Viscoelastic behaviour of basaltic lavas

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Abstract

The rheological properties of basaltic lavas from Etna, Hawai'i and Vesuvius have been investigated at temperatures between ~500 and 1150 °C using a small strain oscillatory shear. The viscoelastic response of the lavas to small, forced, sinusoidal torques (<10^{-3} N m) at frequencies between 0.002 and 20 Hz was measured. 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 may be attributed to the slow closing (healing) of microcracks and small pore spaces resulting in the apparent stiffening of lava samples under annealing. Thus, those parts of lava flows that undergo slow cooling will have more elastic properties. Regions which cool faster possess smaller shear moduli and higher internal friction due to thermal microcracking and less cohesion between crystals and the bulk glassy matrix.

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

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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 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 temperatures below

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their eruption temperature, lavas display viscoelastic behaviour and, at temperatures below the glass transition temperature, T_g, of their glass matrix, they are anelastic (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. Quantification of these processes may be important, for example, in the prediction of lava flow carapace generation or break up. 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 rate of lava flows 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 (G^*) and internal friction (Q^{-1}). 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.

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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 sinusoidal 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).

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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 $cm^3\ s^{-1}$). Inspection of the samples after experiments showed that oxidation (as indicated by growth of Fe-(Ti)-oxides) 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).

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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 ω is the angular velocity (equal to 2π 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, η, can be calculated from

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$$G^* = G' + iG'' = G^* \cos(\varphi) + iG^* \sin(\varphi) \qquad (1),$$

$$Q^{-1} = \tan(\varphi) = \frac{G''}{G'} \quad (2),$$

$$\eta = \eta' - i\eta'' = \frac{G''}{\omega} + i\frac{G'}{\omega} \quad (3),$$

(e.g. Nowick and Berry, 1972). The zero-rate shear viscosity (macroscopic viscosity), η_0, may be obtained from the frequency dependence of G'' by

$$\eta_0 = \lim_{\omega \to 0} \frac{G''(\omega)}{\omega} \quad (4),$$

(Marin, 1998).

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2.1 Sample bonding

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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 ~1°, 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 ~8 N (e.g. Berckhemer et al., 1982). Measurements carried out using a dummy sample of Al_2O_3 have demonstrated that the effect of the cement on phase delay measurements was less than 10^{-3} rad. (Bagdassarov and Dorfman, 1998).

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2.2 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 the correspondingly changed 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.

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2.3 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. Digital images of polished sections were obtained through an optical microscope and computer analysed, omitting corrections for 3D effects. 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 thin sections 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, 2002) at the top of the active flow field.

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b. Hawai'i. The Hawai'ian 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 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 corresponding to that of the glass transition

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for basaltic glass (Ryan and Sammis, 1981). For Pu'u 'O'o paheohoe lavas, the glass transition and melting temperatures are 655°C and 1149°C respectively (Burkhard, 2001). 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. After annealing, the 1999 samples produced similar shear modulus and internal friction results (Fig. 4c, d) to those of the 1992 Etna crustal material. At temperatures >1100 °C, Etna samples demonstrated an increase of fluidity of several orders of magnitude and it was not possible to maintain them in the torsion apparatus.

The experiments carried out on the Hawai'i sample were extended to higher temperatures because of the higher temperature of the sample's softening 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 sample fluidity starts to rapidly increase.

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.

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4. Discussion

Measurements of the complex shear modulus (G^*) and internal friction (Q^{-1}) of lava samples from Etna, Vesuvius and Hawai'i show that they possess an appreciable shear modulus (>0.5 GPa) and an internal friction generally less than 1, at temperatures where field measurements indicate viscosities of ~10^3 Pa s. For comparison, polycrstalline rocks at room temperature have typical values of unrelaxed shear modulus, G_∞, of about 50 GPa (Jackson, 1993) and 25 GPa is representative for silicate glasses (Bagdassarov et al., 1993). Between room and high temperatures prior to the onset of melting, values of Q^{-1} vary between ~10^{-3} and 10^{-1} (Manghnani et al., 1981; Weiner et al., 1987; Jackson, 1993; Bagdassarov, 2002).

