Viscoelastic behaviour of basaltic lavas

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Abstract

The rheological properties of basaltic lavas from Etna, Hawai'i and Vesuvius have been measured at temperatures between ~500 and 1150 °C. The viscoelastic response of the lavas was analysed using small forced oscillatory torques (<10^-3 N m) at frequencies between 0.002 and 20 Hz. A purely viscous regime was only approached during experiments with Hawai'i samples. These experiments indicated that at temperatures between ~1070 and 1130 °C, strain rate independent viscosities (>10^9 Pa s) could be measured at strain rates less than ~10^-2 to 10^-1 s^-1. At 800 °C, temporal variations in complex shear modulus and internal friction suggest that, over durations of up to 120 hours, structural adjustments were occurring within some of the samples. This time-varying behaviour of lava samples is attributed to the slow closing (healing) of microcracks resulting in the apparent stiffening of lava samples under annealing. Thus, those parts of lava flows that undergoing slow cooling have more elastic properties. Regions which cool faster possess smaller shear moduli and higher internal friction due to thermal microcracking.

Key words: basalt lava, Etna, Vesuvius, Hawai'i, shear modulus, shear viscosity, oscillatory rheology

1. Introduction

The rheological properties of basaltic lava are critical parameters in determining the advance rates, morphology and final dimensions of basaltic flows. Rheology is controlled by a number of factors, the most important of which are composition, temperature, crystallinity and bubble content. The effects of these factors on the viscosity of lavas between eruption and solidification (at which lavas become viscoelastic) have been analysed by previous workers (Shaw et al., 1968; Sparks and Pinkerton, 1978; Marsh, 1981; Ryerson et al., 1988; Pinkerton and Stevenson, 1992; Pinkerton and Norton, 1995; Richet et al., 1996; Dingwell et al. 1998).
At lower temperatures lavas behave as viscoelastic materials (Sakuma, 1953; Bagdassarov, 2000). Field measurements carried out at high stresses (~$10^3$ Pa) and typical eruption temperatures indicate that basaltic lavas have a rather moderate viscosity (> $10^2$ Pa s) and an appreciable static yield-strength (~$10^2$ to $10^3$ Pa) (Shaw et al., 1968; Pinkerton and Sparks, 1978; Pinkerton and Norton, 1995; Norton and Pinkerton, 1997).

The stiffness of solidifying lava, the temperature range over which solidification and the anelastic transformation occurs and the kinetics of fracturing and fracture healing processes are important parameters for lava flow modelling. Here, we present the results of viscoelastic measurements carried out on samples of basalt lava flows taken from Etna, Hawai’i and Vesuvius at temperatures in the range ~500 to 1150 °C. These reveal the importance of cooling generated microcracks on the strength of lavas.

Experiments designed to characterise the high temperature anelastic and viscoelastic behaviours of glassy, crystalline and partially molten rocks are based on measurements of elastic modulus and internal friction. Traditionally, these measurements are performed using an inverted torsional pendulum (e.g. Day and Rindone, 1961; Gueguen et al., 1981; Weiner et al., 1987; Versteeg and Kohlstedt, 1994) or by forced torsional oscillation (Beckhemer et al., 1982; Jackson and Paterson, 1987; Bagdassarov and Dingwell, 1993; Gribb and Cooper, 1998). The forced torsional oscillation method used in this work has previously been used to study the behaviour of both rocks (Berckhemer et al., 1982; Jackson and Paterson 1987; Gribb and Cooper, 1998; Bagdassarov and Dorfman, 1998) and glasses (Bagdassarov and Dingwell, 1993; Bagdassarov et al., 1993,1994). This technique allows the magnitude of the complex shear modulus and angle of internal friction to be measured over a range of temperatures and frequencies.
2. Apparatus

