomT	Stable sliding	4.11 - 4.14
U581	M01	500	Stick slips	4.15 - 4.21
U608	M17	roomT	Stable sliding	4.22 – 4.24
U579	M17	500	Stick slips	4.25 – 4.27
U584	BM 5-1	100	Stable sliding	4.28 - 4.32 & 4.40
U575	BM 5-1	roomT	Stick slips	4.33 – 4.39
		• • •		

Table 4.2: Overview of microstructural experiments and their microscope pictures.

Table 4.2 shows an overview of the performed experiments, and the pictures taken of the thin section with the light microscope and SEM. For analyzing these pictures, the terminology and definitions as described by Logan et. al. (1992) are used. In this research, an idealized development of microstructures

in an experimental, monomineralic fault gouge is described. Figure 4.10 shows the fracture array as defined by Logan et. al. (1992). The average angles indicated are 15° for α , 53° for β , 72° for δ , and 16° for λ . At first, shearing is homogeneously spread over the entire sample. then, shear localizes to riedel shears, both R₁ at about 15° angle to the boundary and R₂ at about 70° angle to the boundary. After this, P fractures might initiate at an about 15° angle to the boundary, but in the opposite direction. Lastly, Y shears form, along which most of the shear takes place. They are parallel and close to the boundary.

To compare the array as described by Logan et. al. (1992) with the structures presented here, we need to keep in mind that the stress state within the ring shear is complex, so these structures may look slightly different. Also, the samples used here are multimineralic. In the pictures of the samples shown here, it is unclear if the sense of shear is dextral or sinistral. While the direction of shear within the ring shear is dextral, it depends on the direction from which you look at the sample if it is seen as dextral or sinistral. Looking from the outside in, the structure would appear dextral, while looking from the inside out, the structure would appear sinistral. This makes defining the fractures in the samples more difficult. However,

 R_1 shears are common features in the fracture array, while P shears are rare and hard to see if present. Therefore, I will name all fractures at an angle of about 15° R_1 fractures, unless mentioned otherwise. When looking at the pictures, this also means that most samples are sinistral, meaning we are looking at them from inside out.



Experiment u609 was the felsic sample showing stable sliding behavior. The first two pictures (figures 4.11 & 4.12) are microscope pictures taken with the gypsum plate inserted. In the upper half of the sample, figure 4.11 shows a localized zone in which grain size is reduced parallel to the boundary, which could be called a boundary shear. A few microcracks can also be recognized here. Figure 4.12 is a picture of the same area, but with

200 µm

the stage rotated to show uniform extinction. Grains in the surrounding matrix seem to show a preferred grain shape orientation at an angle 23° to the boundary.



The SEM pictures (figure 4.13 & 4.14) of the stable sliding felsic sample show an up-left to down-right striation, which is most likely due to polishing of the thin section rather than a feature in the sample. At the upside of the sample (figure 4.13), a crack can be seen parallel to the edge of the sample. This specific crack, however, doesn't seem to continue. An R₁ fracture can be seen running from down-left to up-right at an angle of about 15°. Figure 4.14 is a zoom of this crack and the grains surrounding it. The grains surrounding the crack are slightly smaller than further away from the crack.



Left: Figure 4.13: Experiment u609, felsic sample, stable sliding. Arrows indicate fractures in the sample. Right: Figure 4.14: Experiment u609, felsic sample, stable sliding. Smaller grain size near fractures.



Up left: Figure 4.15: Experiment u581, felsic sample, stick slips. Up right: Figure 4.17: Experiment u581, felsic sample, stick slips. This is zoomed in from figure 4.16 (below).





Up: Figure 4.18: Experiment u581, felsic sample, stick slips. Arrows in right picture indicate boundary parallel fracture and band with fine grained material. Picture on the left is zoomed in on the band of finer grain size.



Left: Figure 4.19: EDX map of sample u581, felsic, stick slips. Band of fine grained material is no longer visible, but muscovite grains are elongated as indicated by arrow. Right: Figure 4.20: Experiment u581, felsic sample, stick slips, arrows indicate fractures.



Figure 4.21: Experiment u581, felsic sample, stick slips. Elongated grains with bottom-left to top-right orientation.

The stick slip felsic sample (u581; figure 4.15 - 4.21) has fractures at the boundary of the sample as well. Besides the fractures having a darker color, the structure of the grains in the boundary shear region and the rest of the sample don't seem to be any different in figure 4.15. There is another fracture about half way in the sample that is parallel to the boundary. In figure 4.16, there are several fractures running parallel to the upper boundary, which is at an angle of about 20°. An enlargement of this area (figure 4.17), demonstrates larger grains (in blue) are oriented the same way (top to the right), while other grains, which are also generally smaller, are oriented more randomly. The direction of this rotation (top to the right) is the same as of the cracks.

The SEM picture on the left (figure 4.18a) shows a crack parallel to the boundary, with below a band of finer grained material. Figure 4.18b (on the right) is zoomed in on this band, clearly showing the difference in grain size and also the band being bordered by microcracks. An EDX map (figure 4.19) was made of this band. In this figure, Si is represented in red, Na in blue, and Al in green, all representing different phases in the sample (quartz, plagioclase and muscovite). The band is no longer visible, but in the area where the band is supposed to be, muscovite grains (in green) are elongated and tilted at a 27° angle (top to the right). Figure 4.20 shows multiple fractures parallel to the boundary, and one at an angle of about 10° (possible R_1 shear). There is also a small fracture with an orientation the other way branching off the angled crack (possible P shear). In figure 4.21a, elongated grains seem to be oriented from up-right to down-left. When zoomed in (figure 4.21b), this observation is confirmed.



Up: Figure 4.22: Experiment u608, mafic sample, stable sliding. Arrows indicate possible boundary shears. Below: Figure 4.23: Experiment u608, mafic sample, stable sliding. Riedel shear indicated by arrow. The right picture is zoomed in on the shear, showing a smaller grain size surrounding the fracture.





The quality of the thin section of sample u608 is not very good. It has many scratches due to preparation (polishing). Also, the surface is uneven: on one side the sample is too thin, while on the other side it's too thick for the light to go through and gives a dark image. Some structures are still recognizable. In figure 4.22, there are several fractures visible close to the boundary that are (almost) parallel to the boundary. Material in this region is darker (but this may be due to the quality of the thin section). On the left end, R₁ fractures start with a 14° angle to the boundary. There also seem to be some linear features from downleft to up-right at an 18° angle. Figure 4.23a shows a R₁ fracture (~ 13°-14°). When zoomed in on this feature (figure 4.23b), finer grained material is seen to surround it. An EDX map was made to find possible compositional differences (figure 4.24). While the crack is visible, no differences can be found between the finer grained band next to the fracture, and the material further away from the fracture.