Natural silicate melts and glasses show a linear viscoelastic behaviour in the glass transition temperature range (e.g. Bagdassarov and Dingwell, 1993; Bagdassarov et al., 1993). With increasing temperature and decreasing strain-rate, one expects progressively decreasing shear modulus and increasing internal friction. However, the viscoelastic behaviour of lavas is often controlled by the presence of deformable (vesicles) and non-deformable (crystals) heterogeneities. Viscosity increases with increasing crystal content but decreases with increasing vesicle content (Bagdassarov and Dingwell, 1992) and the presence of these macroscale obstacles extend the shear stress relaxation spectrum towards longer relaxation times. Shear modulus relaxation can be extended over several hundred °C above the T_g of a basaltic groundmass glass (~655 °C) up to ~1100 to 1150 °C. Over the same range, the frequency dependence of shear modulus and internal friction becomes weaker, for example, in lavas $Q^{-1} \propto \omega^{-0.35}$, compared to $\omega^{-0.5}$ for silicate melts.

The shear modulus and internal friction of our lava samples varied significantly with annealing time with, at temperatures between ~500 and 1000 °C, some samples becoming increasingly stiff and elastic (Fig. 4). Similar temporal variation of elastic modulus and internal friction have been previously observed in thermally cycled quartzite, granite, dunite

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and sandstones (Johnson and Toksöz, 1980; Jackson, 1993; Lu and Jackson, 1998) and is thought to be due to the presence and healing of microcracks and pores. 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). Jackson (1993) interpreted the observed increase in shear modulus and decrease in internal friction with the increasing annealing time to be a result of the reduction of porosity and increase of the grain boundary cohesion by a sintering processes. Progressive decreases in Q^{-1} have been associated with a decrease of centres of stress concentration. Thus, we interpret our data, which indicate an increasingly elastic response with time, as demonstrating the effects of healing microcracks and thermally produced pore space (not vesicles) in the samples. It is likely that the microcracks were originally created due to thermal stresses during rapid cooling of the lava flows, however, we cannot exclude the possibility that they could have been produced during sample preparation or the subsequent heating. In thin sections, microcracks within crystals and between crystals and the groundmass can be observed to have undergone some healing after annealing (Fig. 7).

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The production of microcracks suggests that rocks which have undergone slow cooling will possess higher shear moduli and have higher strengths 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. 8a 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,

$$|G^*| = (G_\infty - G(0))\left\{1 - e^{-\left(\frac{t}{\tau}\right)}\right\} + G(0) \quad (5),$$

where G_∞ is the stable shear modulus at any temperature, $G(0)$ is the shear modulus at t=0, the characteristic time, τ, of the crack healing process may be estimated. The results of the fitting are presented in Fig. 8b in the form of an Arrhenian dependence of τ, yielding an activation energy of 150 ± 20 kJ mol^{-1}. According to Atkinson (1984) this value should relate to the activation energy of the effective diffusion coefficient on grain boundaries of species participating in the pinching process at microcrack tips. For example, the activation energy of crack healing in dry quartz at T< 600 °C is about 80 to 85 kJ mol^{-1} (Atkinson and Meredith, 1987), corresponding to the activation energy of water diffusion in quartz (~60 kJ mol^{-1}, Dersch et al. (1997) or ~100 kJ mol^{-1}, Brady (1995)). At high temperatures, crystal lattice diffusion is the rate controlling mechanism of grain boundary diffusion. Thus, the observed activation energy for the time dependence of shear modulus may be associated with an inter-diffusion of Ca, Mg and alkaline species between olivine, clinopyroxene and plagioclase crystals and basalt glass, with activation energies between 130 and 180 kJ mol^{-1} (Brady, 1995). Alternatively, our results are also consistent with a self-diffusion of Si and O, network formers in basaltic glass, with an activation energy of 170 kJ mol^{-1} (Lesher et al., 1996).

