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RHEOLOGICAL PROPERTIES OF MAGMAS

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INTRODUCTION

Recent studies of igneous rocks have taken a refreshing new direction, mainly as the result of a greater awareness of the important role played by physical properties of magmas in determining the eruptive behavior and compositional variations of volcanic rocks. This review summarizes the present state of knowledge, along with some of the recent rheological studies having a direct bearing on interpretations of volcanic phenomena and processes of crystallization and differentiation of shallow magmatic intrusions.

The rheological properties of magmas reflect the inherent structures of molten silicates and will not be thoroughly understood until more is known about the basic nature of silicate liquids. Recent progress in this field has been ably reviewed by Hess (1980) and needs no elaboration here aside from mention of the obvious need for better laboratory measurements in order to place greater constraints on interpretations of silicate melts. For petrogenetic and volcanological processes, the most conspicuous need is for improved data on rheological properties, chiefly viscosity and yield strength, and for more precise equations for predicting these properties under a wide range of natural conditions.

Viscosity, η, is usually defined as the ratio of shear stress to strain rate and is expressed in units of poise (g cm^{-1} s^{-1}) or Pascal seconds (10 poise = 1.0 Pa s). In more general terms, it is the coefficient for transfer of

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momentum in the equation

$$\tau = \tau_0 + \eta\left(\frac{du}{dy}\right)^n, \tag{1}$$

in which τ is the shear stress applied in the direction x parallel to the plane of flow, τ_0 is the minimum stress required to initiate permanent deformation, du/dy is the velocity gradient normal to the plane of shear, and n is a constant with a value of one or less. A fluid having no yield stress ($\tau_0 = 0$) and a direct linear relation between shear stress and rate of strain (i.e. $n = 1$) is said to be Newtonian. Many complex fluids, suspensions, and emulsions are non-Newtonian in the sense that the relation between strain rate and stress is nonlinear, and in some, viscous flow takes place only when the shear stress exceeds a finite value. Although, for convenience, magmas are commonly treated as Newtonian fluids, most are not. Below their liquidus, particularly when they are charged with phenocrysts, their rate of shear is not directly proportional to stress, i.e. $n < 1.0$ in Equation (1).

In 1967, Robson noted that some lavas have a finite yield strength ($\tau_0 > 0$) that must be exceeded for the lava to move under the force of gravity (Robson 1967). Non-Newtonian behavior of basaltic magma was demonstrated by Shaw et al (1968) in studies of the basalts of Hawaiian lava lakes and by Gauthier (1973) and Pinkerton & Sparks (1978) at Mount Etna. Subsequent workers (e.g. Hulme 1974, Sparks et al 1976, Moore et al 1978, Peterson & Tilling 1980, Cigolini et al 1983, Murase et al 1984) have shown how many morphological features of lavas can be explained in terms of these properties.

Yield strength is not strictly a threshold value below which a magma does not deform under any condition. Elastic deformation may result from any stress, with a value either greater or less than the yield strength. It differs from viscous flow in that it is recovered when stress is removed. Small stresses may cause permanent deformation in the form of creep. Thus, the yield strength of magma is more properly thought of as a transition value below which the dominant mechanism of deformation is not viscous flow but another process that, though slower, is still capable of producing permanent strain.

The different modes of permanent deformation within the ranges of temperature and stresses that prevail in crystallizing magmas can be illustrated by means of a "deformation map" of the kind devised by Ashby (1972). Figure 1 is a preliminary attempt to construct such a diagram on the basis of the sparse data now available (H. J. Melosh and A. R. McBirney, unpublished data). The fields for the dominant mechanisms of deformation are shown for different regimes of temperature and deviatoric stress; estimated strain rates are shown as contours crossing the fields. Such a

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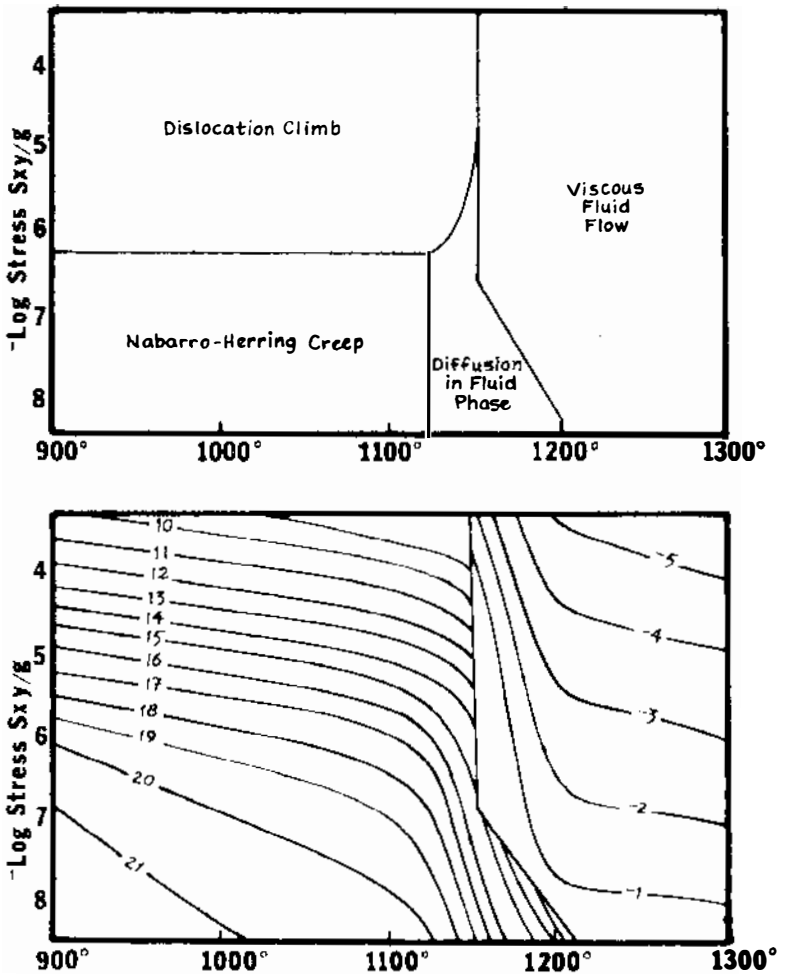

Figure 1 Calculated deformation map for a Hawaiian lava-lake basalt. Upper diagram displays the dominant mechanism of deformation in various ranges of temperature and stress. The lower diagram shows strain-rate contours in negative exponential values (cm s^{-1}). The solidus temperature is assumed to be 1100° and the liquidus 1200°C. Calculations were carried out by Jay Melosh on the basis of properties measured by T. Murase and A. R. McBirney (unpublished data). The method of calculation is similar to that of Stocker & Ashby (1973) and is designed to illustrate the forms and rates of deformation under differing deviatoric stresses over a wide range of temperatures.

diagram, if it were based on reliable data, could provide valuable insights into a number of geological processes occurring within the transition from the liquid to the solid state. Unfortunately, the details of the most interesting regions are poorly known.

In summarizing recent field and laboratory studies of the rheological properties of natural silicate melts, we treat viscosity and yield strength separately, but the two properties are closely related and are only two parts of the much broader phenomenon of deformation of igneous rocks and melts at high temperature. Although we are not concerned here with the behavior of rapidly cooled magmas and glasses, we draw attention to the important work being done in this field, most notably by Ryan and co-workers (e.g. Ryan & Sammis 1981).