Left up: Figure 4.25: Experiment u579, mafic sample, stick slips. Left down: Figure 4.26: Experiment u579, mafic sample, stick slips.

Experiment u579 was done on the mafic sample at 500°C and showed stick slips. Figure 4.25 shows a fracture parallel to the boundary. No real difference can be seen between grains in the boundary shear and the rest of the grains. Figure 4.26 shows two R_1 shears (at an approximate 15° angle to the boundary). The R_1 shear seen in the SEM picture (figure 4.27) has about the same angle. Grains seem to be uniformly spread and randomly orientated with no changes in grain size.



Figure 4.27: Experiment u579, mafic sample, stick slips.

The calc-silicate showed interesting features in its thin sections for both the stable sliding (u584) and the stick slip (u575) sample. They are best seen in PPL as patches of lighter colored material (figure 4.28, 4.29, 4.33 & 4.34). Almost all of them have a color change from darker to lighter. For sample u575 some more detailed pictured were taken in XPL (figure 4.35) and with the gypsum plate (figure 4.36). In XPL

it becomes clear that these patches consist of smaller grained material, and the color change from dark to light correlates with the grain size going from small to even smaller to even a glassy/amorphous looking matrix. The smaller the grain size, the more obvious some linear features become. They run from the boundary with the 'normal' matrix into the patch at almost 90° in the middle, while at the edges, these linear features seem to be leaning at a shallower angle towards the middle of the patch. The boundary between the 'normal' material and these patches is very sharp. Orientation of the patches is random, but they seem slightly elongated. The relationship between cracks and the patches in the sample is random as well: sometimes they follow the patch boundaries (but most boundaries do not have cracks), but they also cut through the patches, and sometimes they seem to avoid them. An EDX map was made and can be seen in figure 4.40. Here, Ca is in red, Al in blue, and Si in green. The patch is still visible, but it does not seem to be a difference in color. It seems like the smaller grain size makes for a slight contrast in brightness, which makes the patch visible.





Up: Figure 4.28: Experiment u584, calc-silicate, stable sliding. PPL, showing patches of lighter colored material. Down: Figure 4.29: Experiment u584, calc-silicate, stable sliding.



Up left (Figure 4.30), up right (Figure 4.31) and below (Figure 4.32): Experiment u584, calc-silicate, stable sliding.





Left: Figure 4.35: Experiment u575, calc-silicate, stick slips, in XPL. Below: Figure 4.36: Experiment u575, calcsilicate, stick slips, with inserted gypsum plate.





Left (Figure 4.37) & Right (Figure 4.38): Experiment u575, calc-silicate, stick slips

In the stable sliding calc-silicate (u584), the bigger grained matrix has randomly oriented grains which do not show any structures. Most fractures are R_1 at a 13-17° angle with the boundary. In the SEM pictures (figures 4.30, 4.31 & 4.32), these fractures are very obvious. Figure 4.31 shows a connecting fracture at 45° with the two R_1 fractures at 15° which it connects. It is also oriented the other way. This small fracture is at a patch boundary: on the left side of the fracture in, bigger, randomly orientated grains can be recognized. On the right side of the fracture, no grains can be recognized. The patches are easily identified in the SEM pictures as well.

Figure 4.33 of the stick slip calc-silicate (u575) seems to show a slight orientation or lineation from downleft to up-right. R_1 fractures are at about 15° angle (mainly figure 4.34) are irregular and mostly discontinuous. In XPL (figure 4.35), the 'normal' matrix grains do not show any features. Figure 4.39 shows a fracture not only from up-left to down-right, but also smaller ones from down-left to up-right at an about 20° angle.



Figure 4.39: Experiment u575, calc-silicate, stick slips.



Figure 4.40: EDX map of sample u575.

5. Discussion

5.1 General comments on friction

The load up of a run can show two different types of behaviors: (1) friction increases quasi-linearly followed by a gradual roll over a yield to a steady state value; or (2) friction increases rapidly to a peak friction, after which friction decreases to a steady state value. This difference is caused by a difference in "static friction", the maximum value of friction following a hold time (Marone, 1998). The ratio of shear stress over normal stress needs to exceed the static friction in order for sliding to begin (Scholz, 1998). Static friction is influenced by porosity: it will be higher for lower porosity. This is because when porosity is low, more work needs to be done against the normal stress in order to allow grains in the sample to move past and over each other. By pre pressing rings and targeting 15% porosity, it was hoped to come close to an equal starting porosity for all samples. However, as soon as the sample is in the pistons, several mechanisms play a role that alter the porosity before shearing starts. The normal stress applied in combination with fluid saturation will decrease the porosity. Porosity, and thus static friction, is timedependent: porosity decreases with increasing 'hold' time (Marone, 1998). The longer the normal stress is applied, the smaller the porosity will be, and the larger the static friction will be. Another factor that influences the starting porosity in the pistons is temperature. Elevated temperatures activate several deformation mechanisms such as pressure solution creep and dislocation creep (decreasing viscosity of the sample). These mechanisms cause a time-dependent decrease in porosity. Static friction thus becomes higher with increasing temperature. For the high temperature experiments, porosity is decreased by both effects: the temperature decreases the porosity, but it also takes a longer time under normal stress for the ring shear to reach and equilibrate at the experimental temperature. For our samples, at 100 MPa, initial friction gradually increases at all temperatures, and has no peak friction. However, at 500°C, all three samples fail multiple times during the load up, and the mafic and felsic sample also fail during loadup at 300°C (but only once). These stress drops automatically mean that there is a little bit of shear displacement. When a sample fails during the load up, it looks like it has overcome its static friction, but the strength keeps increasing after that. A possible explanation could be that interlocked grains break rather than move past each other. This would mean that the breakage of grains requires less work than dilation and frictional resistance. Relatively big (and weak) grains would break first, and the breakage in combination with the caused movement, cause lower porosity and interlocking of now stronger (smaller) grains. This would then increase the "static" friction. However, it has to be pointed out that it's unlikely that grain breakage would cause a macroscopic stress drop. Also, in rate-and-state friction, friction is something that is evolving rather than something that can be described by "static" and "dynamic" friction. What can be noted from the experiments is that the materials show more instabilities during the load up at elevated temperature.