Alternative mechanisms which could contribute to the increase of shear modulus during annealing of samples are redox and re-crystallisation processes in the basalt groundmass glass. Progressive volatile loss could also be responsible for strengthening the samples, however this cannot explain the power law frequency dependence of internal friction in the 1992 Etna samples at low temperatures. A contribution to shear modulus increase from sample oxidation would be due to the production of interface-controlled intergrowths of pyroxene dendrites and Fe-(Ti)-oxides between 850 and 940 °C (Burkhard, 2001). However,

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the activation energy of this process is ~100 kJ mol^{-1}, which is less than that observed from the time variation curves of shear modulus. The oxidation kinetics of basaltic glass are determined primarily by the diffusion of divalent cations Ca^{2+} and Mg^{2+} to the surface of the sample, which is charged compensated by the inward flux of electron holes (Cooper et al., 1996). The activation energy of this process at temperatures below T_g is 210 kJ mol^{-1}. The calculated maximum depth (~1 µm at 600 °C and ~20 µm at 800 °C) of surface oxidation by using the average divalent cation diffusion coefficient (Cooper et al., 1996) is small compared with the size of lava samples used in torsion experiments and is in agreement with observations made from thin sections of the samples. It should also be noted that these rates relate to initially non-oxidised basalt glass samples and the lava samples should be considered as slow cooled, partially crystallised glasses, with long oxidation histories.

At T> 920 °C, the bulk crystallisation of basaltic glass and growth of plagioclase crystals may also contribute to the increase of stiffness of annealed lava samples, but the timescales required are generally much longer than those of the observed changes in shear modulus. For example, a noticeable growth of plagioclase and Fe-Ti-oxides has been observed at 850-934 °C after >200 h of annealing basalt glass (Burkhard, 2001) and this process is likely to have continued for an order of magnitude longer time. In contrast, shear modulus changes occurred much more rapidly in our experiments and generally ceased on a timescale of <100 hours. Thus, neither redox or re-crystallisation processes are believed to significantly contribute to the measured changes in the rheological properties of the samples.

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 (η') up to the highest temperatures attainable during the experiments (Fig. 9). Therefore, no "Newtonian" viscosity can be given for these temperatures at low strain rates. In contrast, the Hawaiian sample demonstrated marked decreases in the frequency dependence of η' at high temperatures and low frequencies (Fig. 10a). The main difference

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between the low strain rate behaviour of, on the one side, Etna and Vesuvius basalts, and on the other, that from Hawai'i, lies in the relative proportion of crystals and vesicles. The small strain-rate and small stress viscosity, η_0, measured for the Hawaiian sample represents a creep-type mechanism underlying the Bingham yield stress which is expressed under conditions of slow, small scale deformation, where the internal structure of the lava sample remains intact. The presence of non-deformable crystals results in increases of this effective viscosity, but also increases the Bingham yield stress (Bagdassarov et al., 1994). Higher ratios of crystals to vesicles in Etna and Vesuvius lavas increase the Bingham creep-viscosity, η_0, and shift it toward smaller strain-rates which were unattainable in our torsion experiments. In contrast, the Hawai'ian lava sample, with a smaller crystals to vesicles ratio, possessed a smaller Bingham creep-viscosity which was observable at the lower strain rates of the torsion apparatus (10^{-3} s^{-1}).

