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Physical properties of the 1980 Mount St. Helens cryptodome magma

Abstract Physical properties of cryptodome and re-melted samples of the Mount St. Helens grey dacite have been measured in the laboratory. The viscosity of cryptodome dacite measured by parallel–plate viscometry ranges from 10.82 to 9.94 $\log_{10}\eta$ (Pa s) ($T = 900$–982 °C), and shrinkage effects were dilatometrically observed at $T > 900$ °C. The viscosity of re-melted dacite samples measured by the micropenetration method is 10.60–9.25 $\log_{10}\eta$ (Pa s) ($T = 736$–802 °C) and viscosities measured by rotational viscometry are 3.22–1.66 $\log_{10}\eta$ (Pa s) ($T = 1298$–1594 °C). Comparison of the measured viscosity of cryptodome dacitic samples with the calculated viscosity of corresponding water-bearing melt demonstrates significant deviations between measured and calculated values. This difference reflects a combination of the effect of crystals and vesicles on the viscosity of dacite as well as the insufficient experimental basis for the calculation of crystal-bearing vesicular melt viscosities at low temperature. Assuming that the cryptodome magma of the 18 May 1980 Mount St. Helens eruption was residing at 900 °C with a phenocryst content of 30 vol.%, a vesicularity of 36 vol.% and a bulk water content of 0.6 wt.%, we estimate the magma viscosity to be $10^{10.8}$ Pa s.

Key words Magma · Mount St. Helens · Cryptodome · Viscosity · Physical · Rheological · Experimental

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Introduction

The paroxysmal explosive eruption of Mount St. Helens on 18 May 1980 was one of the most notable and well-documented volcanic events of this century, characterized by a catastrophic lateral blast. There is general agreement that on 18 May 1980 a magnitude 5+ earthquake at Mount St. Helens caused the failure by landslide of part of the volcanic edifice (Voight 1981; Voight et al. 1981). The landslide produced rapid decompression of the magmatic and hydrothermal system inside the volcano (Voight et al. 1981; Kieffer 1981). An ~0.1 km³ volume of intrusive magma within the volcanic edifice before eruption (the cryptodome) was suddenly depressurized and fragmented explosively in a catastrophic lateral blast that resulted in large areas (~500 km²) of devastation near the volcano (Hoblitt et al. 1981; Kieffer 1981; Eichelberger and Hayes 1982; Sparks et al. 1986; Alidibirov 1995).

Although the process of the cryptodome magma fragmentation is likely to have been strongly influenced by the rheological properties of the cryptodome magma, such properties are poorly known. The experimental study of viscosity of lava of the Mount St. Helens dome formed after the 18 May 1980 eruption by Murase et al. (1985) appears to be the only example. In their study beam-bending and counter-balanced sphere methods were employed on both lava dome and re-melted samples. They found that viscosity decreased from 10^{14} Pa s at 800 °C to 10^2 Pa s at 1500 °C (Murase et al. 1985). To date, no experimental investigation of the rheological properties of the 1980 Mount St. Helens cryptodome magma has been performed, despite the clear need for such data for the modelling of the 18 May 1980 Mount St. Helens blast.

The objective of the present study was (a) the determination of the viscosity of the 1980 Mount St. Helens cryptodome and remelted dacite samples, and (b) estimation of the viscosity of the cryptodome magma before the 18 May 1980 Mount St. Helens blast.

In addition, the data on physical properties (density, vesicularity, velocities of elastic waves, values of elastic moduli and viscosity) of cryptodome dacite presented here are important for the interpretation of recent experimental results on dacite magma fragmentation (Alidibirov and Dingwell 1996).
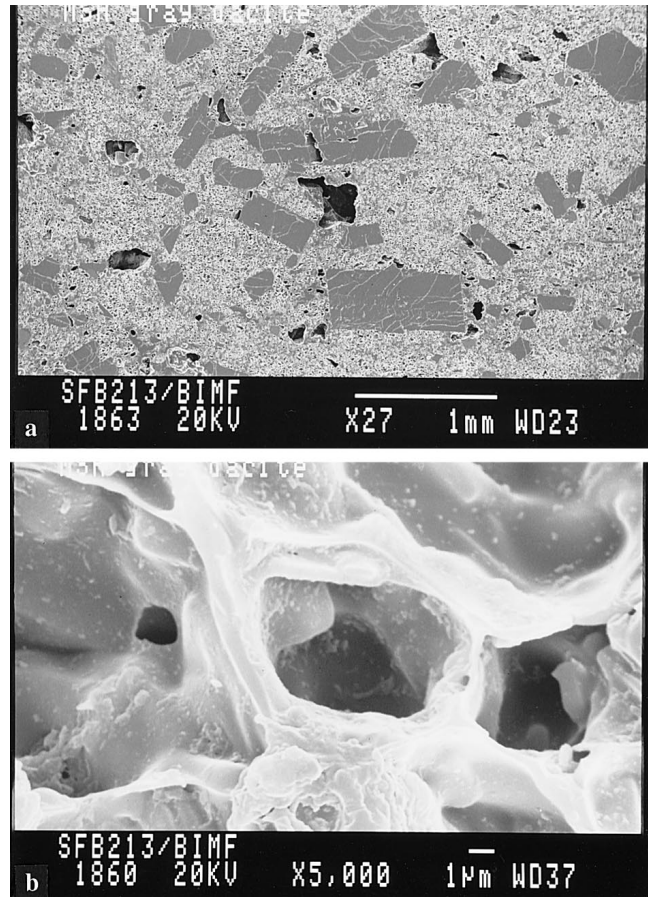
Samples