For some samples, the slope of the load up is steeper. The slopes were calculated and are presented in table 5.1. The steepness has to do with the combination of stiffnesses of the machine and the sample (a higher stiffness gives a steeper slope), and is probably related to porosity. Samples with a higher compaction rate will have lower porosity, and thus a higher stiffness: friction increases faster (i.e. the slope is steeper). There is less displacement needed to achieve higher shear stress.

Normal stress	Т (°С)	Sample	Exp.	Slope				
100 MPa	RoomT	M01	u560	0.39				
		M17	u558	0.41				
		BM 5-1	u576	0.56				
	300C	M01	u559	0.39				
		M17	u556	0.73				
		BM 5-1	u571	0.55				
	500C	M01	u561	0.33				
		M17	u557	0.54				
		BM 5-1	u578	0.46				
200 MPa	500C	M01	u599	0.63				
		M17	u598	0.44				
		BM 5-1	u601	0.42				
250 MPa	500C	M01	u610	0.45				
Table 5.1: s experiment	Table 5.1: slopes of the load up for the experiments.							

For the second (and, in case of the low temperature experiments, third) run in the experiment, the sample has been under normal stress and temperature for a longer time. This, however, has not been a 'hold' time for long since there was shear displacement in the previous run. Reloading does give a peak friction with a gradual decrease in all cases. This means porosity was smaller, increasing the static friction. The displacement vs. friction plots (figure 4.1 & 4.2) show that the peak becomes more distinct (becomes higher) with temperature. To show this even more, the difference between the peak friction and the steady state value is plotted against temperature in figure 5.1. In this graph, the first and second point (100°C & 200°C) are all negative. The friction does show peaks here, but in case of the first point, the peak is very small. Friction drops after the peak, but then increases again to its steady state value. "Dynamic" friction here is higher than "static" friction, which would usually not be possible.

However, in the case of rate-and-state friction, "dynamic" and "static" friction are not really defined, and friction is seen as always evolving. This makes it possible for the steady state friction to be higher than the peak friction. For the second point, the difference is negative, because the "steady state" value here is defined by stick slips, giving them a slightly higher value. This will be discussed further on (in section 5.2).

Figure 5.1 demonstrates the temperature dependence of porosity and static friction. This is a clue that both the activation of plastic deformation mechanisms as well as an increased time under stress with increasing temperature play a role in these samples.

The evolution of the layer thickness may give clues about the porosity of the sample. If porosity decreases, the layer will become thinner. However, it is difficult to measure true changes in layer thickness, because the biggest influence on the evolution of layer thickness is the amount of material lost, which is something that can't be distinguished from layer compaction. It thus needs to be considered whether or not changes in porosity would be visible in the displacement

Figure 5.1: relationship between initial peak in friction and temperature.



vs. layer thickness plots presented in the results. In these plots, layer thickness evolution seems independent of normal stress and sample (composition). Since this trend happens for all the samples, I assume here that material loss is less of an issue, and layer thickness depends (almost) completely on porosity. There are differences in layer thickness with temperature: the decrease of layer thickness decreases with increasing temperature. This indicates that porosity decreases faster at lower temperatures during the experiment. Before the start of the experiment, samples are already being compacted by the normal stress and temperature on the sample. Compaction will be more at high temperatures, as mentioned above (porosity decreases faster). The porosity at the beginning of sliding will be smaller at high temperatures than at low temperatures. This means changes in layer thickness will be much smaller, which is the case here.



Figure 5.2 shows the decrease in thickness during the stick slips. The stick slips in this figure are from all three samples at 100 MPa and 200°C. It can clearly be seen that layer thickness decreases during the 'slip' phase: compaction takes place. This compaction continues for a little while after the sample has stopped sliding in the 'stick' phase. However, the sample shows (almost) no compaction during the 'stick' phase. While all three samples show this behavior, it is most prominent in the mafic sample.

All three samples show the highest friction at 200°C. This is because all three show stick slips at this temperature. The friction taken as "steady state" during the stick slips is the peak friction. This is the maximum strength of the material that the shear stress has had to overcome at the end of the "stick" phase, after which the material slides: the "slip" phase. During the "stick" phase, velocity on the sample is zero (or close to zero; He et. al., 2003; Paterson & Wong, 2005), while the machine rotates with a constant velocity of 1 µm/s (in this case). Both the strength of the material and the shear stress on the material build up, until the shear stress overcomes the strength and the material slips. During this slip, the material 'catches up' with the machine, covering the distance the machine has moved during both the stick and the slip phase. This means that the velocity on the sample is faster than the velocity of the machine. This influences the friction of

the sample, because friction is velocity dependent. As the material is velocity weakening, meaning friction decreases with increasing velocity, the actual friction of the sample is probably lower than the peak friction used, because velocity is very low at the end of the "stick" phase (and thus friction is high). At the end of the slip phase, the minima of the stick slip cycle, is probably caused by machine ("system") stiffness, as well as values of (*a-b*) and *d_c* (Paterson & Wong, 2005). Stick slips are usually more pronounced at low velocities. At higher velocities, less time is needed for the same displacement. This means the shear strength which needs to be overcome is reached sooner, leading to smaller stick slips. In most cases this is true. However, for the mafic sample at 300°C and the felsic sample at 400°C, stick slips are only present at high velocities, and they grow with increasing velocity. This will be discussed in more detail in section 5.2.

Slip Velocities								
M01 M17 BM 5-1								
1	1.28	4.45	2.15					
2	1.19	4.22	1.78					
3	1.41	3.69	1.33					
4	1.26	2.43	1.31					
Average	1.29	3.70	1.64					

Table 5.2: Slip velocities from stick slips at 200°C at 1 μm/s in μm/s for all samples. The slip velocity can be calculated by turning the elastically corrected displacement of the 'slip' phase into a velocity by dividing it by the time. Since the stick slips are not very regular, this velocity is different for each slip. From each sample, four slip phases were picked to calculate slip velocity. Table 5.2 shows these velocities, as

well as the average slip velocity. They are all from the 200°C stick slips at 1 μ m/s. These values are the maximum velocities during a slip phase. Figure 5.3 shows the graphs from the calculation. 5.3a is the regular displacement (as measured from the potentiometer) of the machine plotted against the shear stress. The elastically corrected displacement, which is calculated using the stiffness as measured from the "stick" phase, is plotted against shear stress in figure 5.3b, this is the displacement of the sample. To turn this into velocity, the slope of the time vs. sample displacement plot is calculated. When time is now plotted against this velocity, figure 5.3c is the result. It is plotted over the sample displacement vs. shear stress plot from figure 5.3b. The plots are slightly shifted because the x-axes are not exactly the same ("displacement" in 5.3b and "time"). The peak velocity is picked here, since this is the highest velocity the sample reached during the slip phase. From table 5.2 it can be seen that for all samples, slip velocity is higher than the velocity of the machine. The mafic sample slips a lot faster than the other two. The stiffness used for the calculation is also a lot higher than for the other two samples. Stiffness for the mafic sample was 176, while for the felsic sample it was 161 and for the calc-silicate 156.