If a material has a strain rate dependent rheology, then two different viscosities can be assigned, η_0 and η_∞, which relate to the low and high strain rate limits respectively. The Cross model then gives the viscosity as

$$\eta = \eta_\infty + \frac{\eta_0 - \eta_\infty}{1 + \left(K \cdot \dot{\gamma} \right)^m} \qquad (6),$$

where m and K are empirical constants and $\dot{\gamma}$ is strain rate (e.g. Barnes, 1999), with m characterising a stretching parameter for shear stress relaxation during the viscoelastic transition. For a Maxwell body rheology, m=2. In the operational window of the oscillatory shear apparatus, only η_0 can be measured. The high strain rate viscosity, η_∞, is interpreted as the effective viscosity of a fluid at high strain-rates or stresses. In the case of lavas, it corresponds to the results of field or laboratory experiments in which high strains induced by the measurement disrupt the internal structure of the lava.

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The high temperature data in Fig. 10a have been fitted with Cross curves (assuming that $\eta_0 >> \eta_\infty$, and using m=1) and the values of η_0 obtained are plotted in Fig. 10b along with other viscosity measurements made with a parallel plate viscometer (a Bähr® high temperature dilatometer, University Bayreuth (Bagdassarov, 2000)) and rotational viscometers (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 equivalents of ~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. 10b) is ~950 ± 5 kJ mol^{-1}. The small offset (0.67 log units) between the two lines is due to the effect of the volume (or bulk) viscosity of a porous sample, η_v, which affects the parallel plate deformation results of the dilatometer experiments. The pure shear deformation used in the dilatometer gives a compressional viscosity, η_c, which is usually related to the shear viscosity by $\eta_c/3$. However, this is valid only for incompressible samples (which have an infinite volume viscosity), and is therefore not applicable to material with ~50 % porosity, which will have a finite volume viscosity. For high porosity materials, the relationship

$$\eta_c = \eta_v + \tfrac{4}{3}\eta_s \qquad (7)$$

is appropriate (e.g. Bagdassarov and Dingwell, 1992). Thus measured values should be corrected for a finite volume viscosity, the value of which is unknown. From Fig. 10b the compressional viscosity data from the dilatometer experiments are related to the simple shear viscosity results of the torsion experiments, η_s, by $\eta_c \approx 3\times10^{0.67}\eta_s$, or $10^{1.15}\eta_s$. Applying Eq. 7 thus gives $\eta_v \approx 12.7\eta_s$ (at ~50 % porosity). The theoretical dependence of the volume viscosity as function of porosity has been considered elsewhere (Prud'homme and Bird, 1978; Aksel, 1995), with the classical formula

$$\eta_v = \frac{4}{3}\cdot\frac{\rho^2}{\rho_0(\rho_0-\rho)}\cdot\eta_s^0 \qquad (8),$$

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being derived for small concentrations of vesicles in a fluid, where ρ_0 is the density of fluid, ρ is the density of the suspension and η_s^0 is the viscosity of the fluid without vesicles. Converting η_s^0 to η_s by $\eta_s^0 \approx 0.2\,\eta_s$ (for ~50 % porosity, Fig. 5, Rust and Manga 2002), gives the theoretical relation of $\eta_v \approx 3.3\,\eta_s$. The discrepancy between our result and this theoretical estimation may be easily attributed to experimental errors in viscositiy as well as the non-spherical shape of some vesicles and their non-uniform size distribution in the lava samples.

Alternatives to the Cross model were reviewed by Yue and Brückner (1994) who suggested a strain rate dependent viscosity model depending on three phenomenological constants. Their model is appropriate for parallel plate viscometry in which significant strain increases during experiments can produce thermal effects and violations of non-slip boundary conditions. However, in the torsion experiments the maximum torque was 10^{-3} N m, and maximum angle deformation 10^{-5} rad, resulting in a negligible heat production (<10^{-7} J s^{-1}) even at the highest strain rates used.

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. 10b) is due to the magnitude of the strain used, with rotational viscometry producing high strains and torsion and dilatometer experiments producing small strains.

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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. 10b, 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.