The experimental method consists of exerting small strain oscillatory torsion deformation on cylindrical samples. The equipment (described in detail previously (Berckhemer et al., 1982; Kampfmann and Berckhemer, 1985, Bagdassarov and Dingwell, 1993)) exerts a small sinusoidal torque (of amplitude ~10^{-3} N m) to the end of a cylindrical sample (8 mm in diameter, ~20 to 30 mm in length). A simple schematic of the device is shown in Fig. 1a. The harmonic torque applied to the sample is generated using a pair of electromagnets (two microphone-type coils) connected to a synthesiser via a power amplifier. The sample is fixed between two aligned alumina rods, onto which two sets of light aluminium wings are also attached. The angular deformation across the sample is measured by pairs of capacitive pick-ups which respond to the movement of pure iron plates located at the ends of the aluminium wings. The capacitive signal is detected and amplified using a 5 kHz-frequency bridge which is sampled using a PC. Calibration of the equipment has been described previously (Bagdassarov, 2000), with shear modulus measurements being accurate to 3 to 5 % (due to thermal drift of the calibration at high temperatures).

Although the mechanical design of the equipment has not changed from that used during previous work, the data acquisition hardware and processing software have been significantly improved. For each measurement, data are collected over two periods of the torsional oscillation (Fig. 1b). Data are sampled at up to 10 kHz, allowing 1000 samples per channel to be acquired at the highest frequency used during experiments (20 Hz). At torsional oscillation frequencies of 2 Hz or lower the number of samples per channel is limited to 10,000. Sinusoids are automatically fitted to the collected data using a Levenberg-Marquardt algorithm, and the shear modulus and phase difference between the applied torque and the angular displacement across the sample are calculated from the phase and amplitude parameters of the fitted curves.
Experiments were carried out over the frequency range 0.002 to 20 Hz (at approximately 0.3 log intervals) and at temperatures between ~500 and 1150 °C. At high temperatures, low frequency measurements were prevented by the onset of non-linear sample response. This was revealed by Fourier analysis of the data indicating the presence of harmonics of the torsional driving frequency. Automation of sampling and processing allows repeated measurements and at high frequencies an average of twenty measurements was used for most temperature-frequency points. At low frequencies, individual measurements take up to 17 minutes, so fewer experiments were averaged.

During experiments the furnace was purged with a flow of Ar gas (5 cm³ s⁻¹). Inspection of the samples after experiments showed that oxidation (as indicated by growth of magnetite crystals) had taken place only on the surfaces of the samples. Temperatures were recorded using Pt-PtRd (S-type) or Ni-NiCr (K-type) thermocouples. Direct measurements of the temperature field inside the furnace indicated that temperatures were reduced by up to 15 °C at distances of 10 mm from the hottest point. Although this spatial sensitivity implies that recorded values were only accurate to ~15 °C as indicators of the sample temperature, relative temperature changes within any one experiment are much better constrained (±3 °C at 1000 °C).

The data collected at each temperature-frequency point allowed calculation of the magnitude of the complex shear modulus \( G^*(\omega T) \), and the phase shift \( \varphi(\omega T) \) between the applied torque and the resultant angular strain of the sample, where \( \omega \) is the angular velocity (equal to \( 2\pi \) multiplied by the applied frequency). From these results, the real, \( G' \), and the imaginary, \( G'' \), parts of the complex shear modulus, the internal friction, \( Q^{-1} \), and the complex shear viscosity, \( \mu \), can be calculated from

\[
G^* = G' + iG'' = G' \cos(\varphi) + iG'' \sin(\varphi) \quad (1),
\]
\[ Q^{-1} = \tan(\varphi) = \frac{G''}{G'} \quad (2), \]

\[ \mu = \mu' - i\mu'' = \frac{G''}{\omega} + i\frac{G'}{\omega} \quad (3), \]

(e.g. Nowick and Berry, 1972). The zero-rate shear viscosity (macroscopic viscosity), \( \mu_0 \), may be obtained from the frequency dependence of \( G'' \) by

\[ \mu_0 = \lim_{\omega \to 0} \frac{G''(\omega)}{\omega} \quad (4), \]

(Marin, 1998). If these parameters are frequency dependent, then the material is viscoelastic and a more complex relation must be found for stress and strain (e.g. Bagdassarov and Dingwell, 1993).