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FIELD MEASUREMENTS

Viscosities have been estimated from the velocities of moving lavas by using Jeffery's equation for laminar flow of an extensive sheet on a uniform slope:

$$\eta = \frac{g\rho \sin \alpha d^2}{3V} \tag{2}$$

where ρ is the density, α is the slope angle, d is the thickness, and V is the mean velocity. For flow in restricted channels, the coefficient in the denominator is replaced by 4. This equation yields only crude values because it takes no account of the fact that most of the velocity gradient in lavas may be in a relatively thin basal layer and that viscosity can vary greatly between the base, interior, and crust of the flow.

Most, if not all, lavas move by laminar flow (Einarsson 1949, Booth & Self 1973); no well-documented examples of turbulent flow have been reported, although some have been postulated for lunar basalts. Many lavas with low viscosities are thought to have a motion similar to that of a caterpillar tread, with shear distributed throughout their entire thickness, but others appear to slide on a basal layer in which most of the shear is concentrated. The mean velocity of such a flow cannot be estimated from observations of the surface. An additional problem arises in the case of non-Newtonian lavas in which the apparent viscosity may vary with shear rate. At low velocities and shear rates, the effective viscosity may be greater than it is at high velocities because the ratio of shear rate to shear stress is not linear.

The most reliable field data for lavas are those obtained from measurements made over a range of temperatures and shear stresses in the interiors of flows. Excellent opportunities for measurements of this kind have been provided by recent eruptions of the Hawaiian volcano, Kilauea, where

basalt has ponded in deep pit craters and remained molten for periods of a decade or more. Viscosity is measured by drilling through the crust, inserting a vane-shear apparatus directly into the interior, and measuring the torque required for a constant rate of rotation (Shaw et al 1968).

Yield strength can be estimated from the angle of repose of flow fronts and the minimum slope on which lavas continue to move (Johnson 1970, Hulme 1974). Values obtained by these methods are useful for interpreting the properties of lavas long after they have come to rest, but they are subject to uncertainty owing to the difficulty of evaluating temperatures and time-dependent factors. The measurements may be relevant only to the crust and not to the body as a whole. Again, the best field data are those obtained from ponded lavas, such as those of the Hawaiian lava lakes. Yield strength can be estimated with the same apparatus used to measure viscosity. The value corresponds to the torque required to initiate rotation of a spindle inserted below the crust. More recently, Pinkerton & Sparks (1978) developed a portable instrument that can be used to obtain measurements from flowing lava. Measurements are made by determining the force required to push a cylindrical steel rod into the molten interior of the flow. Yield strengths measured in this way were found to be less than half those estimated from the morphology of flow fronts.

At best, however, the results of field measurements are subject to uncertainties arising mainly from the problem of taking into account all the various factors that affect rheological properties under natural conditions. Most serious, perhaps, are time-dependent factors, which may lead to widely differing values for both viscosity and yield strength, depending on how long the lava has been undisturbed before measurements are made.

LABORATORY MEASUREMENTS

Three methods have been used for laboratory studies of the rheological properties of natural silicate melts. Rotational viscometers of the kind used in the glass industry are satisfactory for measurements at large stresses and strain rates requiring only low precision. For lower stresses and strain rates, the counterbalanced sphere method of Shartsis & Spinner (1959) [Figure 2 (*left*)] is more satisfactory, especially if strain rates are measured electronically (Murase et al 1984). Densities can be measured simultaneously in the same experiment by determining the weight of the platinum sphere while suspended in the molten sample (MacKenzie 1956).

Use of the counterbalanced sphere technique is limited to samples prepared from powdered rock or glass and with low yield strengths and viscosities (below about 10^5 dyn cm^{-2} and 10^8 poise, respectively). Samples with more than 20 or 30% crystals and with high viscosities and yield

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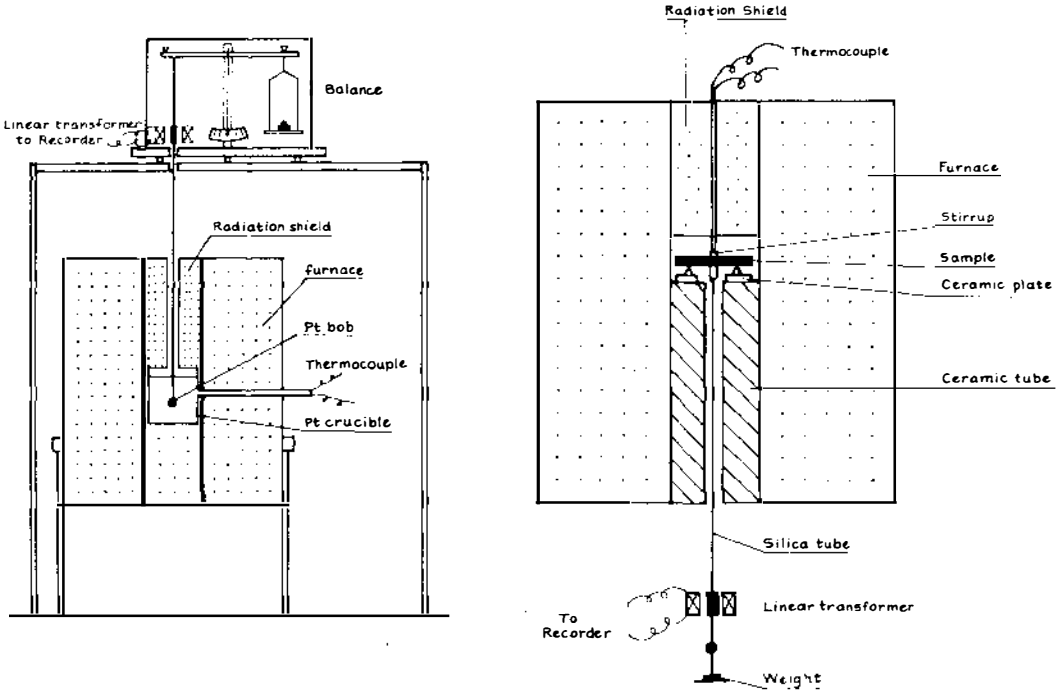

Figure 2 Two types of apparatus used for rheological measurements in the laboratory. (*Left*) In the counterbalanced sphere method, a platinum bob is suspended in the molten sample and balanced by weights on the opposing arm of an analytical balance. A platinum rod between the balance and the wire suspending the sphere in the sample reduces springlike effects in the suspension system. Displacement of the sphere is measured by a linear transformer connected through an amplifying system to a chart recorder. Corrections are made for the effect of surface tension on platinum wire by measuring the stress on the wire when suspended in the sample without a sphere attached. Loss of iron to the crucible and sphere is minimized by prior saturation of the platinum. The atmosphere of the furnace is controlled by mixtures of CO and CO_2 monitored by a zirconia cell. (*Right*) The deflected-beam method utilizes the same basic furnace to measure the bending of a slab of precisely cut rock supported between two ceramic prisms. Stress is transmitted to the center of the beam by means of a system of weights and a silica glass stirrup. As before, displacement is measured by a linear

strengths can be studied by observing the rate of bending of a precisely cut slab under a stress applied to its center [Figure 2 (*right*)].