The small stress and strain rate experiments carried out in torsion and dilatometer equipment demonstrate a second, relatively high viscosity (creep-type rheology), 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. 10b) and the high viscosity plateau is no longer present. Fig. 10b demonstrates the difference between the η_0 (creep-type) and η_∞ (viscoplastic) viscosities. In the small strain and small stress torsion equipment, the internal structure of lava sample is not destroyed, and the measured viscosity is of a creep-type. This type of viscosity may be expressed in lava dome growth or relaxation processes. In experiments with rotational viscometers where strains and stresses are considerable, the internal structure of the lava is destroyed and the apparent viscosity is several magnitudes smaller. This type of viscosity will control lava flows over steep topography. Thus, at the same temperature, lavas may possess two viscosities depending on a stress-strain scale and numerical modelling of lava flows should consider the high viscosity and viscoelasticity of lavas when strain rates are below unity.

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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 creep-type 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 extended anelastic behaviour at temperatures within the "glass transition" temperature of the groundmass, and that may be attributable to the bad cohesion between crystal grains and the groundmass glass. Annealed lava at ~900 to 950°C possesses a shear modulus about 15 to 20 GPa. Below ~800 °C, but still within basalt glass transition temperature range, when the dilatometric effect of between groundmass glass and crystals are significant, intensive microcracking may be expected and can decrease the shear modulus to 7 to 10 GPa. By reheating between 700 and 950°C the characteristic time-constant of the lava hardening process may be on a scale of several to a hundred hours.

Acknowledgements

We thank J. Deubener and one anonymous referee for helpful comments which improved the manuscript.

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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 crystal (from analysis of polished thin sections without corrections for 3D effects).

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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. Thin sections of Etna (**a** and **b**) and Vesuvius (**c** and **d**) lava samples (each image represents 1250 × 950 µm). (**a**) and (**b**) are starting material and show microcracks between crystals and groundmass glass as well as in the interior of phenocrysts. (**b**) shows the Etna sample after being annealed at 812 °C for 118 hours and (**d**) shows the Vesuvius sample after being annealed at 1102 °C for 96 hours. The reduction in the number of microcracks demonstrates that healing has occurred both within the groundmass and the phenocrysts.

Fig. 8. (**a**) The complex shear modulus of 1999 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 ± 20 kJ mol^{-1}.

Fig. 9. 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. 10. 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). Fitting parameter of Eq 5 m= 1 at all temperatures, K=1.6 log(η_o Pa

s)= 9.23 at 1126°c, 3.7 and 9.7 at 1100°, 30 and 10.65 at 1072°. In (**b**), the zero-shear viscosity values η_o obtained for the Hawai'i lava sample are compared with those given by dilatometer (Bagdassarov, 2000) and rotational viscometer (Shaw, 1968, 1969; Ryerson et al., 1988; Ryan and Blevins, 1974; Pinkerton et al. 1995) experiments on Hawai'ian basalt lavas. 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 7.0×10^{-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. High values of η_o obtained in torsion and dilatometer, small strain-stress experiments relate to the creep-type rheology displayed by lava when its internal structure is not destroyed by the measurements. Field experiments carried out at high stresses and strains show η_∞, or a viscoplastic viscosity, expressed when the internal structure (the organisation of vesicles and crystals) of the lava is altered by the viscometer or the flow itself.

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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 ± 20 kJ mol^{-1}.¶
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Fig. 10. Thin sections of Etna (A and B) and Vesuvius (C and D) lava samples. Photos 1250 x 950 µm. A and C is starting material. The contacts between crystals and groundmass glass as well as the interior of phenocrystals are marked by microcracks. B is Etna sample annealed over 118 h at 812°C. D is Vesuv sample annealed over 96 h at 1102°C. The contacts between groundmass glass and crystals as well as the interior of phenocrysts are healed.¶
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lava

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the

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for polycrstalline rocks

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in silicate glasses is

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The

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attenuation

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from room temperature to the temperature range prior to the beginning of

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melting

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ies

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the

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according to the principle of the thermorheological simplicity

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The main difference in

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i

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In this case,

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viscoity

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(effective T_g of a sample)

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in the case of

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increase or

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in the case of th

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e

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This effect is clearly demonstrated by an extended temperature dependence of shear modulus and internal friction

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on frequency and temperature

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The relaxation of the s

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in basalt lava

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c.