1. Sample bonding

During experiments it is essential that each end of the sample is securely bonded to the alumina rods. In order to do this efficiently small conical grips (angle \( \sim 1^\circ \), length 4 mm) were machined at both flat ends of the sample with a diamond tool. Complementary mating grips were produced in the alumina rods and samples were cemented between the rods with a high temperature cement (Polytec). The assembly was placed in the torsion apparatus and the sample was then sintered to the rods for 2 hours at 150 °C and then for 24 hours at 500 °C, under an axial load of \( \sim 8 \) N (e.g. Berckhemer et al., 1982). Measurements carried out using a dummy sample of \( \text{Al}_2\text{O}_3 \) demonstrated that the effect of the cement on phase delay measurements was less than \( 10^{-3} \) rad. (Bagdassarov and Dorfman, 1998).
2.1 Size and shape factors

When temperatures increased during experiments, thermal expansion of the sample and the alumina ceramic rods was accommodated by a spring located at one end of the apparatus. At temperatures sufficiently high for the sample to deform, some of the accumulated stress dissipates by flow shortening of the sample. Changes in sample length were calculated from micrometer readings taking at the spring (to a precision of ~0.02 mm) and corresponding changes in the sample diameter (calculated by assuming conservation of sample volume) were then used to calculate the material properties. However, samples recovered after experiments have shown that flow deformation was not continuously distributed through the samples but was concentrated in the centre, producing distorted, barrel-shaped, cylinders. This is a consequence of temperature gradients across the sample and due to it being supported at both ends. Thus, despite efforts to account for changes in the sample shape, the deviation from a cylindrical form introduced errors (<2 %) in the assumed diameter of the sample, once sample shortening has started.

2.2 Sample description

The experiments were carried out on samples of basaltic lavas collected from three different volcanoes, Etna, Hawai'i and Vesuvius.

a. Etna. Two samples from Etna were collected from lava erupted in 1992 which had ponded after overflowing from a skylight in the Valle del Bove. One of these samples was from the top surface and one was taken from the base (~10 cm down from the surface) and thus represent samples with different cooling regimes. The surface sample has smaller vesicles (~0.2 to 0.5 mm, 15 to 20 vol.%) and smaller crystal content (~20 to 30 vol.%) than the basal sample (~20 vol.% of 1 to 2 mm vesicles and ~30 vol.% phenocrysts). Images of polished
sections of these samples are shown in Fig. 2. A further sample was collected from near the south east cone in 1999. This sample was taken from the least vesicular area found in a recently emplaced flow near hornito H3 (Calvari and Pinkerton, 2001) at the top of the active flow field.

b. Hawai'i. The Hawaiian basalt was sampled in the east rift eruption zone of Kilauea, Hawai'i from a lava flow from a pahoehoe toe in September, 1984 (eruption temperature \~1147 °C). This pahoehoe lava flow corresponds to the episode 25 of the eruption Pu'u 'O'o of Kilauea Volcano. Its bulk chemical composition has been presented elsewhere (Garcia et al., 1992). The chemical composition of the groundmass glass obtained by microprobe analysis differs from the bulk rock composition in SiO\(_2\) (52.3 wt\%) and CaO (9.2%) content (Bagdassarov, 2000). The vesicularity estimated from 2-D image analysis varies from 46.8 to 57.6 vol.\%, or \~50 vol.\% when calculated on the “dry” density rock basis. The sample has about \~10 vol.\% of olivine quenched from magma during sampling and a few percent of other phenocrysts.

c. Vesuvius. The Vesuvius samples were collected from the 1834 flow at Cava Ranieri in the national protected area of Terzigno approximately 6.3 km ESE of the central cone of Vesuvius by the group from University College of London. Chemical analysis of the samples is given in Belkin et al. (1993). The same sample was also used by Rocchi et al. (2002) in experiments to determine Young's modulus and tensile strength.

