Transport properties of glassy and molten lavas as a function of temperature and composition

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Abstract

We provide measurements of thermal diffusivity (D), heat capacity (C_P), and viscosity (η) for 12 remelted natural lavas and 4 synthetic glasses and melts, ranging in composition from leucogranite to low-silica basalt, and calculate their thermal conductivity (k). Both viscosity and the glass transition temperature decrease with decreasing melt polymerization. For basaltic glasses, D is low, $\sim 0.5 \text{ mm}^2 \text{ s}^{-1}$ at room temperature, decreases slightly with increasing temperature, and then drops upon melting to ~ 0.25 to $0.35 \text{ mm}^2 \text{ s}^{-1}$. Other samples behave similarly. Despite scatter, clear correlations exist between D of glass or melt with Si content, density, NBO/T, and, most strongly, with fragility (m). Glass thermal diffusivity is represented by $D = FT^{-G} + HT$, where F, G and H are fitting parameters. For melts, $\partial D / \partial T$ was resolved only for dacite-andesite and MORB: a positive slope is consistent with other iron-bearing samples. Glass and liquid C_P depend on density and other physical properties, but not exactly in the same manner as D . We calculate thermal conductivity (k) from these data and demonstrate that k for glasses is described by a Maier-Kelly formula. Large scatter exists for k at 298 K, but silicic to intermediate melts have k between 1.8 and $1.3 \text{ W m}^{-1} \text{ K}^{-1}$, whereas basaltic melts are constrained to $\sim 1.4 \pm 0.1 \text{ W m}^{-1} \text{ K}^{-1}$. Low values for thermal diffusivity and viscosity for basaltic melts suggests that basalts transfer heat much more efficiently by advection than by conduction alone, and that partially molten zones in the mantle quickly become more thermally insulating than non-molten zones, potentially contributing to melt localization during decompression melting.

Keywords: high-temperature, thermal diffusivity, viscosity, heat capacity, thermal conductivity, lavas, glass, melt

INTRODUCTION

Heat transport plays a crucial role in the thermal evolution of high-temperature, magmatic regimes (e.g., Nabelek et al. 2012). Conduction of heat is quantified by Fourier's law based on thermal conductivity (k). However, large uncertainties exist in k because conventional experimental techniques involving physical contacts are inaccurate for materials that bond poorly with metal contacts or are partially transparent, such as glasses and melts, especially at high temperatures (T) relevant to magmatic processes (e.g., Hofmeister et al. 2007). In contrast, thermal diffusivity (D) can be measured with a high degree of accuracy ($\sim\pm 2\%$) using laser-flash analysis (LFA; Parker et al. 1961). This method is contact-free and lacks systematic errors associated with conventional methods, such as thermal losses at interfaces of $\sim 10\%$ per contact (Hofmeister 2007). Most importantly, removal of boundary-to-boundary radiative transfer gains (Degiovanni et al. 1994; Mehling et al. 1998), which can overwhelm intrinsic behavior at temperatures as low as 1000 K, makes the LFA technique optimal for measuring glasses and melts which are partially transparent in the near-IR. Measurements up to several hundred K above the glass transition (i.e., prior to flow or crystallization) are also possible using LFA, thereby providing information on the liquid state. Combining measurements of heat capacity at constant pressure (C_P), density (ρ), and D allows determination of k associated with vibrational modes from its definition:

$$k_{\text{lat}} = \rho C_P D . \quad (1)$$

For melts, heat can also be transported by advection. Mass transport depends on melt viscosity, which can be measured by various means (e.g. Dingwell 1995), and provides information on the liquid state to high temperatures.

Our previous efforts using LFA focused on the dependence of thermal and mass transport properties on melt chemistry by studying magma analogs. We began with simple glass compositions corresponding to crustal minerals: quartz (SiO_2), alkali feldspar (XAlSi_3O_8 where $\text{X} = \text{Li, K, Na}$), anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$), and clinopyroxene (XYSi_2O_6 where $\text{XY} = \text{CaMg}$ or LiAl) (Pertermann et al. 2008; Hofmeister et al. 2009; Hofmeister and Whittington 2012), and found that mass and heat transport properties are linked: felsic liquids have both high viscosity and high thermal diffusivity compared to mafic liquids. This correlation also holds for more complex melt compositions with high silica contents (71-80 wt. % SiO_2), despite variable quantities of Al_2O_3 , FeO , MgO , CaO , Na_2O , and K_2O (Hofmeister et al. 2014a). The latter measurements further indicate that Ca and Fe strongly affect transport properties. The effect of Fe on D , in the absence of Al, was probed by examining the enstatite-ferrosilite binary, which showed strong and positive $\partial D/\partial T$ at high temperature (Hofmeister et al. 2014b).