5.2 General comments on a, b and (a-b)

The values of a and b are material properties used to describe the velocity dependence of friction in the rate and state friction theory. When the velocity is suddenly increased, this increases the strain rate. As a consequence, contacts have higher shear strength, and thus an instantaneous jump in friction occurs, called the direct effect ("a"). The value of a is dependent on the dominant deformation mechanism, which depends on the controlling mineral in the rock. This makes a also dependent on temperature and normal stress. Beeler et. al. (2007) links a to dislocation glide. Paterson & Wong (2005) point out that for quartz, a is sensitive to slip distance and velocity.



Figure 5.3: Calculating slip velocity (experiment u560 – felsic sample, 200°C at 1 μ m/s).

The evolution of friction after velocity is increased is described by the parameter "b", or the evolution effect. This value is (usually) positive and shows a decrease of friction with time after the direct effect for upward steps in velocity. It describes the evolution of contact area with time until a new steady state is reached. The critical slip distance d_c is the slip distance needed to replenish a new contact population. When phyllosilicates are involved, the value of b can be (near) zero. This indicates "contact saturation" (Saffer & Marone, 2003; Niemeijer & Collettini, 2013), in which the contact area is maximized because phyllosilicates align themselves with the shear direction.

The value of b may be composed of two values, b1 and b2. In this case, the value of b is represented by two deformation processes taking place at the same time but at different length scales. One (b1) is dominant right after the velocity step, and the other one (b2) takes over after a certain amount of displacement. In all the models that could be fitted to the data for all the samples, there were both b1 and b2values necessary.

The velocities tested here are high compared to natural fault velocities, i.e. the tested strain rate (10^{-4} s^{-1}) is higher than the actual strain rate in natural faults $(10^{-15} \text{ s}^{-1})$. When *(a-b)* values are seen to be decreasing with decreasing velocity

in these experiments, this might be interesting because these results might be extrapolated to produce negative (*a-b*) values to velocities (and strain rates) representative for the faults of the East African Rift on which earthquakes nucleate at depth. This decreasing (*a-b*) with decreasing velocity trend is seen at several temperatures and pressures, but often (*a-b*) values are already negative at the velocities tested. Since we're looking for unstable behavior at high pressures and temperatures, some of these trends are potentially interesting.

In most of the (*a-b*) vs. temperature plots, a parabolic trend can be recognized, showing aseismic – seismic – aseismic behavior of the samples with temperature (or depth). This trend is recognized by many authors (e.g. Chester & Higgs, 1992; Blanpied et. al, 1995; Chester, 1995; Den Hartog & Spiers, 2012). At shallow depths, faults are generally aseismic. This is because faults contain loose wear detritus (fault gouge). Such

poorly consolidated material involves additional hardening (involving dilatancy) when the material is sheared, which causes more positive (*a-b*) values (Marone & Scholz, 1988; Scholz, 1998). At depth, pressure and temperature solidify the gouge and decreases this effect and thus lowers (*a-b*), allowing for the aseismic – seismic transition. The second and deeper seismic – aseismic transition is usually associated with the onset of crystal plasticity for the dominant mineral in the material (often quartz). The depth at which this transition occurs depends on the geothermal gradient of the area (Sibson, 1982; Scholz, 1998). In the data presented here, at the first velocity step, all (*a-b*) values at 300°C are positive. They are negative again at 400°C, which makes for multiple transitions between seismic and aseismic. For the first two velocity steps, especially for the mafic sample, (*a-b*) values remain negative at high temperatures.

For the felsic sample at 400°C and the mafic sample at 300°C, there are only stick slips at high velocities. Also, the stick slips grow with increasing speed. This is opposite to what is expected from rate and state friction, because the higher the velocity, the faster the sample will reach its peak strength and slip, making stick slips smaller with increasing velocity. Since here stick slips grow with increasing velocity, (*a-b*) values, which are calculated from peak friction, are calculated as positive, while the sample is unstable. The sample velocity at which the samples start to slip at the different motor velocities could influence the peak friction, since friction is velocity dependent. For the felsic sample (at 400°C) at 30 μ m/s motor velocity, the velocity of the sample at peak friction (the start of the slip phase) is 3 μ m/s. At 100 μ m/s motor velocity at 30 μ m/s motor velocity, and 9.9 μ m/s sample velocity at 100 μ m/s motor velocity. This means that for both samples, peak friction also becomes higher with sample velocity, indicating the sample would still be velocity strengthening. There should be a different explanation why the sample shows stick slips at high velocities. Possibly trapped pore fluid due to low porosity may have something to do with this, changing effective normal stress and shear stress states within the sample.

5.3 General comments on microstructures

The cracks seen in the microstructures are dilatant fractures. When the stress is released from the sample, the sample expands, and it is likely to break on a weak plane. These fractures thus accentuate weak planes within the sample due to shear. The stress state within the ring shear is complex, so these structures may look slightly different from what Logan et. al. (1992) described. Also, Logan et. al. (1992) used a monomineralic gouge, while here we have multimineralic gouges. This means the stress field is altered to the weaker minerals in the sample, rather than on the entire sample. What can be said here about all the fractures at an angle to the boundary, is that the angle is around 15°. These are most likely R₁ (classic) riedel shears. R₁ shears begin to form in the post-yield region and with continued deformation link up with B or Y boundary-parallel shears (Marone, 1998). Both homogeneously distributed shear and slip localized on riedel shears is associated with velocity strengthening behavior (Logan et. al., 1992; Marone, 1998). This behavior (and velocity neutral) persists until a well-developed set of Y shears form. When these become prominent, stable frictional strength is reached (Marone, 1998). The development of these structures is a result of the development of the orientation of the stress field within the gouge (Logan et. al., 1992), and is in accordance with the development of the friction coefficient.