In order to study the behavior of crystals with densities close to the liquid in which they are suspended, Roeder & Dixon (1977) developed an ingenious centrifuge furnace capable of generating forces equivalent to 700 times the acceleration of gravity.

These various techniques can be used in controlled atmospheres, but only at low pressures; more elaborate methods must be used at elevated pressures and volatile contents. Measurements of the latter type have been carried out by Shaw (1963), Burnham (1963), Carron (1969), Scarfe (1973), Kushiro et al (1976), and Khitarov et al (1978). Most of these studies have used a falling-sphere technique, in which the sample, enclosed in a small capsule, contains a platinum sphere or round mineral grain of known size and density, which is allowed to settle through the liquid while the sample is held at an elevated temperature and pressure.

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VISCOSITY

Viscosity is sensitive to a variety of factors, including composition, temperature, pressure, volatile content, abundance and size of crystals, and thermal and mechanical history of the melt. Many of these factors are interrelated, and their individual effects are not easy to unravel.

Effects of Temperature and Composition

Typical viscosities for some common igneous compositions are shown in Figure 3. As in all determinations for silicate melts above their liquidus, these viscosities have an inverse logarithmic variation with temperature according to a relation having the form

$$\eta = A_\eta \exp E_\eta/RT, \quad (3)$$

where A_η is a constant, E_η the activation energy of viscous flow, and R the gas constant. Activation energy seems to be related to the proportion of bridging silicon and aluminum atoms as expressed by the ratio (Si + Al)/O for the melt (Euler & Winkler 1957), but only at ratios greater than about 0.4 (Murase & McBirney 1973). Structural differences reflecting the proportions of bridging and nonbridging silicon and aluminum atoms seem to provide the best empirical correlation between composition and viscosity (Murase 1962), but no simple rule has been found to account for the differing effects of individual network-modifying elements, such as Mg, Fe, Ca, K, and Na, each of which tends to lower viscosity but by slightly different amounts.

Viscosities of crystal-free liquids can be calculated from the empirical

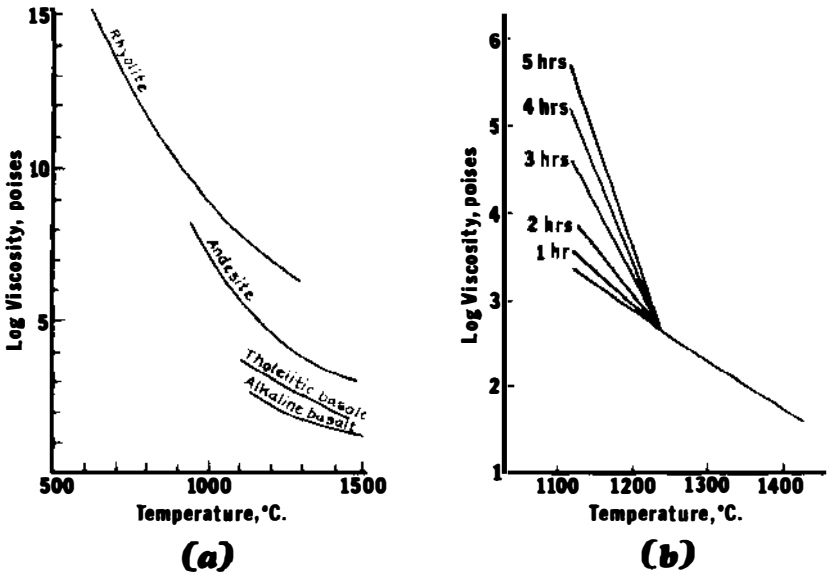

Figure 3 (*a*) Viscosities of some common igneous rocks at temperatures in and above their melting ranges (after Murase & McBirney 1973). (*b*) The viscosities shown in (*a*) are metastable below the liquidus temperatures of the respective samples. The increase that comes with time as crystals nucleate and grow is illustrated by the behavior of the same tholeiitic basalt shown in (*a*).

equations of Shaw (1972) and Bottinga & Weill (1972). Values predicted by these equations are in close accord with laboratory measurements, at least for low pressures and water contents and for temperatures above the liquidus.

Effects of Water and Load Pressure

Shaw (1963), Burnham (1963), and Carron (1969) obtained data on the effect of water on the viscosities of rhyolitic liquids, and Scarfe (1973) and Khitarov & Lebedev (1978) collected similar data for basaltic compositions (Figure 4). At constant temperature, one weight percent of water lowers the viscosity of silica-rich liquids by nearly an order of magnitude, but the effect is much less pronounced in liquids of lower silica content, probably because the polymerization of mafic melts is already low and cannot be substantially reduced by the addition of water. The effects of other volatile components, such as CO_2 and S, have yet to be explored.

Kushiro (1980) has reviewed the work that he and others have done on the effects of load pressure on the viscosity and other properties of natural silicate melts. Load pressure has been found to reduce the viscosity of basaltic and andesitic melts at constant temperatures above the liquidus. A tholeiitic basalt was found to fall from about 170 poise at atmospheric

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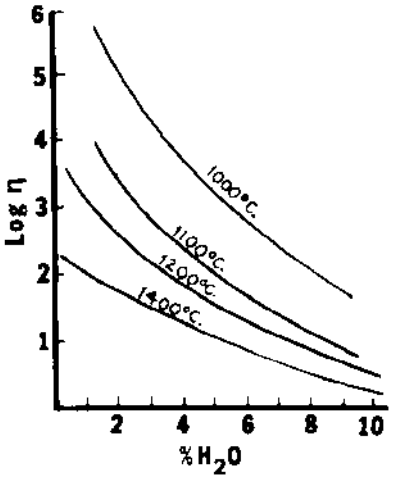

Figure 4 The effect of water on the viscosity of basaltic liquids at various temperatures (after Khitarov & Lebedev 1978).

pressure to 40, 25, and 8 poise at 15, 20, and 30 kbar, respectively. Corresponding values for an andesite from Crater Lake were about 3200 and 900 poise at 7.5 and 20 kbar (Figure 5). Since pressure also has an effect on the liquidus temperature, and since viscosity is a function of temperature

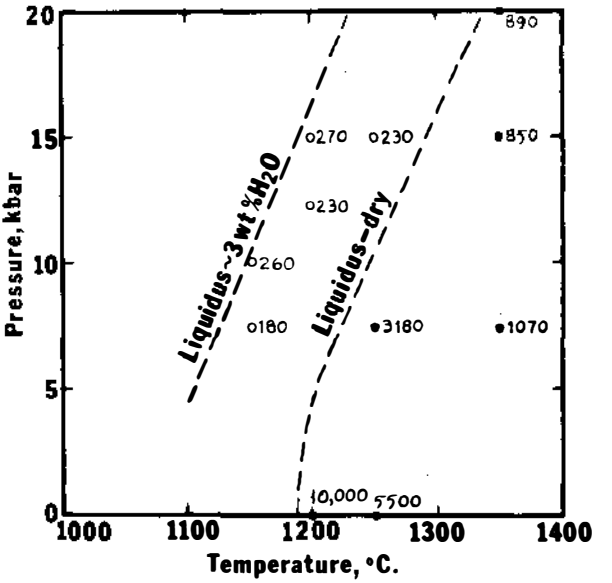

Figure 5 Viscosity of an andesite from Crater Lake, Oregon, as a function of pressure and temperature above the liquidus. Data points shown by filled circles are for water-free conditions; open circles denote samples with about 3 wt% H_2O. Corresponding liquidus curves are shown by broken lines [from data of Kushiro et al (1976) and Kushiro (1978)].