Cryptodome dacite

Light-grey, vesicular, hypersthene–hornblende dacite was the most abundant lithology (72%) erupted during the lateral blast of 18 May 1980 (Hoblitt et al. 1981; Hoblitt and Harmon 1993; Eichelberger and Hayes 1982; Cashman 1988, 1992). In the present experiments, dacite samples corresponding to grey dacite of 1980 Mount St. Helens cryptodome were employed. The grey dacite contains ~ 30 vol.% phenocrysts (on a vesicle-free basis) of 210 μm average size, comprising plagioclase, hypersthene and hornblende and Fe–Ti oxides, which are easily visible on SEM images of samples (Cashman 1988; Eichelberger and Hayes 1982). The proportion of microlites and microphenocrysts in the microcrystalline groundmass is ≈ 45 vol.% calculated on a vesicle-free basis from a BSEM image of grey blast dacite in Cashman (1988). Microlites and microphenocrysts occupying some of the interbubble partitions are typically plagioclase, and have widths ranging from 1 to 10 μm. Most vesicles range in size from several-microns width (average 5 μm) to several tens of microns. The smaller vesicles are usually elongated with aspect ratios (ratio of longest to shortest apparent axis) less than 8. Many vesicles are interconnected and the characteristic thickness of interbubble partitions is 1–4 μm (Fig. 1).

Bulk chemical composition of Mount St. Helens grey dacite samples used in experiments was determined by solution ICP-AES methods. The chemical composition of the matrix glass of cryptodome dacite and composition of samples of remelted glass was determined by microprobe (CAMECA SX-50 Courbevoie) at 15 kV and 15 nA, with a defocussed beam spot of 15 μm. Compositional data are summarized in the Table 1.

Data on density, vesicularity, water content, linear expansion coefficient and glass transition temperature of crytodome grey dacite are presented in Table 2. For p-wave velocity measurements a pulse transmission technique with 1-MHz p-wave transducers was employed. Cylinders 30 mm length and 30 mm diameter were cut and oven dried for 48 h at $T = 40\,°C$. Omnidirectional p-wave velocity measurements were performed by rotating a cylindrical sample, within a 400-MPa pressure vessel fitted with a fixed arrangement of four transducers (Zang et al., in press; Zinke 1996) with an accuracy of $\pm 1\%$. Samples were sealed using an epoxy resin to ensure stable confining pressures and prevent leakage of hydraulic oil into the sample. As these sam-

Fig. 1a, b SEM image of the 18 May 1980 Mount St. Helens grey dacite cryptodome material. **a** Polished surface; **b** glassy matrix fragment

Table 1 Chemical compositions of samples used in viscometry

	Rock sample[a]	Matrix glass of rock sample[b]	Remelted sample[c]
SiO_2	63.3	76.05	65.28 (0.53)
Al_2O_3	17.6	14.62	17.05 (0.20)
Fe_2O_3	5.0	0.44	4.47 (0.11)
MgO	1.7	0.02	1.82 (0.03)
CaO	5.2	3.01	4.70 (0.09)
Na_2O	4.7	4.96	4.34 (0.05)
K_2O	1.2	0.56	1.29 (0.02)
TiO_2	0.63	0.15	0.59 (0.03)
P_2O_5	0.12	0.10	0.13 (0.03)
MnO	0.09	0.04	0.08 (0.02)
Total	99.54	99.95	99.75

[a] Composition was determined by ICP-AES method
[b] Microprobe measurements (analysis with the highest silica content chosen to be representative of analysis with minimum external contamination)
[c] Microprobe measurements; average values for 12 points; standard deviation in parentheses

ples were of high porosity, the sealing in some cases was not strong enough.

At low confining pressures (1 MPa) p-wave velocities depend on the sample orientation and varies over

Table 2 Some physical properties of the 1980 Mount St. Helens cryptodome dacite

Properties	Values
Density	1.59×10^3 kg/m^3
Vesicularity	35.9%
Water content[a]	0.61–0.62 wt.%
Linear thermal expansion coefficient[b]	4.85 ± 10^{-6} °C^{-1}
Glass transition temperature[c]	800–810 °C

[a] Water content of the cryptodome dacite was measured by Behrens using Karl-Fischer titration method (Behrens, 1995)
[b] Linear thermal expansion was estimated dilatometrically for the temperature range 20–800 °C
[c] Glass transition temperature was determined using the position of the dilatometric peak at a heating rate 10 °C/min (Webb et al. 1992)

the range 2900–3100 m/s (Fig. 2a). This variation is likely a function of sample fracture anisotropy. In contrast, at high confining pressures (350 MPa) the measured p-wave velocity is 4000 m/s. At these confining pressures, there is no directional dependence (with respect to sample orientation) of p-wave velocity and the sample appears isotropic. The dependence of p-wave velocity on confining pressure is illustrated in Fig. 2b.

Remelted dacite

Cryptodome dacite samples were crushed, remelted at 1400–1600 °C, and stirred in a $Pt_{80}Rh_{20}$ crucible over 18 h to remove any bubbles. After the resulting bubble-free melt was cooled in air, cylindrical samples of glass were drilled from the crucible. The chemical composition of remelted dacite samples is presented in Table 1. The density of the glass was 2.48×10^3 kg/m^3.

The room temperature compressional and shear wave velocities were determined by the phase comparision method at 20 MHz using the 1 and 2 transducer technique of Jackson et al. (1981). The p-wave velocity (v_l) was determined as 6021 ± 9 m s^{-1} and s-wave velocity (v_t) as 3612 ± 4 m s^{-1}. The elastic moduli calculated using equations presented in Jaeger and Cook (1979) are presented in Table 3.