In general, it is seen that riedel shears are well developed in all samples, and boundary shears are more visible and better developed in the stick slip samples. This is in agreement with Logan et. al. (1992) and Marone (1998). Shear localization takes place and grain size is reduced, producing weak planes in the sample, on which the shear takes place. When the stress is removed from the sample and the sample expands, the sample fractures on these planes, making R₁, P and Y shears visible. Detailed images of these, show that surrounding these fractures, grain size is smaller than in the rest of the sample. EDX maps showed that there is no preferred mineral on these localized planes, so all minerals are fractured and present in the R₁, P and Y shears.

5.4 The felsic sample (M01)

Besides the high friction coefficient at 200°C due to stick slips, the felsic sample also has a high friction coefficient at 400°C. It shows some unstable behavior here, but no real stick slips. It makes sense for this sample to have its highest friction value at this temperature because this sample consists mainly of feldspar. According to Scholz (1988), feldspar becomes plastic at around 450°C, so above this temperature, the sample strength should decrease, which is exactly what it does. However, this temperature for the onset of plasticity (and plasticity itself) is strain rate dependent (Scholz, 2002), which means this transition temperature will not be accurate at the velocities tested here.

When the normal stress is increased to 200 MPa, the friction of the felsic sample (M01) increases noticeably, and becomes the strongest sample of these three. Also, initial friction increases faster at 200 MPa than for the other two samples (the slope of the load up is steeper, see table 5.1), so by increasing the normal stress, compaction rate of the felsic sample is higher than for the mafic or calc-silicate samples. The 250 MPa experiment was done to see if the trend of increasing normal stress would carry on. This is not the case. Purely frictional behavior is normal stress dependent, but at temperatures of 500°C and 600°C, it is unlikely that deformation is purely frictional. Shear stress becomes normal stress independent when plasticity starts to take over. This means (apparent) friction would decrease with increasing normal stress. Figure 5.4 shows the normal stress plotted against the shear stress. If deformation would be purely plastic, and shear stress would thus be completely normal stress independent, the slope would be horizontal. While this is not the case, the shear stress clearly bends from 200 MPa to 250 MPa normal stress. This indicates plastic deformation processes are becoming more important, since shear stress is less normal stress dependent. So, it is very likely that at 250 MPa (at 500°C and 600°C), plastic behavior is more dominant. This could be due to increasing normal stress, since increasing the normal stress means more energy would be lost when deforming by frictional deformation, and it may be easier to deform the sample by plastic deformation.

The direct effect *a* seems to increase with increasing velocity. This would be in accordance with the possibility of a trend in *a* with slip distance and speed for quartz, something Paterson & Wong (2005) discussed. Saffer & Marone (2003) describe the exact opposite trend (decrease in *a* with increasing velocity) from their velocity stepping experiments on clay-rich material (smectite and illite). Here, the felsic sample, containing mostly feldspar, does follow the trend described by Scholz (2002), who states

there is a strong temperature dependence of *a*. Following this trend, the material also shows an increase of *b* with temperature.



The values for (*a-b*) generally decrease with velocity for the felsic sample at 200 MPa at 500°C, meaning values may be negative at natural strain rates, and the material is potentially seismogenic. At 250 MPa, the sample fails a couple of times during the load up, but is not triggered to unstable behavior by the velocity steps, and (*a-b*) values are all positive. They become more positive with temperature, again indicating that plastic deformation processes could be becoming more important at these temperatures and normal stresses.

For the felsic sample, SEM pictures of the stable sliding sample show clearly developed R_1 shears, while these can't be seen for the stick slip sample. Also, the stable sliding sample shows a

much wider boundary shear. The boundary shear for the stick slip sample is more localized. In both cases, grains have a preferred orientation. For the stick slips sample, shear mostly took place on localized boundary shears. These boundary shears have increased strain rates due to localization: most of the shear takes place here. Stick slips were created at 500°C. At this temperature, the material is thus capable of localizing to a small area. Strain rate is increased high enough that the material creates unstable behavior. According to Logan et. al. (1992), shear takes place on the entire sample, and after about 2 mm of displacement, R₁ shears develop. The displacement now takes place on the R₁ shears. Shear that takes place here is usually velocity strengthening or neutral, and stick slips don't occur until after the fracture array has reached the steady state condition when Y-shears have developed (Logan et. al., 1992; Marone, 1998). This is in agreement with what we see here. Also, at room temperature, boundary shears are much wider and thus not yet fully developed. Localization of the material is less effective at room temperature.

5.5 The mafic sample (M17)

At 100 MPa and especially at higher temperatures, initial friction increases faster for the mafic sample than for the other two samples (figures 4.1 & 4.2; table 5.1). Compaction rate here is thus faster. The friction coefficient for the mafic sample is very stable (around 0.7) for all normal stresses and temperatures. This makes it the strongest sample at high temperatures at 100 MPa. As the friction coefficient does not change (decreases) with temperature, there may be no switch from frictional to (more) plastic behavior. This is probably because mafic material has a high viscous strength (Mackwell, 1998). This means it is easier for the material to deform by brittle behavior than by plastic behavior up to high temperatures. This is also the reason why mafic material was proposed to play a big role in earthquakes at great depths like in the East African Rift Zone (Fagereng, 2013; Albaric et. al., 2009;

Shudofsky et. al., 1987; Nyblade & Langston, 1995). However, the sample as tested here, mainly consists of hornblende, of which not much is known.

That friction hardly changes with temperature and that this may be due to the material remaining in the brittle regime, does not immediately explain earthquakes at depth. Another argument for the mafic material possibly causing these deep earthquakes, are the friction parameters *a* and *b*. The values for (*a*-*b*) at high temperatures (400°C, 500°C & 600°C) for both 100 MPa and 200 MPa all show a decreasing trend with decreasing velocity. Most of them are already negative at the velocities tested here. This velocity weakening behavior of mafic material at high temperatures and normal stresses may give another clue and be another argument for mafic material being (partially) responsible for deep earthquakes in the East African Rift Zone.

Like the felsic sample, the mafic sample also follow the trend of increasing *a* with temperature described by Scholz (2002). An increase in *b* with temperature can also be seen.