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above the liquidus, the two effects of pressure tend to offset one another. It appears likely, however, that the viscosities of melts at mantle conditions are low enough to facilitate segregation of liquids from a crystal-liquid assemblage under relatively small differential stresses.

Effects of Crystal Content

Because most magmas have temperatures below their liquidus and contain differing proportions of crystals in suspension, estimates of properties on the basis of measurements carried out at temperatures above the liquidus have only limited relevance to natural processes. A number of attempts have been made to correct for these differences. The effective viscosities of crystal-rich magmas are commonly estimated from the Einstein-Roscoe equation

$$\eta = \eta_0(1 - R\phi)^{-2.5}, \tag{4}$$

in which η is the effective viscosity of a liquid containing a volume fraction Φ of suspended solids, and η_0 is the viscosity of the liquid alone. The constant R is based on the volumetric ratio of solids at maximum packing and is usually taken as 1.35 for spheres of uniform size or as 1.0 for crystals with serial sizes. Marsh (1981) has considered the packing ratios of crystals in lavas and has concluded that a value of 1.67 is more appropriate.

The large differences between measured viscosities of crystal-rich magmas and those predicted by Equation (4) are illustrated by the data for Hawaiian tholeiitic basalts and the dacitic dome of Mount St. Helens (Figure 6). The discrepancies are attributable in part to the sizes and high concentrations of crystals, both of which exceed the range of values for which Equation (4) was derived. Sherman (1968) examined the effects of a wide range of concentrations of crystals of differing size and found that the relative viscosity increases with the mean diameter and concentration of suspended solids according to the relation

$$\ln \eta_R = \ln (\eta/\eta_0) = \frac{\alpha D_m}{\left(\frac{\Phi_{max}}{\Phi}\right)^{1/3} - 1} - 0.15, \tag{5}$$

in which η_0 is the viscosity of the crystal-free liquid, α is a constant that varies with the mean diameter D_m, Φ_{max} is the concentration of solids at maximum packing, and Φ is the concentration of suspended particles.

Equations (4) and (5) assume that the solid fraction has no physical-chemical relation to the liquid in which it is suspended. This is obviously not the case for magmas if crystals are nucleating and growing during cooling. Under such conditions, the only limit of Φ_{max} is 1.0, when the liquid

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is totally crystallized at the solidus. As Φ increases, the mean diameter D_m also changes. It does not necessarily increase with cooling; in lavas the opposite is more likely to be the case because large numbers of small crystals growing from the groundmass more than offset the slight increase in size of phenocrysts.

An even greater effect arises from the fact that the composition of the liquid is not independent of Φ but changes at an accelerating rate with progressive crystallization. Thus, as the composition of the liquid fraction changes, so too does η_0, and the magnitude of this effect may be much greater than that of the suspended crystals alone.

We know of no simple way of integrating these effects, other than by empirically matching the physical constants in Equation (5) to experimentally determined values of η_0, Φ, and D_m. If the groundmass composition of the sample is determined, either by a mass-balance calculation or direct analysis, η_0 can be calculated from the equations of Shaw (1972). The value of D_m can be estimated petrographically, and Φ can be determined in the same way or by mass balance if the compositions of all phases are known.

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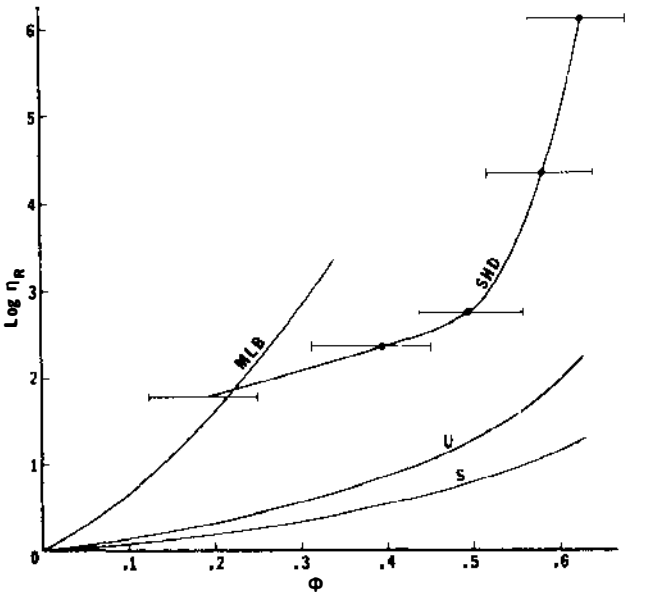

Figure 6 Effects

(MLB) and a Mount St. Helens dacite (SHD). Relative viscosity is defined as the ratio of the viscosity of the liquid-crystal suspension to that of the crystal-free liquid at the same temperature. Curves labeled U and S are calculated for spheres of uniform and serial sizes, respectively, using Equation (4) [after Shaw (1969, Figure 4) and Murase et al (1984)].

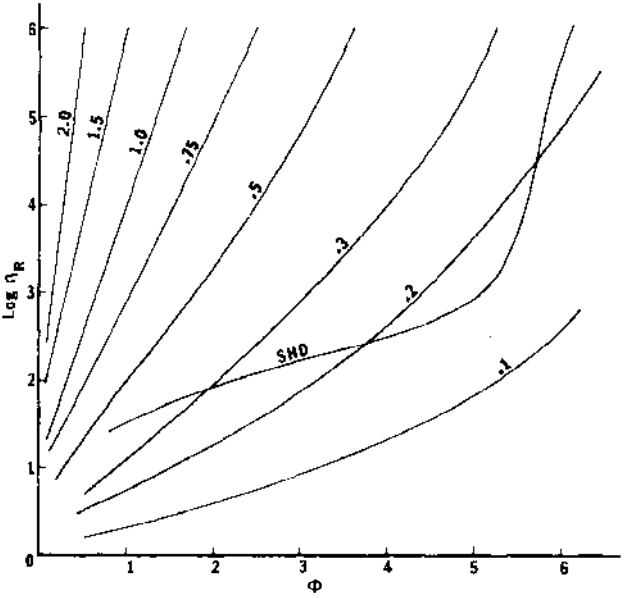

Figure 7 Effect of crystal size on the relative viscosity of a Mount St. Helens dacite. Sloping lines indicate mean diameter in millimeters. Curve labeled SHD is plotted from measured viscosities, crystal contents, and compositions of glass. It forms the basis for values used in Equation (5) and the computer program in the Appendix (data from Murase et al 1984).