Rheological properties

Experimental methods

Rotational viscometry

A sample of the Mount St. Helens cryptodome grey dacite was remelted at 1600 °C and gas bubbles were expelled by stirring the melt in the crucible over 24 h. This melt, termed "remelted sample", was used for sub-
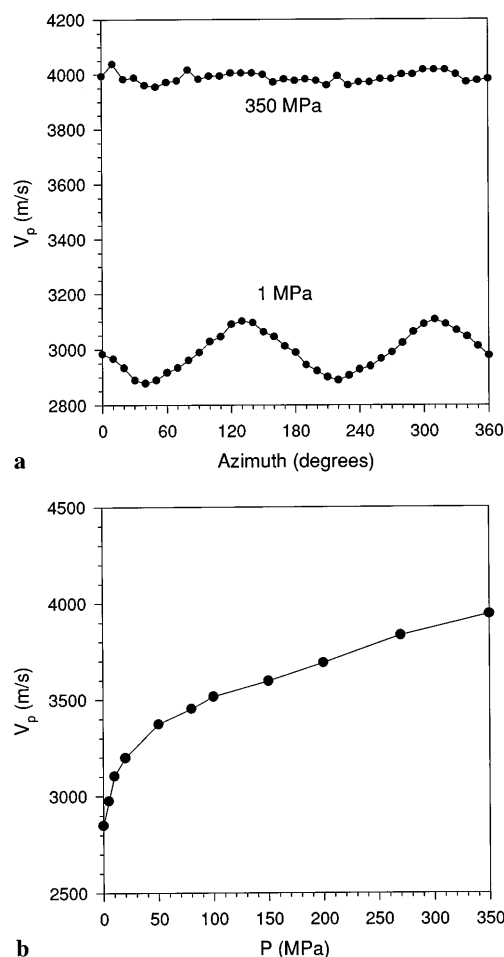
Fig. 2a, b *P*-wave velocities in raw dacite samples. **a** Azimuthal variations for confining pressures of 1 and 350 MPa; **b** dependence of the *p*-wave velocity on the confining pressure

Table 3 Elastic moduli of remelted samples of Mount St. Helens grey dacite

Moduli	Value
Longitudinal modulus $L = \rho v_l^2$	89.91×10^9 Pa $= 89.91$ GPa
Shear modulus $G = \rho v_t^2$	32.35×10^9 Pa $= 32.35$ GPa
Bulk modulus $K = L - 4/3G = \rho(3v_l^2 - 4v_t^2)/3$	46.78×10^9 Pa $= 46.78$ GPa
Young's modulus $E = \rho v_t^2(3v_l^2 - 4v_t^2)/3(v_l^2 - v_t^2)$ $= (9KG)/(3K+G)$	78.87×10^9 Pa $= 78.87$ GPa
Poisson ratio $\mu = \frac{1}{2} - v_t^2/(2v_l^2 - v_t^2)) = E/(2G) - 1$	0.2190

sequent viscosity measurements. The shear viscosity of the melt was measured in the temperature range 1300–1600 °C in air, using the concentric cylinder method (Dingwell 1989a, b). The viscometry furnace is a box furnace with $MoSi_2$ heating elements. The viscometry

crucible and spindle, fabricated from $Pt_{80}Rh_{20}$, are identical to those described by Dingwell and Virgo (1988). The high viscosity spindle from that study was employed in the present measurements. The viscometer heads are Brookfield models HBTD and RVTD (with full-scale torques of 5.75×10^{-2} N m and 7.20×10^{-2} N m, respectively). The measurements of viscosity are relative. The values of torque are converted to viscosity values based on a family of viscosity-torque calibration curves obtained from viscosity measurements of "Deutsche Glastechnische Gesellschaft" standard DGG1. The precision of these measurements is $\pm 2.5\%$ at 2σ. The calibration using viscosity standards yields an accuracy of $\pm 5\%$ at 2σ.

The rotation rates vary with the spindle and viscosity of the sample between 0.5 and 100 rpm. In general, the method employed here results in higher shear strain rates for the determination of lower viscosity data. For any individual determination, a range of approximately 0.67 orders of magnitude in rotation speed, and therefore shear strain rate, is accessible within a reasonable error range. No shear strain rate dependence of viscosity was observed. The Newtonian viscosities resulting from this observation are consistent with the proposal that the time scale of structural relaxation for melts of these viscosities is much shorter than the inverse of the experimental strain rate. Thus, we are measuring what may be usefully defined as liquid, rather than glassy, properties.

Measurements were made starting with the highest temperatures and reducing the control temperature in steps of 50 °C until an instrumental or sample limit was reached. The last measurement was always a re-occupation of the highest temperature determination. No difference was observed between the first and last viscosity determinations, indicating that no compositional or instrumental drift took place during the viscometry determination.

Micropenetration method

The low-temperature viscosities (735–800 °C) of melt were measured using the micropenetration technique of Hess et al. (1995). This involves determining the rate at which an Ir indenter under a fixed load moves into the melt surface. These measurements were performed in a BÄHR DIL 802 V vertical push-rod dilatometer. The sample was placed in a silica sample holder under an Ar gas flow. The intender is attached to one end of an alumina rod, attached at the other end to a weight pan. The metal connection between the alumina rod and the weight pan acts as the core of a calibrated linear voltage displacement transducer (LVDT). The movement of this metal core as the indenter is pushed into the melt yields the displacement.

The present system uses a hemispherical Ir indenter with a diameter of 2 mm and a force of 1.2 N. The absolute shear viscosity is determined from

$$\eta = \frac{0.1875 \, Pt}{r^{0.5} \, \alpha^{1.5}} \tag{1}$$

(Pocklington 1940; Tobolsky and Taylor 1963) for the radius of the hemisphere r, the applied force P, indent distance α, and time t ($t=0$, $\alpha=0$ upon application of the force). The measurements are performed over indentation distances of ~ 100 μm. Viscosities determined on the NBS (NIST) standard SRM 711 are reproduced within an error of ± 0.06 log units (Hess et al. 1995). Cylindrical slabs of remelted dacite with a thickness of 2–3 mm and diameter of 5 mm were used in experiments.

Parallel-plate viscometry

The methodology employed has been described previously (Bagdassarov and Dingwell 1992, 1993; Stevenson et al. 1995). Cylindrical samples of cryptodome (not remelted) vesicular grey dacite (8 mm in diameter and 7–10 mm in length) were cored from a grey dacite cobble using diamond tools and the ends of the cylinders were ground flat and parallel. Samples were washed in water using an ultrasonic bath and dried overnight at 120 °C. The dimensions of the cylindrical samples before and after experiments were measured and their volume and cumulative deformation was calculated.