SEM and microscope pictures for the mafic samples show riedel shears for both the stable sliding and the stick slip sample. They are more developed in the stick slip sample. In the stick slip sample, boundary shears can also be recognized, again agreeing with Logan et. al. (1992) and Marone (1998). Localization is easier at high temperatures for the mafic sample.

5.6 The calc-silicate (BM 5-1)

The calc-silicate consists of mostly diopside, which is a clinopyroxene. Crystals are usually elongated. Friction increases with increasing temperature at low temperatures. At these low temperatures, the calc-silicate is also stronger than the felsic and mafic samples. Above 400°C, friction drops noticeably. This might be the onset of crystal plasticity for diopside, but no further information was found on this. For higher normal stress (200 MPa), friction is slightly higher.

The a, b, and (a-b) values for the calc-silicate seem completely random and follow no specific trend. The material dependent friction parameters are thus not affected by changes in temperature, normal stress or velocity. Surprisingly, it does have two negative b values, both at 600°C at 100 MPa and 200 MPa. While a value for b of zero or close to zero could be explained by contact saturation due to phyllosilicates, a negative b cannot be explained by contact area evolution. Negative values are usually seen for phyllosilicates, however, here they are seen in an almost pure diopside sample. A possibility is that the second phase in this sample is not plagioclase, but vermiculite, as discussed in the materials section. Vermiculite is a phyllosilicate and thus will line up with the shear direction. Now the hypotheses of Niemeijer & Collettini (2013) of why phyllosilicates may give negative b values might work. There is a fluid film between phyllosilicate layers along which the shear takes place (Niemeijer et. al., 2010; Moore & Lockner, 2004). This fluid film becomes thinner with displacement, increasing the strain rate within the film. This would cause an increase in friction. Friction here is thus rather a property of the fluid in between the phyllosilicate layer, than of the mineral itself. Several questions arise from this issue, assuming vermiculite is the second phase in the sample. Is vermiculite the controlling mineral in the calc-silicate sample? If so, only at 600°C, or at other temperatures as well? This may change the conclusion drawn for this sample, which will be discussed later. It is, however, more likely that vermiculite is not the second

phase (but plagioclase, like assumed), but in this case there has to be another explanation for the negative b values. The friction coefficient at 600° C is also noticeably lower than for the lower temperatures. This can probably be linked to the negative b values. A mineral transition seems unlikely, since both diopside and plagioclase are stable up to much higher temperatures. A negative b value means friction is increased after a velocity step. Following the equation $\mu = \tau/\sigma$, there are two possibilities to increase the strain rate and friction. (1) Increasing the shear stress on the sample; (2) decreasing the effective normal stress on the sample. An increased shear stress will be caused by a (locally) increase in strain rate. This means the sudden localization of shear to a narrow shear zone may have taken place. As presented in the result and will be discussed later, this material shows 'patches' of very small grain size. If grain size becomes very small, porosity strongly decreases. This may lead to the potential issue that pore fluid inside the sample becomes disconnected from the controlling outside reservoir of the ring shear. The actual pore fluid pressure within the sample may be higher than the measured fluid pressure, because fluid becomes 'trapped' within the sample because low porosity leads to low permeability. This increased fluid pressure decreases the effective normal stress ($\bar{\sigma}_n = \sigma_n - P_f$), which will increase the apparent friction. However, negative b values are only for the first and second velocity step. It is possible that increasing the velocity, 'opens' some pores to increase the permeability, which decreases the fluid pressure again.

There are very clear R₁ shears in the stable sliding sample, and even P shears can be seen in the SEM pictures. No shears parallel to the boundary can be seen. In the stick slip sample, there are parts that could be defined as Y shears, but they seem underdeveloped. R₁ shears are in discontinuous fragments, with no specific orientation. Both the stable sliding and stick slip sample show areas of small grain size material. These areas seem randomly spread throughout the sample and have no real orientation. Striations can be noticed within these areas. Because this feature is seen only in the diopside sample, and in both the stick slip and stable sliding samples, it looks like this is due to composition of the sample. Grain size seem only reduced in these areas.

For this calc-silicate/diopside sample, the friction parameters do not seem to be affected by temperature, normal stress or velocity. The material does show changes in material strength with temperature and normal stress. When looking at the microstructures, the extreme case of grain size reduction down to the size that the material looks amorphous, is possibly linked to this. Grain size can be reduced by dynamic recrystallization, which is important during dislocation creep. However, this is very unlikely at the temperatures tested here (room temperature and 100°C), because these are high temperature processes. Another way to reduce grain size is by cataclasis during shearing. Why the material only reduces grain size in certain areas is unclear. This is likely linked to shear and shear direction, and maybe has to do with small grains sticking together, and possibly a positive feedback effect of smaller grains becoming even smaller (shear taking place easier within areas of smaller grain size, reducing grain size even more).

5.7 Localization of shear stress on faults

The graphs for temperature vs. friction look different for each sample because a different mineral composition will lead to different deformation processes active in the sample. This is because when looking at the flow law, these variables are material properties and thus depending on the mineral. Also in rate and state friction the parameters a, b and d_c are material properties. In a certain sample, it is not

necessarily the dominating mineral that controls the behavior of a sample. For example, phyllosilicates are needed only in small amounts to control deformation within the sample (Niemeijer et. al., 2010). Besides the geometry of the structure, deformation within the sample also has a lot to do with localization and weakening processes, which are also mostly material dependent processes. If a specific zone is weakened, shear stress will localize to this zone, because it is easier to deform. Deformation will then weaken this zone even more, and further localization will take place. The strain rate on this localized zone is increased. The increased strain rates may lead to unexpected deformation processes. It is easier to maintain brittle behavior in a region of elevated strain rate, than to have ductile behavior. So, as Fagereng (2013) pointed out, to increase the strain rate and have potential earthquakes at depth, there has to be a material present that stays in the brittle regime at depths. This material would accumulate its strain on planes of weakness, while in the rest of the material, strain is so low that hardly any deformation takes place. With the increased strain rate at the localized zones, it would then be possible for deformation to remain in the brittle regime. For this to happen, we thus need material on which it is easy for shear localization to take place. When looking at the samples used here, there are several things to be said about this.