Using data of these kinds for the dacite of Mount St. Helens and a Hawaiian tholeiitic basalt, we find that the best agreement with measured viscosities is obtained when $\alpha = 0.011$ and $\Phi_{max} = 1.0$. If these values are substituted into Equation (5), a graphical plot (Figure 7) can be constructed for the relative viscosity (Log η_R) as a function of the proportions (Φ) and mean diameter of crystals (in mm). A computer program for calculating effective viscosities of magmas at subliquidus temperatures is given in the Appendix.

Time-Dependent Factors

Most measurements of viscosity, either in the field or in the laboratory, show an increase of viscosity as the melt remains at subliquidus temperatures for increasing periods of time (Figure 8*a*). Most of this increase is attributable to the growth of crystals, but part may be due to the ordering and polymerization of the liquid. Samples that have remained undisturbed for many hours may revert to a lower effective viscosity if they are stirred or otherwise disturbed. In a similar way, apparent viscosities of many crystal-rich silicate liquids tend to be lower at high rates of shear. Most of this difference is due to the non-Newtonian nature of the sample, which can only be determined if rates of strain are measured at more than one rate of shear.

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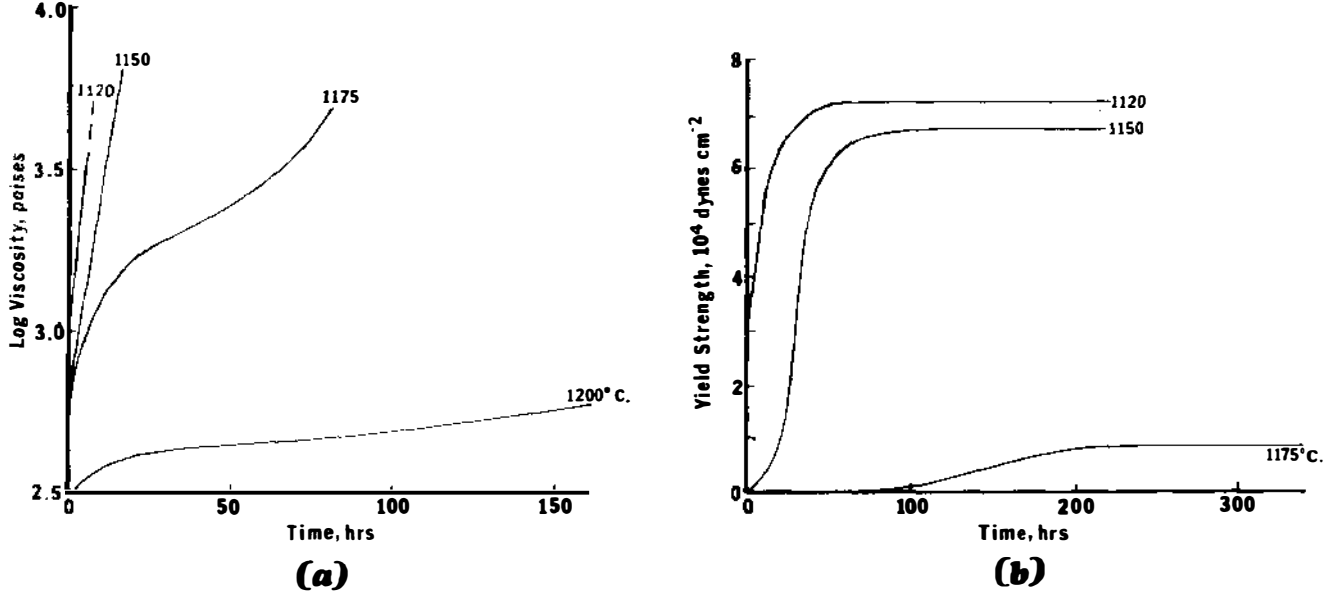

Figure 8 (*a*) A sample that was initially at a higher temperature increases in viscosity with time, as shown by the behavior a Hawaiian tholeiitic basalt from Makaopuhi lava lake. Most of the increase is due to crystallization, but a similar effect is observed after the melt is strongly stirred at constant temperature. (*b*) The same sample shows a similar increase in yield strength with time. The measurements were made in air; reliable values for yield strength have not yet been obtained under reducing conditions, but preliminary results indicate that they are much lower.

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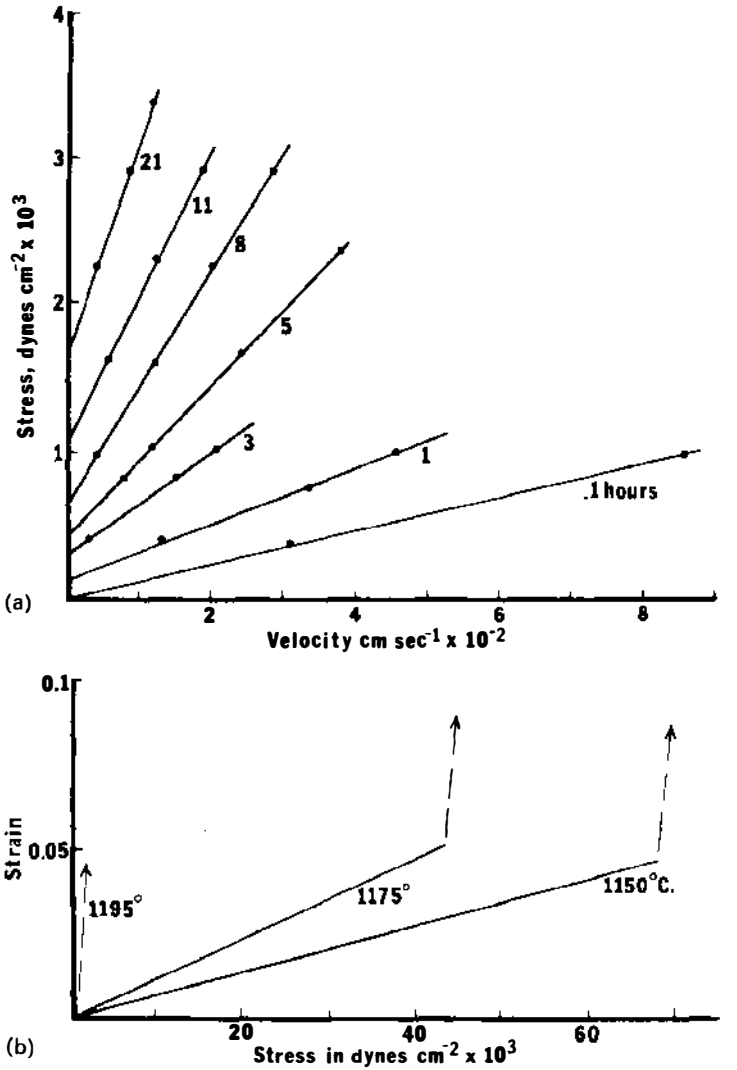

Figure 9 Yield strength can be determined by (*a*) measuring the slope of a curve for velocity of the ascending sphere as a function of the applied stress and then extrapolating to a velocity of zero to find the minimum stress at which motion begins. Alternatively, the elastic strain at low stresses (*b*) can be measured to find the stress at which the elastic limit is exceeded and permanent strain begins. In the latter method, each measurement of strain for a given stress produces a characteristic strain-time (*c*). The effect of surface tension (σ_s) is determined without a sphere attached to the wire and is subtracted to obtain the stress required to produce strain in the interior of the liquid. At low stresses (below the yield strength), elastic strain results when stress is applied, and is recovered when the stress is removed. Stresses greater than the yield strength result in viscous flow and strain that is not recovered when the stress is removed.