A commercial dilatometer (Model TMA 402®, Netzsch Gerätebau, Selb, Germany) was used for several sets of experiments. Experiments were initially performed at low load to measure the thermal expansion coefficient and the glass transition temperature (T_g). The sample was heated at a rate 10 °C/min and sample expansion was determined as a linear temperature expansion coefficient converted to volume expansion. For another set of experiments, samples were held at a constant temperature at negligible load during 8 h. In such experiments we observed shrinkage effects in the cryptodome dacite samples.

The dilatometer was adapted to a parallel-plate geometry for viscometry measurements in the temperature range 900–990 °C. The sample was placed between two polished alumina plates and deformed perpendicular to the cylindrical end surfaces using stresses and strain rates described below. Strain rates were measured by recording the change in length of a sample as a function of time upon application of a series of stresses. After heating each sample at a rate of 10 °C/min up to the experimental temperature of interest, it was stabilized for approximately 1 h at a constant temperature under no load. The stresses applied to samples in experiments were in the range of 10^5–10^6 Pa. A loading and unloading procedure was used for viscosity measurements in these dacite samples. After the stepwise application of each load the compression of the cylinder was recorded at 30-s intervals for 25 min. The measured strain rate varied in the range of 10^{-7}–10^{-6} s^{-1} for the vesicular grey dacite samples.

Owing to volumetrically minor changes as a result of shrinkage (discussed later), as a first approximation, the shear viscosity was calculated with the following expression:

$$\eta_s = \frac{2\pi M g h^5}{3V \cdot (2\pi h^3 + V)} \qquad (2)$$

where g is the acceleration due to gravity, V is the sample volume, h is sample height, M is applied mass, $\dot{}$ is the strain rate (Bagdassarov and Dingwell 1992).

Experimental results

Melt viscosity

Measurements of viscosity on remelted samples in the temperature range 1300–1600 °C by rotational viscometry demonstrated that viscosity was in the range $10^{1.66}$ Pa s–$10^{3.22}$ Pa s (Table 4). The melt viscosity determination by the micropenetration technique in the temperature range 735–800 °C yielded viscosities in the range of $10^{9.25}$–$10^{10.60}$ Pa s. These viscosity data are also presented in Table 4. The high and low temperature data for the remelted samples are presented in Fig. 3. Collectively, these data have been fitted to the Tamann-Vogel-Fulcher (TVF) equation:

$$\log_{10}\eta = a + b/(T - c) \qquad (3)$$

to describe the temperature dependence of viscosity. The TVF coefficients resulting from the fit are A = -5.0784 ± 0.3947, B = 10009 ± 791.6, C = 372.45 ± 35.25.

Cryptodome dacite viscosity

Parallel-plate measurements on cryptodome dacite samples were performed in the temperature range of 900–982 °C. Viscosity was determined to be in the range $10^{9.94}$–$10^{10.82}$ Pa s (Table 4). These data are presented as a function of temperature in Fig. 4. Viscosity determinations have a high standard deviation owing to experiments being conducted on a crystal-rich sample near the upper limit of this technique. For the higher-temperature samples, a Newtonian stress–strain rate dependence was observed, whereas for the low-temperature measurements some hysteresis during the loading and unloading procedure was observed. At high viscosities and low loads, this hysteresis could be attributable to the mechanical system of the viscometer (Bagdassarov and Dingwell 1992). The standard error of this method is calculated as 0.08–0.14 $\log_{10}$ Pa s. In addition, the sample was crystal rich and texturally complex, and experiments under higher stresses and temperatures were not possible. As mentioned previously, viscosity calculations were performed using an expression for incompressible melts, as the crystal content was high, and

Table 4 Viscosity data for remelted and natural samples of Mount St. Helens grey dacite

T (°C)	$1/T \times 10^{-4}$ (K^{-1})	η ($\log_{10}$ Pa s)	Note
1298	6.36	3.22	a
1347	6.17	2.91	a
1397	5.99	2.61	a
1446	5.82	2.35	a
1495	5.65	2.11	a
1544	5.50	1.87	a
1594	5.35	1.66	a
735.9	9.91	10.60	b
750.3	9.77	10.25	b
764.5	9.64	10.00	b
783.9	9.46	9.53	b
802.7	9.30	9.25	b
900.4	8.52	10.82	c
924.0	8.35	10.64	c
943.3	8.22	10.27	c
962.5	8.09	10.20	c
982.4	7.97	9.94	c

a = rational viscosity data, remelted sample; b = micropenetration viscosity data, remelted sample; c = parallel-plate viscosity data, natural sample
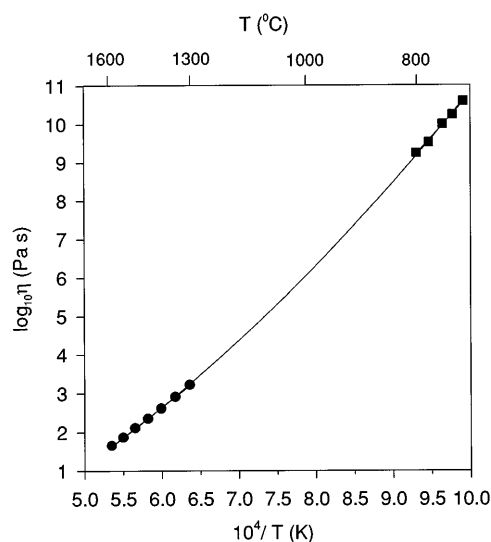

Fig. 3 Combined (rotational viscometry (circles) + micropenetration method (squares)) data on viscosity vs reciprocal absolute temperature (*squares*) for remelted grey dacite

any shrinkage effects were considered volumetrically minor.