3. Results

The shear modulus and internal friction results for the samples are given in Fig. 3 to 6. The 1992 Etna samples (Fig. 3) show marked differences in shear modulus and internal friction between the rapidly cooled crustal sample and the more slowly cooled basal sample. As a result of its smaller and smaller volume fraction of vesicles, the crustal sample has a shear modulus \~1.5 to 2 times greater than that of the basal sample at high temperatures (>1050 °C),
increasing to ~6 times at low temperatures (~500 °C). The internal friction of the crustal sample shows a relatively low dependency on frequency at high temperatures (>800 °C). At temperatures between ~600 and 800 °C (Fig. 3c) a small, wide peak in internal friction suggests a complicated behaviour at temperatures well below that at which the matrix softens. In contrast, at temperatures below 1000 °C, the internal friction of the basal sample (Fig. 3d) is a strong power law of frequency, $Q^{-1} \propto \omega^{-0.35}$. The response of a purely elastic material would be independent of frequency, and this frequency dependence indicates a partially viscous response even at ~600 °C.

Similar experiments carried out on cores cut from the 1999 Etna sample not only reproduced the internal friction peak at low temperatures (~700 °C) but demonstrated considerable temporal variations in the results (Fig 4). Over periods of up to ~120 hours, the measured shear modulus increased and internal friction decreased. Similar changes have been previously observed in thermally cycled quartzite, granite and sandstones (Johnson and Toksöz, 1980; Lu and Jackson, 1998) and are thought to be due to the presence and healing of microcracks. For example, rapid thermal cycling of granite and diabase samples from room temperature to ~800 °C results in order of magnitude increases in crack porosity and a factor of 2 increase in $Q^{-1}$ due to the thermal production of cracks (Johnson and Toksöz, 1980). Thus we interpret our data, which indicate an increasingly elastic response with time, as demonstrating the effects of healing microcracks in the samples. It is likely that the microcracks were originally created due to thermal stresses during cooling. Progressive volatile loss could also be responsible for strengthening the samples, however this could not be used in order to explain the power law dependence of internal friction in the 1992 Etna samples at low temperatures. After annealing, the 1999 samples produced similar shear modulus and internal friction results (Fig. 4c, d) to those of the 1992 Etna crustal material.
The experiments carried out on the Hawai‘i sample were extended to higher temperatures because of the sample’s higher melting point. Similar temporal variations to those observed with the Etna lavas were found and are shown in Fig. 5 at a temperature of 1102 °C. The results are broadly similar to those of the Etna samples with shear moduli of ~18 GPa decreasing to ~0.2 GPa as the melting temperature is approached.

The results from the experiments carried out on the Vesuvius sample are given in Fig. 6. During sample annealing the shear modulus increased but to a lesser extent than in Etna samples. An increase in tensile strength and Young’s modulus after annealing was also reported during bending experiments (Rocchi et al., 2002). At the highest temperature attained during experiments (1132 °C) the shear modulus was less than 0.5 GPa and practically frequency independent. With a further increase of temperature the sample became too fluid for it to be held within the torsion apparatus.

4. Discussion

Measurements of the complex shear modulus \( (G^*) \) and internal friction \( (Q^{-1}) \) of lavas from Etna, Vesuvius and Hawai‘i samples show that lava possess an appreciable shear modulus and the internal friction is generally less than 1.