For the calc-silicate, if the composition of diopside/plagioclase is assumed, it seems hard for deformation to localize and to form planes of weakness. More energy seems to go into dynamic recrystallization of the grains, which seems to be in random patches rather than oriented zones. There is hardly any evidence that localization has taken place within these samples. This is also an argument against the composition diopside/vermiculite. Vermiculite, as a phyllosilicate, would easily form a localized zone, which are hardly found within the sample. Also, if vermiculite were to be the second phase, this would be a weak dominant mineral. Friction is not particularly low as one would expect for a phyllosilicate. It is extremely unlikely that under the conditions at depth of the earthquakes in the East African Rift Zone, vermiculite would be the dominating mineral in terms of deformation, because this would probably not lead to seismogenic behavior.

In the felsic sample, grains are oriented in the shear direction, and localization is not uncommon. However, the microstructures seem underdeveloped, especially at low temperatures. At higher temperatures, boundary shear zones seem to localize more easily, leading to velocity weakening behavior and stick slips. However, when increasing the normal stress to 250 MPa, friction relatively decreases and it looks like deformation becomes more dominated by plastic behavior, leading to velocity strengthening and stable sliding. The sample is feldspar dominated, which behaves plastic from about 450°C on (at natural strain rates). Also, viscous strength of felsic material is not as high as is needed to keep the material in the brittle field up to large depths. The comparison to mafic material is made in a strength profile by Fagereng (2013) and can be seen in figure 1.5 (from the Introduction). Even though the sample seems to localize relatively easily, minerals within the sample would have to have a high viscous strength, which is not the case.

Comparing this, the minerals in mafic material have a high viscous strength, thus maintaining brittle behavior up to high temperatures. This can be linked to the data here, which shows that the friction coefficient for the mafic sample hardly changes with temperature and normal stress. Localization is easier at high temperatures, which leads to velocity weakening behavior above 400°C. Mafic rock, not only deforms brittle up to great depths, it is also likely to form localized zones at high temperatures and normal

stress. Frictional properties of the sample tested here also shows that mafic rocks are possibly seismogenic at great depths.

6. Conclusions

Three samples from the Malawi rift zone were sheared at various temperatures and normal stresses. The results may give clues to explaining earthquakes at great depths (35 – 45 km) in the East African Rift Zone, an extensional setting.

- A diopside dominated sample (calc-silicate) has no trends in the friction parameters *a* and *b*, but friction itself did seem to respond to changes in temperature. Microstructural analysis shows patches of small grain size material. These are likely a result of shear, however the exact processes taking place here are unclear.
- For the calc-silicate sample, there are negative values for *b*, while there is assumed that there are no phyllosilicates present in the sample. This may be due to the extremely small grain sizes, in combination with increasing the load up velocity. This could suddenly decrease the porosity, which decreases the permeability, disconnecting the pore fluid in the sample from the externally controlled pore fluid reservoir. Pore fluid pressures within the sample could increase. This decreases the effective normal stress, which increases the friction, leading to negative *b* values.
- The felsic sample shows a decrease in friction from 400°C with increasing temperature. When the normal stress is increased to 200 MPa, friction increases noticeably, but at 250°C this does not carry on as a trend. This could be due to shear stress becoming more normal stress independent when crystal plasticity becomes dominant at higher normal stress. There could be a change in deformation process between 200 MPa and 250 MPa, because at higher normal stress it becomes more difficult to deform by frictional deformation, and it may be easier to deform by plastic deformation processes.
- In the felsic sample, it is very obvious that when stick slips are present, localization features within the microstructures are better developed. Boundary shears are present and more localized than in the stable sliding experiment. This can also be seen in the mafic sample.
- The mafic sample has a friction coefficient of about 0.7 which is relatively independent of temperature and normal stress changes. This could possibly give a clue to there being no change from brittle to ductile behavior within this rock. This would be in agreement with earlier statements by Mackwell (1998) and others of high viscous strength in mafic rock.
- Velocity weakening behavior (or trends of decreasing (*a-b*) with decreasing velocity) are especially
 noticeable for the mafic sample at high temperatures (> 400°C). This means the mafic rock could
 be potentially seismogenic at high temperatures (and thus greater depths).

The microstructures show that for both the felsic and mafic sample, localization takes place. This seems easier at higher temperatures. Especially for the mafic rock, this is interesting. The combination of high

viscous strength, negative (*a*-*b*) values and strong localization could be part of the explanation of why deep earthquakes take place in a zone of crustal thinning and elevated geotherms.

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Appendix A – Values for a, b1, b2, b and a-b

The tables below show the values of a, b1, b2, b and a-b for the velocity steps. They are either modelled with a rate and state friction program, or calculated by hand. In case the steps were calculated by hand, they only show values for a-b.

M01 – 100 MPa	

T = Roo	T = RoomT										
V step	Velocity	а	b1	b2	b	(a-b)					
1	3	0.012864	0.007658	0.00339	0.011048	0.001815					
2	10	0.014187	0.008845	0.004449	0.013294	0.000892					
3	30	0.016425	0.007155	0.006046	0.013201	0.003224					
4	100	0.018059	0.009714	0.003692	0.013406	0.004653					

T = 100 °	T = 100°C									
V step	Velocity	а	b1	b2	b	(a-b)				
1	3	0.014771	0.01053	0.005294	0.015824	-0.00105				
2	10	0.017527	0.011699	0.005553	0.017252	0.000275				
3	30	0.018862	0.012704	0.005082	0.017786	0.001075				
4	100	0.021178	0.010336	0.005836	0.016172	0.005006				

T = 200°	T = 200°C										
V step	Velocity	а	b1	b2	b	(a-b)					
1	3					-0.02445					
2	10					-0.01725					
3	30	0.016748	0.017953	0.007395	0.025348	-0.0086					
4	100	0.021345	0.02036	0.002579	0.022939	-0.00159					

T = 300°	T = 300°C										
V step	Velocity	а	b1	b2	b	(a-b)					
1	3					0.0034					
2	10										
3	30					-0.01567					
4	100	0.027671	0.025369	0.004552	0.029921	-0.00225					

T = 400°	T = 400°C										
V step	Velocity	а	b1	b2	b	(a-b)					
1	3					-0.0082					
2	10					-0.00346					
3	30					0.00818					
4	100					0.00824					

T = 500°C									
V step	Velocity	а	b1	b2	b	(a-b)			
1	3					-0.00893			
2	10					-0.00603			
3	30					-0.00464			
4	100					-0.05456			

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.058762	0.053713	0.007526	0.061239	-0.00248		
2	10					-0.00019		
3	30					0.00118		
4	100					0.00069		

M01 – 200 MPa

T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3					-0.00449		
2	10	0.032315	0.0281	0.007749	0.035849	-0.00353		
3	30	0.049359	0.043719	0.00617	0.049889	-0.00053		
4	100					-0.00146		