Over this measured viscosity range, the $\log_{10}$ viscosity–inverse temperature relationship for the dacite sample is Arrhenian, described by:

$$\log_{10}\eta_s = \log_{10} A_\eta + \frac{E_\eta}{2.303\,RT} \qquad (4)$$

where η_s is the viscosity at temperature $T(K)$, and A_η and E_η are the pre-exponential factor and the activation energy of viscous flow, respectively, and R is the gas constant. The fitted values of the coefficients are
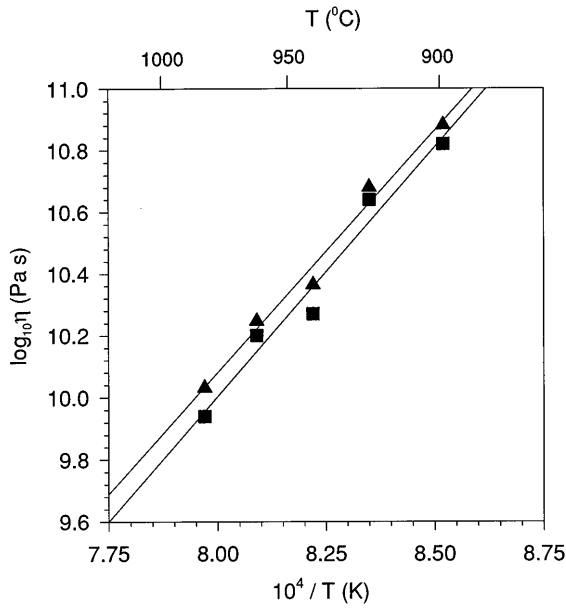
Fig. 4 Parallel-plate viscometry data vs reciprocal absolute temperature uncorrected (*squares*) and corrected (*triangles*) for the shrinkage effect for cryptodome grey dacite samples

$\log_{10} A_\eta = -2.823 \pm 0.072$ and $E_\eta = 306.88 \pm 31.68$ kJ/mol.

Discussion

Shrinkage effect on viscosity of cryptodome dacite

The experimental data presented above (Fig. 4; Table 4) were obtained by the application of standard procedures used to calculate viscosity directly using strain rate–stress Eq. (2). However, the samples of dacite at temperatures above 900 °C exhibit decreasing volumes, an effect also observed by Westrich and Eichelberger (1994). In our experiments the change in sample length reached a maximum of 3% for samples held at 970 °C over 8 h.

It is necessary therefore to estimate the effect of shrinkage on viscosity measurements. The total strain rates $\dot{\epsilon}_t$ can be related to corrected strain rates $\dot{\epsilon}_c$ (without shrinkage effect) through a term $\dot{\epsilon}_{sh}$ for the shrinkage without any load:

$$\dot{\epsilon}_t = \dot{\epsilon}_c + \dot{\epsilon}_{sh} \tag{5}$$

Equation (2) may be replaced to account for the shrinkage effect by replacing the strain rate $\dot{\epsilon}_t$ used for viscosity calculations by the corrected strain rate $\dot{\epsilon}_c$:

$$\dot{\epsilon}_c = \dot{\epsilon}_t - \dot{\epsilon}_{sh} \tag{6}$$

Concerning the loading segment the parameters M, h and V can be considered the same as for the case of the "real" viscosity (without the shrinkage effect). The real viscosity is related to the effective viscosity determined above using:

$$\eta_r \approx \frac{\eta_t}{1 - \frac{\dot{\epsilon}_{sh}}{\dot{\epsilon}_t}} \tag{7}$$

The logarithmic viscosities are related by:

$$\log_{10} \nu_\eta \approx \log_{10} \eta_t - \log_{10}\left(1 - \frac{\dot{\epsilon}_{sh}}{\dot{\epsilon}_t}\right) \tag{8}$$

The results of these calculations presented in Fig. 4 demonstrate that the shrinkage effect on viscosity of cryptodome dacite is less than 0.1 log units.

Effect of crystals and bubbles on the cryptodome grey dacite viscosity

The apparent viscosity of cryptodome dacite samples is a function of the viscosity of matrix melt as well as the crystallinity and vesicularity of samples. The viscosity of the matrix melt itself is a function of the matrix glass composition and the amount of water available in the melt. For such complex textures both of these values can be determined only with considerable uncertainty. As for the composition of the matrix melt, it is worth noting that all electron microprobe analyses on the groundmass of the cryptodome dacite are mixed owing to microlite-rich nature of the matrix. Therefore, the analysis with the highest SiO_2 content was chosen as our best approximation to the melt composition (see Table 1).

The water content of the matrix melt is the bulk value of 0.6 wt.% (presented in Table 2) obtained by the Karl-Fisher titration method, corrected to take into account the crystallinity of samples. Assuming that the sample only has 30 vol.% phenocrysts, and that one sixth of these phenocrysts are hornblende (i.e. 5 vol.%), we can estimate the water content of matrix melt as ≈ 1 wt.%. This estimate should be considered as the upper limit as it is very difficult to take into account the possible decrease of water content of matrix melt during the viscosity measurements at $T \geq 900$ °C. It should, however, be noted that we did not observe any additional vesiculation of samples during measurements implying the absence of significant degassing of the matrix melt. This corroborates our experience of viscosity measurements of other highly viscous water-bearing melts (Dingwell et al. 1996).

The calculated viscosity of matrix glass of rhyolitic composition (Table 1) obtained using the method of Hess and Dingwell (1996) for the case of water content of the matrix melt 1 wt.% are plotted in Fig. 5 together with viscosity data for the cryptodome dacite.