Converting the shear modulus and internal friction data into viscosity (Eq. 3) shows that both the Vesuvius and Etna samples maintained a frequency dependent real component of their viscosity \( (\mu') \) up to the highest temperatures attainable during the experiments (Fig. 7). Thus, no Newtonian viscosity can be given for these temperatures at low strain rates. By contrast, the Hawaiian sample demonstrated marked decreases in the frequency dependence of \( \mu' \) at high temperatures and low frequencies (Fig. 8a). If a material has a strain rate dependent yield stress then two different viscosities, \( \mu_0 \) and \( \mu_{\infty} \), can be assigned, which relate to the low and high strain rate limits respectively. The Cross model then gives the viscosity as
\[ \mu = \mu_\infty + \frac{\mu_0 - \mu_\infty}{1 + \left( K \cdot \dot{\gamma} \right)^n} \]  

(5)

where \( m \) and \( K \) are empirical constants and \( \dot{\gamma} \) is strain rate (e.g. Barnes, 1999). The high temperature data in Fig. 8a have been fitted with Cross curves (assuming that \( \mu_0 \gg \mu_\infty \)) and the values of \( \mu_0 \) obtained are plotted in Fig. 8b along with other viscosity measurements made with a parallel plate viscometer (a Bähr® high temperature dilatometer, University Bayreuth (Bagdassarov, 2000)) and a rotational viscometer (Shaw, 1968, 1969; Ryan and Blevins, 1987; Ryerson et al., 1988; Pinkerton et al., 1995). The parallel plate viscometer uses shear rates of \( 10^{-5} \) to \( 10^{-7} \) s\(^{-1}\), compared with an equivalent of \( \sim 10^{-2} \) to \( 10^2 \) s\(^{-1}\) for the torsion apparatus. The activation energy for viscous flow obtained by the dilatometer and torsion experiments (given by the gradient of the dashed lines in Fig. 8b) is \( \sim 950 \pm 5 \) kJ mol\(^{-1}\). The small offset (0.67 log units) between the two lines is due to the effect of the volume viscosity, \( \mu_v \), in pure shear deformation, which affects the results of the dilatometer experiments. Thus, assuming that the measured dilatometer viscosity is a pure compressional viscosity, \( \mu_c \), and the torsion experiments provides the viscosity for simple shear, \( \mu_s \),

\[ \mu_c \approx 10^{0.67} \mu_s \]  

(6)

Applying the relationship \( \mu_c = \mu_s + 4/3 \mu_s \) (e.g. Bagdassarov and Dingwell, 1992) this gives the volume viscosity as estimated from the observed difference between the two types of experiments (pure and simple shear) as \( \sim 3.35 \mu_s \).

Field measurements, corrected to unit strain rate, have reported significantly smaller shear viscosities for Hawaiian basalts. Shaw et al. (1968) measured plastic viscosities of 650 to 750 Pa s at 1130 °C and Pinkerton et al. (1995) measured viscosities of 234 to 548 Pa s at 1146 °C (Fig. 8b). Unit shear rate results from the torsion experiments (as given by the fitted Cross models) are five orders of magnitude greater than these field measurements. For Mount
St. Helens dacite, Pinkerton and Stevenson (1992) demonstrated that a combination of factors was responsible for a 10 order of magnitude variation within measured and calculated apparent viscosities at sub-liquidus temperatures. For Hawaiian basalt, although differences in volatile content, crystallinity and vesicularity exist between the experimental and the fieldwork samples, the main difference in the results (Fig. 8b) is due to the magnitude of the strain used, with rotational viscometry producing high strains and torsion and dilatometer experiments producing small strains.

When magmas or lavas are subjected to small strains, the rheology is influenced by the deformation of gas bubbles and by the rotation, interaction and small displacement of crystals suspended in the viscous melt matrix. In this case, the rheology is controlled by the “structure” of the material and is consequently relatively high (see Fig. 8b, torsion and dilatometric experiments). For rotational viscometry (either in the field or in laboratory experiments) the strains are much higher and, with time, the material “structure” becomes disrupted. The increasingly “unstructured” nature of the material results in a viscous shear thinning response (Barnes, 1997), producing decreased apparent viscosities.