The present paper reports coordinated LFA, heat capacity, and viscosity measurements on synthetic and remelted natural lavas. Effects of oxidation state and minor crystallites are also probed. Our focus is on basaltic compositions, which are important due to their ubiquitous occurrence on Earth and other planetary bodies. We find that D depends roughly on density, NBO/T, and fragility, which in turn largely depend on Si content, although a predictive model for k remains elusive. Viscosity and D for basaltic melts are particularly low, implying that heat transfer by advection is relatively more efficient than conduction. Our new measurements pave the way for an improved understanding of magmatic processes.

EXPERIMENTAL METHODS

Synthesis procedures

Synthetic rhyodacite glass was prepared by weighing and mixing reagent grade oxides in the desired proportions with acetone in an iron-free platinum crucible. The mixture was decarbonated slowly during heating over many hours from 900 to 1100 °C prior to fusion in the same platinum crucible in a muffle furnace in air at 1600 °C for 60 minutes. The crucible was removed from the furnace and allowed to cool to room temperature, producing a clear crystal-free glass. If not, samples were reground and remelted until a homogeneous glass was obtained. Chemical homogeneity and absence of crystallites were ascertained by visual inspection with an optical microscope, and by electron microprobe analysis and scanning electron microscopy, see below. Synthetic andesite, basaltic andesite, and basalt were made in a similar way.

Likewise, twelve glasses were made by remelting the lavas described in Table 1. Crushed rock was placed in iron-saturated platinum or platinum-rhodium crucibles and remelted in a muffle furnace in air at 1600°C for between 1 and 2 hours, stirred, then quenched to form glasses either by dipping the bottom of the crucible in water, or by pouring on to a copper plate.

The as received sample of MORB had glassy rims sufficient in size to make measurements of thermal diffusivity. The small size of the rims, and rapid crystallization at high T , prohibited remelting and viscosity measurements.

Sample preparation

For parallel plate viscometry, cylindrical samples were cored from glass lumps using a diamond core drill, avoiding visible bubbles. The cylinders were cut to lengths of 5 to 10 mm using a diamond wafer saw, polished on successively finer grit papers, and parallel faces were

verified using a micrometer. For LFA, sections were sawed and ground into disks of ~12 mm diameter with 0.2 to 1.1 mm thicknesses and nearly parallel surfaces, and sand-blasted with 50–150 μm alumina grit. For calorimetry, samples were cut into flat disks (4–5 mm in diameter) with uniform thickness, of ~600–1100 μm , and weights between 40 to 60 mg. For spectroscopic measurements, double-polished sections were prepared. Polished chips were used in electron microprobe analysis.

Chemical analyses

Samples were characterized by wavelength dispersive analysis (WDS) using standard procedures on the JXA-8200 electron microprobe at Washington University. We used “Probe for Windows” for data reduction (see <http://www.probesoftware.com/>). The measured data were corrected using CITZAF after Armstrong (1995). Oxide and silicate standards were used for calibration (e.g., Amelia albite for Na, Si; microcline for K; Gates wollastonite for Ca; Alaska Anorthite for Al; synthetic fayalite for Fe; synthetic forsterite for Mg; synthetic TiO_2 for Ti; synthetic Mn-olivine for Mn; synthetic Cr_2O_3 for Cr). Large beam size and measurement of Na at first during each analysis was used to avoid volatilization, which was confirmed by high totals obtained (see Table 2). Millimeter-scale homogeneity was verified by analyzing different areas of glass chips.

Wet chemistry was performed to determine the $\text{Fe}^{3+}/\text{Fe}^{2+}$ in the remelted dolerite and continental basalt. Procedures are described by Sehlke et al. (2014), who measured the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio for OIB2.

Spectroscopic characterization of impurity ions

Unpolarized ultraviolet (UV) to visible spectra from 190-1100 nm (9090-52630 cm^{-1}) were collected using a double-beam Shimadzu UV-1800 with 1 nm resolution. For lower frequencies ($\sim 2000\text{-}14000\text{ cm}^{-1}$), a Bomen Fourier transform infrared (FTIR) spectrometer was used with an InSb detector and each of a CaF_2 and quartz beamsplitter. Absorption coefficients (A) were calculated using thickness (L) measured with a digital micrometer and from $AL = -\log(I_{\text{trans}}/I_0)$.