The phenocrysts and microlites increase the viscosity of the dacite compared with the calculated viscosities of the matrix glass (Lejeune and Richet 1995; Pinkerton and Stevenson 1992; Ryerson et al. 1988). There are
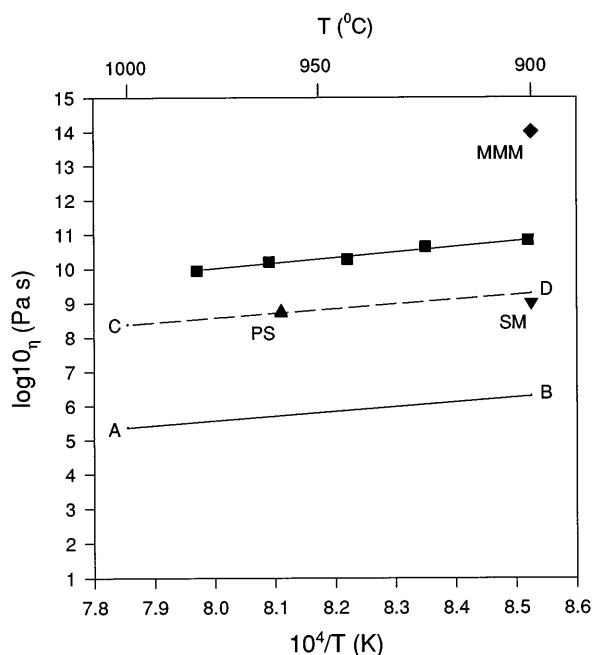
Fig. 5 Comparison of measured (*squares*) viscosities of crypto-dome dacite, calculated viscosities and those estimated by other investigators for other samples of Mount St. Helens lava. Calculation of the viscosity of the crystal-free matrix melt with 1 wt.% water (curve AB) has been performed using method of Hess and Dingwell (1996; glass matrix composition; see Table 1). Viscosity of the matrix melt + crystals (*curve CD*) has been calculated using the method of Pinkerton and Stevenson (1993). Here are also presented the estimates of viscosity of the 1980 Mount St. Helens lavas from Pinkerton and Stevenson (1993), *PS*, Scandone and Malone (1985), *SM* and Murase et al. (1985) *MMM*

several factors (crystal fraction, size distributions, shape) which can influence the effects of crystals, but quantifying these effects is difficult. The method of viscosity calculation of crystal-bearing magmas suggested by Pinkerton and Stevenson (1992) seems to be the best. Using the appropriate equation (Pinkerton and Stevenson 1992) the relative viscosity η_r can be calculated as

$$\eta_r = \eta_s/\eta_l = \left(\frac{\phi_{\max}}{\phi_{\max} - \phi}\right)^{2.5} \qquad (9)$$

where η_s and η_l are the viscosity of suspension and melt phase respectively, ϕ is the volume concentration of crystals, and ϕ_{max} is the maximum packing concentration of crystals. This is higher when we have large phenocrysts together with small microlites between the phenocrysts.

We performed our experiments at relatively low strain rates (10^{-6} s^{-1}). The mean phenocryst width was 210 μm. On a vesicle-free basis, phenocryst contents, microlite contents and melt contents were 30, 45 and 25 vol.%, respectively. We can consider three different cases of crystallinity for which different values of ϕ_{max} can be used (Pinkerton and Stevenson 1992): (a) 30 vol.% phenocrysts only (assuming $\phi_{max}=0.6$); (b) 50 vol.% crystals only (assuming $\phi_{max}=0.6$); (c) 75 vol.% crystals (phenocrysts + microlites; assuming $\phi_{max}=0.6$). The relative viscosities, which reflect the physical effect of the crystals on the viscosity of the suspension, are as follows for the different cases above: (a) $\log\eta_r=0.75$; (b) $\log\eta_r=1.94$; and (c) $\log\eta_r=3.0$. Based on these rough calculations, we can conclude that the effect of crystallinity of the Mount St. Helens dacite is to increase the apparent viscosity of the magma by 3 log units. But even in this case the increased calculated viscosity of magma having matrix melt with 1.0 wt.% H$_2$O will still be more than 1.5 log units lower than the experimentally measured viscosity (Fig. 5).

The effect of vesicles on the viscosity of matrix melt is much more difficult to predict because of the absence of reliable algorithms. For vesicular highly viscous melts containing no crystals at small strains (10^{-3}) and strain rates 10^{-5} to 10^{-7} s^{-1} a decrease of melt viscosity with vesicularity increase has been observed in parallel plate viscometry experiments (Bagdassarov and Dingwell 1992). In contrast, for the case of dilute emulsions of inviscid bubbles in melts, the shear viscosity increases with bubble content and exibits a power-law dependence (Stein and Spera 1992; Dingwell et al. 1993).

The comparison of our experimental data on the cryptodome dacite viscosity with calculated viscosities (Fig. 5) does not give an unequivocal answer as to the main influences on viscosity. Assuming a constant water content of the matrix melt we suggest that the observed difference between measured and calculated viscosity of the crystal suspension (Fig. 5) is due to the effect of vesicles.

It is instructive to compare our results with previously published estimates of the viscosity of Mount St. Helens magma for different stages of the 1980 eruption. Earlier experimental determination of the viscosity of samples of Mount St. Helens dome lava using the bending-beam method at $T<1150\,°C$ gave a viscosity $\eta \approx 10^{14}$ Pa s at $T=900\,°C$ (Murase et al. 1985). These viscosities, three orders of magnitude higher than our experimental data, can be explained by a different sample crystallinity (ca. 60%) used by Murase et al. (1985) and, perhaps more importantly, lower strain rates (10^{-10} s^{-1}; Pinkerton and Stevenson 1992). Scandone and Malone (1985) calculated the viscosity of crystal-free magma at $T=900\,°C$ to be $\eta \approx 10^6$ Pa s and suggested that the presence of 30% of crystals could increase the viscosity by three orders of magnitude (i.e. up to $\eta \approx 10^9$ Pa s). Pinkerton and Stevenson (1992) used a more rigorous method to calculate the effects of 39 vol.% of crystals. Their values, based on an eruptive temperature 960 °C range from 4.98 10^8 Pa s (for 0.4 wt.% water) to 13.7 10^8 Pa s (for 0.1 wt.% water). There are also estimates of the viscosity of the Mount St. Helens dome lava erupted after the 18 May 1980 blast based on field observations which give the viscosities of lava flows 4.8 10^8–4.2 10^9 Pa s (Chadwick et al. 1988). The published viscosity estimates are mostly

lower than our measurements of the cryptodome dacite viscosity; the only exception is the estimate of Murase et al. (1985) which is higher than our data (Fig. 5).