YIELD STRENGTH

Two methods have been used to estimate yield strength from laboratory measurements. By measuring strain rates as a function of stress and plotting one aginst the other (Figure 9*a*), one can obtain a curve that extrapolates to the stress axis to give the minimum stress required to produce measurable viscous flow. This intercept corresponds to τ_0 in Equation (1). In the second method, strain is measured as a function of stress (Figure 9*b*). Elastic strain does not increase with time after the initial loading, and it is recovered when stress is removed. At stresses greater than the yield strength, however, viscous flow results in permanent strain that continues to increase with time. A series of experiments can be run at progressively greater stresses to find where this transition occurs (Figure 9*c*).

Most laboratory measurements of yield strength show a marked increase as temperature falls below the liquidus and crystallinity increases. No unambiguous evidence has yet been reported that crystal-free liquids above their liquidus temperatures have measurable yield strengths. In the critical range of temperatures between the liquidus and solidus in which the liquid is saturated with one or more crystalline phases, the bulk crystal-liquid assemblages of most natural compositions have a yield strength that increases with both time and increasing crystallinity (Figure 8*b*).

Hulme (1974) has shown that the apparent yield strength estimated from the morphologies of arrested lava flows increases with silica content (Figure 10), but the relationship is ambiguous because the estimates do not provide a way of separating the effects of silica content from those of temperature and crystal content (Murase et al 1984). The relationship may be similar to

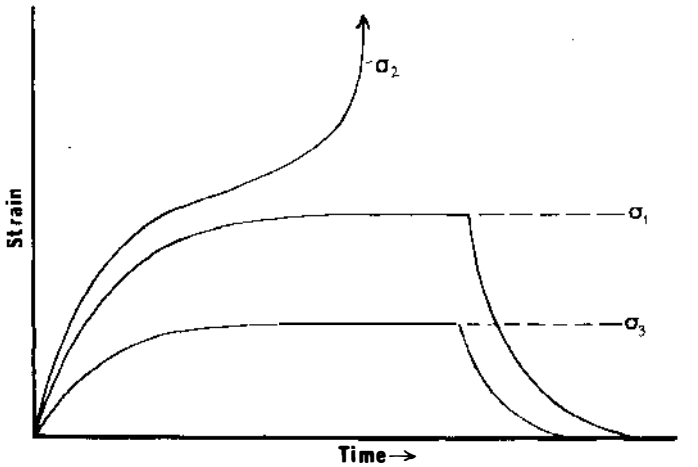

Figure 9 (continued)

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that found by Marsh (1981), who noted that beyond a certain limiting viscosity the rates of cooling and crystallization are rapid compared with the rate of flow, and the magma is simply demobilized by its slow response to shear stress.

Laboratory measurements show less correlation of yield strength with silica content than with the amount of cooling below the liquidus. If silica content is a factor, it cannot be the only one, because silica-rich liquids with cation ratios of alkalies to alumina greater than 1.0 have little or no yield strength, even at temperatures well below their liquidi (T. Murase and A. R. McBirney, unpublished data). Iron may also be an important factor, particularly at a high degree of oxidation. Exploratory experiments we have carried out on basalts show that samples having high yield strengths in air have much lower values under reducing conditions. We have no explanation for this difference.

No simple relation has yet been found between yield strength and the

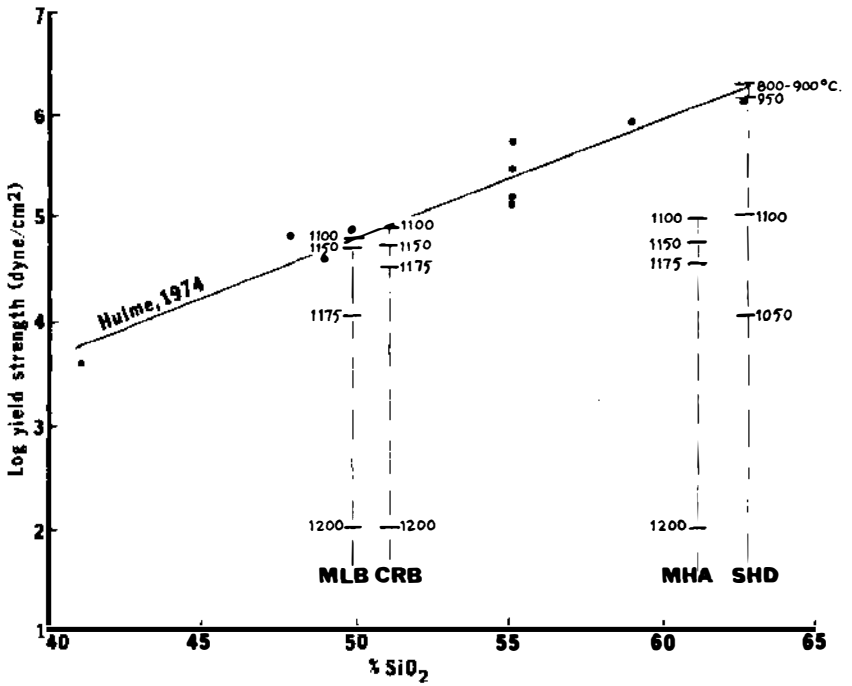

Figure 10 Estimates of yield strength of lavas of different silica contents. The sloping line has been defined by Hulme (1974) from the morphologies of arrested lava flows (solid dots). The last point at the high-SiO_2 end of the line is based on the shape of the dome of Mount St. Helens. Values for four rocks studied in the laboratory (MLB, CRB, MHA, SHD) are plotted as functions of temperature to show the increase of yield strength and the approach to Hulme's line as temperature falls and crystallization advances.

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volumetric proportions of crystals. A basic question still to be resolved is whether any part of the yield strength of a crystallizing silicate is inherent in the saturated liquid and not a bulk property of the crystal-liquid suspension. The magnitude of values measured in complex alumina-silicates containing only a few percent of small crystals suggests an extensive polymerization of the liquid phase under saturated conditions. None of the techniques developed thus far are able to make this distinction.

NEEDS FOR FURTHER RESEARCH

Much remains to be done in order to quantify the relations between rheological properties of natural silicates and such factors as composition, pressure, crystal content, shear rate, and a variety of time-dependent factors. Next to nothing is known about rates and mechanisms of deformation at very low long-term stresses, and little has been done on the effects of nonuniform temperature distributions and the simultaneous transfer of heat and momentum. The early studies by Shaw (1969) of viscous heating and mechanisms of thermal feedback are of unusual geological interest but have never been expanded, mainly because of the serious experimental difficulties they entail.

Rheological studies are certain to advance rapidly as experimental techniques are improved and more detailed measurements are made, particularly in the range of conditions between the liquidus and solidus temperatures of natural silicates. Few fields of geological research hold greater potential for enhancing our understanding of basic magmatic and tectonic processes.