In torsion experiments the viscosity results from the Hawai’i sample demonstrate the effect of a large volume percentage of deformable vesicles. Only with the Hawai’i sample (~50 vol.% of vesicles) were our experiments close to a shear rate independent viscosity. For the Etna and Vesuvius samples (<10 vol.% of vesicles) a frequency independent viscosity was not detected, even at temperatures well above the glass transition temperature of basalt glass (~820 to 850 °C). This is in agreement with previous measurements (on partially molten rocks and melt-crystal suspensions) which indicate that a frequency independent shear viscosity is unobtainable at low strains and stresses for samples with a melt phase <40 vol.% (Bagdassarov et al., 1994; Bagdassarov and Dorfman, 1998).
Our small stress and strain rate experiments demonstrate a second, relatively high viscosity, plateau in the viscosity-strain rate dependence (Barnes, 1999) which cannot be observed with high stress or strain rate measurements. However, at temperatures above \(~1145\) to \(1150\) °C the groundmass of lava becomes so fluid that the viscosity decreases a few orders of magnitude (Fig. 8b) and the high viscosity plateau is no longer present. At temperatures about their glass transition point, lavas exhibit viscoelastic behaviour which can be of importance in problems such as dome growth and collapse and the slow development of lava flow fronts. Numerical modelling of lava flows should consider the high viscosity and viscoelasticity of lavas when the strain rates are below unity.

Our experiments have also illustrated the effect of annealing on the strength of lava samples. With time at temperatures between \(~500\) and \(1000\) °C, some samples became increasingly stiff and elastic, an effect we attribute to the slow closure and healing of microfractures initially produced during cooling. The presence of microfractures dictates that rocks that have undergone slow cooling will possess higher shear moduli and have higher strength than rocks which cooled more rapidly. Thus, before annealing, samples taken from the centre of a cooled flow would be expected to be more elastic than samples taken from the top of a flow. Under annealing at high temperature, the crack healing process consists of cracks pinching off and healing to a pore-like shape before subsequent decreasing of the pore diameter (Atkinson, 1984). During this period, the shear modulus may significantly increase and the internal friction decrease. The time dependence of changes in crack lengths can be described as an Arrhenius function of temperature and, as a first approximation, a linear dependence between shear modulus and a crack density parameter may be assumed (O'Connel and Budiansky, 1974). In Fig. 9a the results of annealing on the complex shear modulus of the 1999 Etna lava sample are presented as a function of time, \(t\). By fitting these data with an exponential time-dependence,
\[ G^* = G_\infty \left[ 1 - e^{-\frac{t}{\tau}} \right] \]  

where \( G_\infty \) is the stable shear modulus at any temperature, the characteristic time \( \tau \) of the crack healing process may be estimated. The results of the fitting are presented in Fig. 9b in the form of an Arrhenian dependence of \( \tau \), yielding an activation energy of 150 ± 20 kJ mol\(^{-1}\). According to Atkinson (1984) this value should relate to the activation energy of the grain boundary diffusion coefficient.

5. Conclusions

1. The oscillatory torsional apparatus allows low strain investigations of shear modulus and internal friction. Experiments carried out on lavas over a range of temperatures (~500 to 1150 °C) and frequencies (20 to 0.002 Hz) demonstrate that in general the complex shear modulus of lavas decreases with decreasing frequency and increasing temperature. Internal friction is usually ≤1.

2. At temperatures close to their eruption temperature (~1080 to 1100 °C) no purely viscous regime was detected for the Etna and Vesuvius samples and their measured viscosity remained frequency dependent over the range 20 to 0.002 Hz. However, results for the Hawai'i sample indicate a shear rate independent regime for low shear rates at temperatures between ~1070 and 1130 °C, with viscosities >10\(^9\) Pa s.

3. The samples from Etna and Vesuvius exhibited anelastic behaviour at temperatures much lower than the “glass transition” temperature of the groundmass, and that this may be attributable to the presence of microcracks. Annealed lava at ~900 to 950°C possesses a shear modulus about 15 to 20 GPa. Below ~800 °C, when the lava becomes anelastic, intensive microcracking may be expected and can decrease the shear modulus to 7 to 10
GPa. Between 700 and 950°C the characteristic time-constant of the crack healing process in lavas may be on a scale of several to several hundred hours.
References


Fig. 1. (a) Schematic of the torsion oscillation apparatus (redrawn from Berckhemer et al., 1982). Capacitive pickups detect the motion of the iron plates at the ends of the aluminium wings providing two data channels which can be calibrated to provide angular deflection and the applied torque. (b) An example of two periods of data collected in order to measure internal friction (from the phase shift) and magnitude of the complex shear modulus (from the relative amplitudes of the curves). Further details are given in Bagdassarov (2000).