Implications for viscosity estimates of the 1980 Mount St. Helens cryptodome magma

For modeling the lateral blast of 18 May 1980 it is important to know rheological properties of the cryptodome magma immediately before the blast. The temperature of the 1980 Mount St. Helens cryptodome magma before the 18 May lateral blast has been estimated to be 900 °C and the characteristic lithostatic pressure has been estimated to be approximately 20 MPa (Eichelberger and Hayes 1982). The crystallinity of cryptodome magma could not change during the blast and from this point of view our samples used for viscosity measurements should be considered identical to "original" cryptodome magma. But there are two main questions which cannot be answered so easily: (a) What was the water content of the cryptodome magma? and (b) What was the vesicularity of cryptodome magma immediately before the blast? Data on water content of ejected fragments of cryptodome show water concentrations less than 1 wt.% and for most of samples of black and grey dacite the range is 0.2–0.5 wt.% H_2O (Hoblitt and Harmon 1993). Our measurements of the bulk water content gave 0.6 wt.%. The initial water content of dacite magma of the 1980 Mount St. Helens eruption has been estimated by Rutherford and Devine (1988) to be approximately 4.6 wt.%. Was this amount of water preserved in the cryptodome magma which has been intruded slowly enough before the 18 May 1980 blast? Perhaps, but a more reasonable scenario seems to us to be the significant degassing of magma during its ascent and formation of the cryptodome. For the lithostatic pressure of 20 MPa the equilibrium amount of water which could be dissolved in the melt can be estimated at ~1.5 wt.% (Burnham 1979). The rest of the water (~3 wt.%) had to exsolve during the cryptodome formation. Thus, the samples used here, with 0.6 wt.% bulk water content and an estimated 1 wt.% water content in the melt matrix, seem to have an appropriate amount of water for those conditions.

As a result of significant water exsolution during relatively slow decompression of magma to pressures ≤ 20 MPa, vesiculation of cryptodome magma during the days preceding the 18 May 1980 blast seems probable. This suggestion is also supported by our observation of a quite homogeneous vesicularity of large grey dacite clasts with size ~0.5 m. If the vesiculation occurred mainly during the decompression–induced blast event, we might expect to observe a significant elongation of vesicles and a more inhomogeneous distribution. In the case of pre-blast vesiculation of the highly viscous cryptodome magma the probable scenario of the development of the 18 May 1980 blast might well be similar to that suggested by Alidibirov (1995). Howev-

er, it should be noted that the presence of breadcrust bombs suggests that some additional vesiculation of some cryptodome magma took place. This phenomena can be explained by the expansion of pre-existing vesicles after their decompression during quite long flight time of the bombs. It seems clear that future experiments on rapid decompression of water-bearing magmas could clarify this problem.

At a bulk water content of 0.6 wt.% and $T = 900$ °C the viscosity of cryptodome magma can be estimated from our experimental data as $10^{10.8}$ Pa s. This estimate confirms the proposal of a highly viscous state for the cryptodome magma before the 18 May 1980 lateral blast (Hoblitt et al. 1981). The different vesicularities of dacite ejected during the blast (Hoblitt and Harmon 1993) can be related to different vesicularities of magma inside the cryptodome. Different lithostatic pressure inside cryptodome could cause variations in magma–water content. There could also be variations of temperature inside the cryptodome: the outer parts could be cooler than the interior of the cryptodome. All these factors could change the viscosity of the cryptodome magma and yield magma body with high but spatially variable viscosity.

Conclusions

The following conclusions were reached as a result of this study:

1. Mount St. Helens dacite from the 18 May 1980 eruption was characterized using different techniques. Data on density, vesicularity and specific surface area of vesicular samples of grey dacite were obtained. The sound speed measurements of natural and remelted dacite samples ranged from $v_p = 3000$–4000 m/s for cryptodome dacite samples and $v_p = 6021 \pm 9$ m/s and $v_s = 3612 \pm 4$ m/s for remelted samples. These data are important for analysis of the cryptodome magma response during the 18 May 1980 blast.

2. Rheological properties of Mount St. Helens dacite (cryptodome and remelted dacite samples) were investigated. The viscosity of vesicular cryptodome dacite samples measured by the parallel-plate viscometry method lies in the range $10^{9.94}$–$10^{10.82}$ Pa s for the temperature range 900–982 °C. Viscosities of the remelted samples determined by rotational viscometry and the micropenetration method lie in the range $10^{1.66}$–$10^{10.60}$ Pa s for temperatures in the range 736–1594 °C.

3. Assuming that the grey dacite vesicularity is representative of the 1980 Mount St. Helens cryptodome magma prior to fragmentation on 18 May 1980, the corresponding viscosity should be approximately $10^{10.8}$ Pa s. The high viscosity of the cryptodome magma results from a combination of the low water and high silica content of the melt as well as the high values of crystallinity and vesicularity.

References

Alidibirov MA (1995) A model for the mechanism of the May 18, 1980 Mount St. Helens blast. J Volcanol Geotherm Res 66:217–225

Alidibirov M, Dingwell DB (1996) Magma fragmentation by rapid decompression. Nature 380:146–149

Bagdassarov NS, Dingwell DB (1992) A rheological investigation of vesicular rhyolite. J Volcanol Geotherm Res 50:307–322

Bagdassarov NS, Dingwell DB (1993) Deformation of foamed rhyolites under internal and external stresses: an experimental investigation. Bull Volcanol 55:147–154

Behrens H (1995) Determination of water solubilities in high-viscosity melts: an experimental study of $NaAlSi_3O_8$ and $KAlSi_3O_8$ melts. Eur J Mineral 7:905–920

Burnham CW (1979) Magmas and hydrothermal fluids. In: Barnes HL (ed) Geochemistry of hydrothermal ore deposits. Wiley, New York, pp 71–136