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APPENDIX

Calculation of effective viscosity of a subliquidus magma Knowing the proportions and size of phenocrysts, it is possible to calculate the effective viscosity of a magma by means of relations outlined in the text, provided the composition of the groundmass liquid can be determined. The following program is designed to do this by first estimating the compositions of crystals that would form in a liquid having the bulk composition of the magma and then by subtracting these crystals to obtain the composition of the remaining liquid.

Volumetric proportions of crystals are converted to weight proportions

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by use of simple density relations. The density values used to do this need not be precise because the calculation of viscosity is much less sensitive to this factor than to others that are much more loosely constrained.

Densities and compositions of crystals in equilibrium with the liquid are estimated using the equilibrium coefficients compiled by Nathan & Van Kirk (1978) with subsequent modifications (H. D. Nathan, personal communication). Plagioclase is assumed to have perfect normal zoning, with a core in equilibrium with the initial liquid and subsequent growth reflecting the changing composition of the liquid as increments of 0.1 vol% are crystallized. The loop used to achieve this effect prolongs the run-time, and if less precision is needed, the growth step can be made larger by increasing the value of the step in line 1050. All other minerals are assumed to be unzoned, with compositions in equilibrium with the original (bulk) composition.

The viscosity of the remaining (groundmass) liquid is calculated using Shaw's (1972) equations and then corrected for the effect of suspended solids by using the equation of Sherman (1968), given as Equation (5) in the text. Constants for the Sherman equation were selected to give a result conforming to experimental measurements of the dacite of Mount St. Helens and a Hawaiian tholeiitic basalt.

The program is written in BASIC for an Apple computer and, in the interest of clarity, is given here in a more expanded form than necessary. A more condensed version can be obtained from either of the authors.

Table 1 Program to calculate effective viscosity of subliquidous magmas

```
100  REM : CALC VISC OF SUBLIQUIDUS MAGMAS
110  DEF  FN A(X) = ( INT (X * 100 + .5)) / 100
130  HOME : PRINT "ENTER OXIDES IN WT.%"
140  PRINT : INPUT "   SIO2    ";SI
150  INPUT "   TIO2    ";TI
160  INPUT "   AL203   ";AL
170  INPUT "   FE203   ";F3
180  INPUT "   FEO     ";F2
190  INPUT "   MNO     ";MN
200  INPUT "   MGO     ";MG
210  INPUT "   CAO     ";CA
220  INPUT "   NA2O    ";NA
230  INPUT "   K2O     ";K
240  INPUT "   P2O5    ";P
250  INPUT "ENTER H2O FOR VISCOSITY: ";H
255 DL = 2.2 + .027 * (F3 + F2 + TI): REM : APPROX DENSITY OF LIQUID
260 T = (SI + TI + AL + F3 + F2 + MN + MG + CA + NA + K + P) / (100 - H
300 SI = SI / T / 60.0843
310 TI = TI / T / 79.8988
320 AL = AL / T / 50.9806
330 F3 = F3 / T / 79.8461
340 F2 = F2 / T / 71.8464
350 MN = MN / T / 70.9374
```

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Table 1 (*continued*)

```
360 MG = MG / T / 40.3044
370 CA = CA / T / 56.0794
380 NA = NA / T / 30.9895
390 K = K / T / 47.0980
400 H = H / T / 18.0152
410 P = P / T / 70.9723
420 FM = F3 + F2 + MN + MG
430 TN = SI + TI + AL + FM + CA + NA + K + P + H
440  INPUT "ENTER MEAN DIAM OF XLS IN MM.: ";DM:DM = DM * 1000
500  PRINT "ENTER VOL. FRACTION OF:": PRINT
510  INPUT "   PLAGIOCLASE:    ";PF
520  INPUT "   K-FELDSPAR:     ";KF
530  INPUT "   OLIVINE:        ";OL
540  INPUT "   AUGITE:         ";AU
550  INPUT "   HYPERSTHENE:    ";HY
560  INPUT "   MT-ILM:         ";MT
570  INPUT "   QUARTZ:         ";QZ
580 PH = OL + PF + KF + AU + HY + MT + QZ
590  INPUT "ENTER TEMP, DEG C: ";TE:TE = TE + 273.15
670  IF OL = 0 THEN 760
680 MF = 3.33 * MG / F2: REM   MG/F2 OF CRYSTALS
690 DO = 4.39 - 1.17 * MF / (1 + MF):OL = OL * TN * DO / DL
700 SS = SS + OL * .333
710 SF = OL * .667
760  IF HY = 0 THEN 850
770 MF = 4.15 * MG / F2: REM   MG/F2 OF CRYSTALS
780 DH = 3.96 - .75 * MF / (1 + MF):HY = HY * TN * DH / DL
790 SS = SS + HY * .5
800 SF = SF + HY * .5
850  IF AU = 0 THEN 990
860 MF = 5.53 * MG / F2: REM   MG/F2 OF CRYSTALS
870 DA = 3.56 - .26 * MF / (1 + MF)
880 AU = AU * TN * DA / DL
900 SS = SS + AU * .5
920 SF = SF + AU * .25
940 SC = SC + AU * .25
990  IF QZ = 0 THEN 1040
1000 QZ = QZ * 2.65 * TN / DL
1020 SS = SS + QZ
1040  IF PF = 0 THEN 1140
1050  FOR Z = 0 TO PF STEP .001
1060 CN = 3.68 * (CA * AL) / (NA * SI): REM    CA/NA OF PLAG
1070 DP = 2.62 + .14 * CN / (1 + CN):PL = .001 * TN * DP / DL
1080 SS = SS + PL * .6 - PL * .2 * CN / (1 + CN)
1090 SA = SA + PL * .2 + PL * .2 * CN / (1 + CN)
1100 SC = SC + PL * .2 * CN / (1 + CN)
1110 SN = SN + PL * .2 / (1 + CN)
1120  NEXT Z
1140  IF KF = 0 THEN 1200
1150 KF = KF * TN * 2.62 / DL
1160 SS = SS + KF * .6 * TN
1170 SA = SA + KF * .2 * TN
1180 SK = KF * .2
1200  IF MT = 0 THEN 1300
1230 FT = .92 * F3 * F3 / (F2 * TI): REM   MAGNETITE/ULVOSPINEL
1240 MT = MT * TN * 5.18 / DL
1250 ST = MT * .1200 / (1 + FT)
```

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Table 1 (*continued*)

```
1260 SF = SF + MT * .88
1300 SI = SI - SS
1310 TI = TI - ST
1320 AL = AL - SA
1330 FM = FM - SF
1360 CA = CA - SC
1370 NK = NA - SN + K - SK
1380 H = H / (1 - PH)
1400 TV = SI + TI + AL + FM + CA + NK + P + H:TR = TN / TV
1410 SI = SI * TR
1420 TI = TI * TR
1430 AL = AL * TR
1440 FM = FM * TR
1480 CA = CA * TR
1490 NK = NK * TR
1510 P = P * TR
1520 H = H * TR
1530 TC = SI + TI + AL + FM + CA + (NK + P) / 2 + H
2000  REM     CALCULATE VISC OF LIQ FRACTION
2010 Q1 = AL * 6.7
2020 Q2 = FM * 3.4
2030 Q3 = (CA + TI) * 4.5
2040 Q4 = NK * 1.4
2050 Q5 = H * 2
2060 QT = (Q1 + Q2 + Q3 + Q4 + Q5) * SI / TC ^ 2
2070 MU = (QT / (1 - SI / TC)) * (1E4 / TE - 1.5) - 6.4
2100  REM : CALC VISC OF LIQ + XLS
2110  IF PH = 0 THEN NE = MU: GOTO 2170
2120  IF DM = 0 THEN 2070
2130 PH = (1 / PH) ^ .333
2140 SH = .011 * DM / (PH - 1) - .15
2150 NE = MU + SH
2170  PRINT "LOG VISC OF GLASS = " FN A(MU / 2.3026)
2180  PRINT "LOG VISCOSITY = " FN A(NE / 2.3026)
3000  END
```