Fig. 2. (a) Thin section of the crustal sample from Etna, 1992. The lava contains ~15 to 20 vol.% of vesicles (mean diameter between ~0.2 to 0.5 mm) and ~20 to 25 vol.% of plagioclase, olivine and magnetite phenocrysts. (b) Thin section of the 1992 Etna sample which was collected from the base of an overflow. This sample contains ~20 vol.% of vesicles with a mean diameter of 1 to 2 mm and about 30 vol.% of crystals.

Fig. 3. Shear modulus (a, b) and internal friction (c, d) results from the 1992 Etna lavas. The crustal sample (a, c) shows a higher shear modulus than the basal sample, and neither sample has an internal friction approaching 1 over the conditions investigated.

Fig. 4. In (a) and (b), temporal variations in shear modulus and internal friction are given for the 1999 Etna sample. After annealing for 118 hours at 800 °C the sample had attained a greater shear modulus than it originally possessed at 700 °C. The increasing shear modulus and decreasing internal friction with time indicate the sample was becoming increasingly elastic and less viscous. The shear modulus and internal friction results collected after annealing are given in (c) and (d).
Fig. 5. Time dependent shear modulus (a) and internal friction (b) during annealing experiments on the Hawai’i sample. Numbers in the legend indicate annealing time in hours.

Fig. 6. Shear modulus (a) and internal friction (b) results from the Vesuvius lava.

Fig. 7. Dynamic viscosity as given by $G''(\omega)/\omega$ (see Eq. 4) for the Vesuvius (a) and Etna (b) samples. At the highest temperatures and lowest frequencies used these samples maintained a frequency dependent rheology.

Fig. 8. Dynamic viscosity of the Hawai’i sample. In (a), the high temperature values of $G''(\omega)/\omega$ are given, demonstrating the decreasing dependence on frequency at low frequencies. The curves show the results of using a Cross model in order to extract the zero-shear viscosity (see text). In (b), the zero-shear viscosity values obtained are compared with those given by dilatometer (Rocchi et al., 2002) and rotational viscometer (Shaw, 1968, 1969; Ryerson et al., 1988; Ryan and Blevins, 1974; Pinkerton et al. 1995) experiments. There is good agreement between the dilatometer and the torsional results, with the slightly greater values from the dilatometer being expected due to the effect of compression viscosity on extracting shear viscosity values from the pure shear experiments (Bagdassarov and Dingwell, 1992). The gradient of the dashed best fit lines represents an activation energy for viscous flow of ~950 kJ mol$^{-1}$. The plot demonstrates the ~3 order of magnitude change in viscosity which occurs around 7.0x10$^{-4}$ K$^{-1}$ (~1155 °C) and separates the high temperature, Newtonian region from the lower temperature, viscoplastic region. The 1 s$^{-1}$ line indicates the torsion results at unit shear rates which are more applicable for comparison with the rotational viscometer measurements.
Fig. 9. (a) The complex shear modulus of Etna lava measured at 20 Hz and different temperatures. The temporal change at each temperature is modelled using a curve of characteristic time-constant which is believed to represent the characteristic time for crack healing (see text). In (b), these characteristic times are plotted against absolute reciprocal temperature. The straight line fit represents an Arrhenian dependence with an activation energy of $150 \pm 20 \text{kJ mol}^{-1}$. 

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Figure 9

(a) Complex shear modulus (GPa) vs. Time (hours) for different temperatures (°C): 900, 850, 800, and 700.

(b) Characteristic time (s) vs. Reciprocal temp. (K⁻¹)