Cashman KV (1988) Crystallization of Mount St. Helens 1980–1986 dacite: a quantitative textural approach. Bull Volcanol 50:194–209

Cashman KV (1992) Groundmass crystallization of Mount St. Helens dacite, 1980–1986: a tool for interpreting shallow magmatic processes. Contrib Mineral Petrol 109:431–449

Chadwick WW, Archuleta RJ, Swanson DA (1988) The mechanics of ground deformation precursory to dome-building extrusions at Mount St. Helens 1981–1982. J Geophys Res 93:4351–4366

Dingwell DB (1989a) Effect of fluorine on the viscosity of diopside liquid. Am Mineral 74:333–338

Dingwell DB (1989b) Shear viscosities of ferrosilicate liquids. Am Mineral 74:1038–1044

Dingwell DB, Virgo D (1988) Melt viscosities in the Na_2O–FeO–Fe_2O_3–SiO_2 system and factors controlling the relative viscosities of fully polymerized silicate melts. Geochem Cosmochim Acta 52:395–403

Dingwell DB, Bagdassarov NS, Bussod GY, Webb SL (1993) Magma Rheology. MAC Short Course Handbook 21:131–196

Dingwell DB, Romano C, Hess K-U (1996) The effect of water on the viscosity of a haplogranitic melt under P-T-X conditions relevant to silicic volcanism. Contrib Mineral Petrol 124:19–28

Eichelberger JC, Hayes DB (1982) Magmatic model for the Mount St. Helens blast of May 18, 1980. J Geophys Res 87:7727–7738

Hess K-U, Dingwell DB, Webb SL (1995) The influence of excess alkalis on the viscosity of a haplogranitic melt. Am Mineral 80:297–304

Hess K-U, Dingwell DB (1996) Viscosities of hydrous leucogranitic melts: a non-Arrhenian model. Am Mineral 81:1297–1300

Hoblitt RP, Miller CD, Vallance JW (1981) Origin and stratigraphy of the deposit produced by the May 18 directed blast. US Geol Surv Prof Pap 1250:401–419

Hoblitt RP, Harmon RS (1993) Bimodal density distribution of cryptodome dacite from the 1980 eruption of Mount St. Helens, Washington. Bull Volcanol 55:421–437

Jackson I, Niesler H, Weidner DJ (1981) Explicit correction of ultrasonically determined elastic wave velocities for transducer-bond phase shift. J Geophys Res 86:3736–3748

Jaeger JC, Cook N (1979) Fundamentals of rock mechanics. Chapman and Hall, London

Kieffer SW (1981) Fluid dynamics of the May 18 blast at Mount St. Helens. US Geol Surv Prof Pap 1250:379–400

Lejeune A-M, Richet P (1995) Rheology of crystal-bearing silicate melts: an experimental study at high viscosities. J Geophys Res 100:4215–4229

Murase T, McBirney AR, Melson WG (1985) Viscosity of the dome of Mount St. Helens. J Volcanol Geotherm Res 24:193–204

Pinkerton H, Stevenson RJ (1992) Methods of determining the rheological properties of magmas at sub-liquidus temperatures. J Volcanol Geotherm Res 53:47–66

Pocklington HC (1940) Rough measurement of high viscosities. Proceedings of the Cambridge Philosophical Society 36:507–508

Rutherford MJ, Devine JD (1988) The May 18, 1980, eruption of Mount St. Helens, 3. Stability and chemistry of amphibole in the magma chamber. J Geophys Res 93:11949–11959

Ryerson FJ, Weed HC, Piwinskii AJ (1988) Rheology of subliquidus magmas. I. Picritic compositions. J Geophys Res 93:3421–3436

Scandone R, Malone SD (1985) Magma supply, magma discharge and readjustment of the feeding system of Mount St. Helens during 1980. J Volcanol Geotherm Res 23:239–262

Shaw H (1972) Viscosities of magmatic silicate liquids. Am J Sci 272:870–893

Sparks RSJ, Moore JG, Rice CJ (1986) The initial giant umbrella cloud of the May 18th, 1980, explosive eruption of Mount St. Helens. J Volcanol Geotherm Res 28:257–274

Stein DJ, Spera FJ (1992) Rheology and microstructure of magmatic emulsions: theory and experiments. J Volcanol Geotherm Res 49:157–174

Stevenson RJ, Dingwell DB, Webb SL, Bagdassarov NS (1995) The equivalence of enthalpy and shear stress relaxation in rhyolitic obsidians and quantification of the liquid–glass transition in volcanic processes. J Volcanol Geotherm Res 68:297–306

Stevenson RJ, Dingwell DB, Webb SL, Sharp T. The viscosity of microlite-bearing obsidians: an experimental study. Bull Volcanol 58:298–309

Tobolsky AV, Taylor RB (1963) Viscoelastic properties of a simple organic glass. J Phys Chem 67:2439–2442

Voight B (1981) Time scale for the first moments of the May 18, eruption. US Geol Surv Prof Pap 1250:69–86

Voight B, Glicken H, Janda RJ, Douglass PM (1981) Catastrophic rockslide avalanche of May 18. US Geol Surv Prof Pap 1250:347–377

Webb SL, Knoche R, Dingwell DB (1992) Determination of silicate liquid thermal expansivity using dilatometry and calorimetry. Eur J Mineral 4:95–104

Westrich HR, Eichelberger JC (1994) Gas transport and bubble collapse in rhyolitic magma: an experimental approach. Bull Volcanol 56:447–458

Zang A, Lienert M, Zinke J, Berckhemer H (in press) Residual strain, wave speed, and crack analysis of crystalline cores from the KTB-VB well. Tectonophysics

Zinke J (1996) Mikrorissuntersuchungen an KTB-Bohrekernen-Beziehungen zu den elastischen Gesteinsparametern. Frankfurter Geowissenschaftliche Arbeiten, A13, 195 pp, Frankfurt am Main, Germany