Literature Cited

Ashby, M. F. 1972. A first report on deformation-mechanism maps. *Acta Metall.* 20:887–97

Booth, B., Self, S. 1973. Rheological features of the 1971 Mount Etna lavas. *Philos. Trans. R. Soc. London Ser. A* 274:99–106

Bottinga, Y., Weill, D. F. 1972. The viscosity of magmatic silicate liquids: a model for calculation. *Am. J. Sci.* 272:438–75

Burnham, C. W. 1963. Viscosity of a water-rich pegmatite melt at high pressures. *Geol. Soc. Am. Spec. Pap. No. 76*, p. 26

Carron, J. 1969. Recherches sur la viscosite et les phénomènes de transport des ions alcalins dans les obsidiennes grantiques. *Trav. Lab. Geol., Éc. Norm. Super., Paris.* 112 pp.

Cigolini, C., Borgia, A., Cassertano, L. 1983. Intracrateric activity, aa–block lava, viscosity and flow dynamics: Arenal Volcano, Costa Rica. *J. Volcanol. Geotherm. Res.* In press

Einarsson, T. 1949. The eruption of Hekla, 1949–1948, Pt. IV, 3. *Soc. Sci. Isl.* 74 pp.

Euler, R., Winkler, H. G. F. 1957. Über die Viskositaten von Gesteins und Silikatschmelzen. *Glastech. Ber.* 30:325–32

Gauthier, F. 1973. Field and laboratory studies of the rheology of Mount Etna lava. *Philos. Trans. R. Soc. London Ser. A* 274:83–98

Hess, P. C. 1980. Polymerization model for silicate melts. In *Physics of Magmatic Processes*, ed. R. B. Hargraves, pp. 3–48. Princeton, NJ: Princeton Univ. Press

Hulme, G. 1974. The interpretation of lava flow morphology. *Geophys. J. R. Astron. Soc.* 39:361–83

Annu. Rev. Earth Planet. Sci. 1984.12:337-357. Downloaded from www.annualreviews.org by Brown University on 02/10/14. For personal use only.

Johnson, A. M. 1970. *Physical Processes in Geology*. San Francisco: Freeman & Cooper. 577 pp.
Khitarov, N. I., Lebedev, E. B. 1978. The peculiarities of magma rise in presence of water. *Bull. Volcanol.* 41:354–59
Khitarov, N. I., Lebedev, E. B., Dorfman, A. M., Slutsky, A. B. 1978. Viscosity of dry and water-bearing basalt melts under the pressure. *Geokhimiya* 6:900–5
Kushiro, I. 1978. Density and viscosity of hydrous calc-alkalic andesite magma at high pressures. *Carnegie Inst. Washington Yearb.* 77:675–77
Kushiro, I. 1980. Viscosity, density, and structure of silicate melts at high pressures, and their petrological applications. In *Physics of Magmatic Processes*, ed. R. B. Hargraves, pp. 93–120. Princeton, NJ: Princeton Univ. Press
Kushiro, I., Yoder, H. S. Jr., Mysen, B. O. 1976. Viscosities of basalt and andesite melts at high pressures. *J. Geophys. Res.* 81:6351–56
MacKenzie, J. D. 1956. Simultaneous measurements of density, viscosity, and electrical conductivity of melts. *Rev. Sci. Instrum.* 37:297–99
Marsh, B. D. 1981. On the crystallinity, probability of occurrence, and rheology of lava and magma. *Contrib. Mineral. Petrol.* 78:85–98
Moore, H. J., Arthur, D. W. G., Schaber, G. G. 1978. Yield strengths of flows on the Earth, Mars, and Moon. *Proc. Lunar Planet. Sci. Conf., 9th*, pp. 3351–78
Murase, T. 1962. Viscosity and related properties of volcanic rocks at 800° to 1400°C. *J. Fac. Sci. Hokkaido Univ. Ser. 7* 1:487–584
Murase, T., McBirney, A. R. 1973. Properties of some common igneous rocks and their melts at high temperatures. *Geol. Soc. Am. Bull.* 84:3563–92
Murase, T., McBirney, A. R., Melson, W. G. 1984. The viscosity of the dome of Mt. St. Helens. *J. Volcanol. Geotherm. Res.* In press
Nathan, H. D., Van Kirk, C. K. 1978. A model of magmatic crystallization. *J. Petrol.* 19:66–94
Peterson, D. W., Tilling, R. I. 1980. Transition of basaltic lavas from pahoehoe to aa, Kilauea Volcano, Hawaii. *J. Volcanol. Geotherm. Res.* 7:271–94
Pinkerton, H., Sparks, R. S. J. 1978. Field measurements of the rheology of lava. *Nature* 276:383–85
Robson, G. R. 1967. Thickness of Etnean lavas. *Nature* 216:251–52
Roeder, P. L., Dixon, J. M. 1977. A centrifuge furnace for separating phases at high temperature in experimental petrology. *Can. J. Earth Sci.* 14:1077–84
Ryan, M. P., Sammis, C. G. 1981. The glass transition in basalt. *J. Geophys. Res.* 86:9519–35
Scarfe, C. M. 1973. Viscosity of basic magmas at varying pressure. *Nature* 241:101–2
Shartsis, L., Spinner, S. 1959. Viscosity and density of molten glasses. *J. Res. US Nat. Bur. Stand.* 46:176–94
Shaw, H. R. 1963. Obsidian-H_2O viscosities at 1000 and 2000 bars in the temperature range 700° to 900°C. *J. Geophys. Res.* 68:6337–42
Shaw, H. R. 1969. Rheology of basalt in the melting range. *J. Petrol.* 10:510–35
Shaw, H. R. 1972. Viscosities of magmatic silicate liquids: an empirical method of prediction. *Am. J. Sci.* 272:870–93
Shaw, H. R., Wright, T. L., Peck, D. L., Okamura, R. 1968. The viscosity of basaltic magma: an analysis of field measurements in Makaopuhi lava lake, Hawaii. *Am. J. Sci.* 266:255–64
Sherman, P. 1968. *Emulsion Science*. New York: Academic. 351 pp.
Sparks, R. S. J., Pinkerton, H., Hulme G. 1976. Classification and formation of lava levees on Mount Etna, Sicily. *Geology* 4:269–71
Stocker, R. L., Ashby, M. F. 1973.

Annu. Rev. Earth Planet. Sci. 1984.12:337-357. Downloaded from www.annualreviews.org by Brown University on 02/10/14. For personal use only.