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)

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At the same time

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the

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the

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rather

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than

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instead of

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the

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in

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Aheim

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At T> 920 °C, the bulk crystallisation of basaltic glass and growth of plagioclase crystals may also contribute to the increase of stiffness of annealed lava samples. Analysis of thin sections of samples after experiments revealed the textural changes only in thin surface layer (few μm) and not in the bulk of samples. A noticeable growth of plagioclase and Fe-Ti-oxides was observed at 850-934°C after >200 h of annealing (Burkhard, 2001) that is longer that the duration of torsion annealing experiments. The end of these processes must take 10 times longer time, despite the constant steady state values of shear modulus were observed after <100 h.

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the additional annealing of lava samples in laboratory is not believed to

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significantly contribute to further oxidation

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in the bulk of the samples.

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Bingham yield stress and

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under conditions of slow, small scale deformation, where the

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internal structure of lava samples remains intact

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of lava samples is not destroyed

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Parameter *m* characterises a stretching character of shear stress relaxation during viscoelastic transition. For a Maxwell body rheology *m*=2 in Eq 5.

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8a

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μ

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a

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Parameter *m*=1 in Eq 5 and indicates on a distribution of a shear stress relaxation spectrum.

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μ

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, in pure shear deformation,

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Parallel plate

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s

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samples. Thus, the viscosity calculated from a simple shear deformation

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in a dilatometer and shown in Fig. 8b is ⅓ “

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shear viscosity from parallel plate viscometry. The correction of the measured viscosity by dividing to the factor 3 stems from the assumption that the material is incompressible and possesses an infinite volume viscosity. It is not true for a sample with porosity of c. 50 vol.%.

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e

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must

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Thus, assuming that the measured

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dilatometer viscosity is a

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compressional viscosity, μ_c, and, f

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, is r

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from the torsion experiments

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provides the viscosity for simple shear, η

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$_s$, and they are related as follows

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μ

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$\approx$

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x

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(6).

Applying the relationship $\mu_c = \mu_v + 4/3 \cdot \mu_s$

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appropriate for materials with high porosity

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(e.g. Bagdassarov and Dingwell, 1992), thisthis

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the volume viscosity as estimated from the observed difference between the two

types of experiments (pure and simple shear) as ~

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3.35

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μ

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c.

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of

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4/3 (Prud'homme and Bird, 1978)

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difference between our result and

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Eq

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which

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to

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the

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where

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significantly

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the

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. High strains

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of deformation,

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some

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, transient flow after applying a new load etc.

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which

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s

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of experiment

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This heat production cannot result in a significant temperature effect.

Transient flow in oscillatory shear does not play any significant role, amplitude and phase of the angle deformation were calculated from sinusoidal signals after a many repeated oscillations.

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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).

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Our

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8b

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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. The main difference

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between the low strain rate behaviour of, on the one side, Etna and Vesuvius basalts

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, and on the other, that from Hawai'

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lies in the relative proportion of crystals and vesicles.

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presence of deformable inclusions leads to

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decreases at small strain-rates and small stresses (Bagdassarov and Dingwell, 1992). The presence of non-deformable crystals results in increases

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the Bingham yield stress (Bagdassarov et al., 1994). Higher ratios of crystals to

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vesicles in Etna and Vesuvius

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lavas increase the Bingham creep-viscosity, η_0,

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which was observable at the strain rates of the torsion apparatus.

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clearly

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strain-rates which were unattainable in

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the operational window of

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our torsion experiments (10^{-3} s^{-1}). In contr

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was much smaller and because of this fact

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Bingham creep-viscosity

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