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.039937	0.052662	-0.02782	0.024842	0.015098		
2	10	0.050699	0.043801	0.01544	0.059241	-0.00854		
3	30	0.051216	0.04534	-0.00339	0.041947	0.009268		
4	100	0.073979	0.058269	0.01305	0.071319	0.00266		

M01 – 250 MPa

	L							
T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.05348	0.045478	0.007217	0.052695	0.000785		
2	10	0.039827	0.030674	0.007723	0.038397	0.001431		
3	30	0.038797	0.02938	0.007241	0.036621	0.002176		
4	100	0.044076	0.035269	0.007319	0.042588	0.001488		

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.049138	0.03482	-0.00395	0.030866	0.018273		
2	10	0.086737	0.052784	0.02043	0.073214	0.013523		
3	30	0.116742	0.080181	0.02057	0.100751	0.015992		
4	100	0.091078	0.074733	0.01157	0.086303	0.004772		

M17 – 100 MPa

T = RoomT								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.022836	0.014289	0.003201	0.01749	0.005345		
2	10	0.014701	0.008128	0.003798	0.011926	0.002775		
3	30	0.015675	0.008297	0.004456	0.012753	0.002922		
4	100	0.022662	0.013962	0.005836	0.019798	0.002865		

T = 100°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.012818	0.010839	0.004001	0.01484	-0.00202		
2	10	0.014414	0.012296	0.006563	0.018859	-0.00445		
3	30	0.021118	0.015439	0.004733	0.020172	0.000946		
4	100	0.041353	0.034903	0.004479	0.039382	0.00197		

T = 200°C							
V step	Velocity	а	b1	b2	b	(a-b)	
1	3					-0.00296	
2	10					-0.01136	
3	30	0.016	0.0185	0.017	0.0355	-0.0195	
4	100	0.012208	0.013167	0.004868	0.018035	-0.00583	

T = 300°C							
V step	Velocity	а	b1	b2	b	(a-b)	
1	3					0.00571	
2	10					-0.00065	
3	30					0.01125	
4	100					0.00131	

T = 400°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.023343	0.025711	0.007345	0.033056	-0.00971		
2	10	0.038336	0.033409	0.01021	0.043619	-0.00528		
3	30	0.055026	0.049008	0.006683	0.055691	-0.00067		
4	100	0.040997	0.041783	-0.00043	0.041355	-0.00036		

T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3					-0.01059		
2	10	0.043033	0.033963	0.01296	0.046923	-0.00389		
3	30	0.033081	0.023322	0.009403	0.032725	0.000355		
4	100	0.037906	0.025131	0.008947	0.034078	0.003827		

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.04627	0.044992	0.006363	0.051355	-0.00509		
2	10	0.067742	0.055859	0.01287	0.068729	-0.00098		
3	30	0.060006	0.048424	0.01029	0.058714	0.001293		
4	100	0.056886	0.046598	0.006251	0.052849	0.004037		

M17 – 200 MPa

T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.045324	0.038955	0.01134	0.050295	-0.00497		
2	10	0.02789	0.022538	0.00856	0.031098	-0.00321		
3	30	0.02363	0.01382	0.0071	0.02092	0.00271		
4	100					0.00155		

T = 600°C							
V step	Velocity	а	b1	b2	b	(a-b)	
1	3	0.03536	0.031488	0.004667	0.036155	-0.0008	
2	10	0.040153	0.027716	0.009908	0.037624	0.002529	
3	30	0.04581	0.034601	0.007671	0.042272	0.003538	
4	100	0.035843	0.027368	0.003024	0.030392	0.005452	

BM 5-1 – 100 MPa

T = RoomT								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.014016	0.004908	0.000546	0.005454	0.008562		
2	10	0.014575	0.018177	-0.00941	0.008768	0.005807		
3	30	0.016211	0.007186	0.004383	0.011569	0.004642		
4	100	0.018135	0.006627	0.007212	0.013839	0.004295		

T = 100°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.015598	0.010406	0.007873	0.018279	-0.00268		
2	10	0.015423	0.010262	0.008672	0.018934	-0.00351		
3	30	0.016927	0.010451	0.007725	0.018176	-0.00125		
4	100	0.019443	0.01205	0.008067	0.020117	-0.00068		

T = 200°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3					-0.00623		
2	10					-0.01274		
3	30	0.021311	0.022409	0.004627	0.027036	-0.00573		
4	100	0.014535	0.014742	0.006329	0.021071	-0.00654		

T = 300°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3					0.00191		
2	10					5.00E-05		
3	30	0.023691	0.024525	0.005805	0.03033	-0.00664		
4	100	0.02129	0.020812	0.004874	0.025686	-0.0044		

T = 400°C							
V step	Velocity	а	b1	b2	b	(a-b)	
1	3	0.01545	0.017795	0.004815	0.02261	-0.00716	
2	10	0.020864	0.020663	0.009739	0.030402	-0.00954	
3	30	0.017128	0.017636	0.004841	0.022477	-0.00535	
4	100	0.012214	0.012216	0.008462	0.020678	-0.00846	

T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3					0.00146		
2	10	0.029457	0.024284	0.005469	0.029753	-0.0003		
3	30	0.035356	0.031827	0.004544	0.036371	-0.00102		
4	100	0.017334	0.013721	0.004362	0.018083	-0.00075		

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.020432	0.005858	-0.00714	-0.00128	0.021712		
2	10	0.016561	0.008098	0.004395	0.012493	0.004069		
3	30	0.016095	0.012396	-0.00188	0.01052	0.005576		
4	100	0.0149	0.012246	-0.00262	0.00963	0.005269		

BM 5-1 – 200 MPa

T = 500°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.024701	0.01905	0.005138	0.024188	0.000513		
2	10	0.020465	0.014825	0.003829	0.018654	0.001811		
3	30	0.01471	0.012789	0.001987	0.014776	-6.60E-05		
4	100	0.014501	0.011539	0.004537	0.016076	-0.00158		

T = 600°C								
V step	Velocity	а	b1	b2	b	(a-b)		
1	3	0.006381	0.01009	-0.01333	-0.00324	0.009616		
2	10	0.008154	0.004776	-0.01365	-0.00887	0.017026		
3	30	0.008843	0.004167	0.000903	0.00507	0.003774		
4	100	0.010974	0.006281	0.002348	0.008629	0.002345		