Evaluation of crystallinity

All transparent samples, whether colorless, green, or brown, were visibly free of crystals, which was confirmed using high magnification electron microscopy during chemical analysis. It is possible that the dacite has very tiny crystallites in small amounts, given its color, as in natural rhyolites. Pieces of the various basalts used for thermal diffusivity analysis are opaque at the thickness used, so it is possible that the interiors of these outwardly glassy black samples had some crystals, even when electron microscopy did not reveal any crystallites. To recognize when data are affected by crystallites, we include sample OIB1 in our study for which a small amount of crystallites was consistently detected by electron microscopy, although they were not visible on its surface. Sample P-MORB may have crystallites, because these were detected in one of the three chips examined. Unlike the other black basalts and transparent glasses, D of OIB1 depends nearly linearly on T . This behavior was also seen in reheated samples which have some crystallites (data not reported). Moreover, the viscosity measurements on OIB1 are discordant with the other basalts and shifted to higher viscosity for a given temperature, suggesting the presence of several volume percent crystallites. From these comparisons, the data we report on all other black samples represent glass with very low amounts of crystallites, if any.

Density determinations

Glass density was obtained using the Archimedean method, with ethanol as the immersion liquid. Repeat measurements indicate precision is $\pm 2 \text{ kg m}^{-3}$ except for MORB, which showed variation of $\pm 17 \text{ kg m}^{-3}$ among three pieces. This (not remelted) sample contained minor quantities of microscopic bubbles, which may explain the larger variability.

Viscosity measurements

Cylindrical glass cores of ~6 mm diameter were drilled and then sliced to ~6 to 20 mm length. The ends were polished until parallel. Viscosity above the glass transition but below liquidus temperatures was measured using a Theta Instruments Rheotronic III parallel plate viscometer, with a constant uniaxial load of 1500 g, and a maximum temperature of 1000°C. Viscosity is calculated from the measured longitudinal strain rate, known load and calculated instantaneous surface area, assuming perfect slip between sample and plates. Relatively low finite strains were used ($\leq 20\%$) and samples remained cylindrical after measurement, justifying this assumption. Measurements made in this study ranged in temperature from 630 to 889°C, and in viscosity from 8.7×10^8 to $1.3 \times 10^{13} \text{ Pa.s}$. The accuracy and precision of the measurements are ± 0.06 log units, confirmed by repeat measurements on multiple glass cores where sufficient sample was available. Whittington et al. (2009b) give details regarding experimental protocol and instrument calibration. In one case, crystallization caused two measurements to deviate to higher viscosity than predicted from the trend of previous measurements on the same sample (MORB, Table 3). These data were not used in fitting the VFT equations and should not be used in viscosity modeling because they no longer represent a pure liquid phase.

Viscosity was also measured at superliquidus temperatures using a Theta Instruments Rheotronic II 1600C Rotating Viscometer, equipped with a Brookfield HBDV-III Ultra measuring head, which can measure over the range 1 to 10^5 Pa s with accuracy and precision of ± 0.03 log units, based on repeat measurements of NIST standard glasses. Calibration and experimental protocols are described fully by Getson and Whittington (2007). Data collection began at 1600°C and measurements were typically made at 20 to 30°C intervals. At some point the liquidus was crossed, and crystallization would cause measurements to deviate to higher viscosity than predicted from the trend of previous measurements (e.g. P-MORB, Table 3). These data were not used in fitting the VFT equations and should not be used in viscosity modeling because they no longer represent a pure liquid phase. In each case we reheated the sample above its liquidus, and reproduced the earlier measurements within experimental uncertainty. The low rmsd of each fit reported in Table 4 shows that the results of both viscometric techniques are in excellent agreement.

Thermal diffusivity measurements

Our LFA 427 apparatus is manufactured by Netzsch Gerätebau, Germany. Glasses were sawed and ground into disks of ~8-12 mm diameter with 0.2-1.1 mm thicknesses and nearly parallel surfaces, and sand-blasted with 50–150 μ m alumina grit. Specimens are held in a furnace in an Ar gas atmosphere using graphite holders. The temperature dependence of D is obtained by varying furnace temperature, which is measured to within $\sim 1^\circ\text{C}$ using calibrated W-Re thermocouples. A pulse with ~ 0.5 ms width from a 400 W Nd-GGG laser heats the sample from below, providing a difference of $\sim <4^\circ\text{C}$ across the sample. As heat diffuses from the bottom to the top of the sample, emissions upward are recorded as a function of time with an InSb detector.

Graphite coatings on the sample serve to block laser light, enhance absorption of the laser pulse and sample emissions, and buffer oxygen fugacity at high temperatures to C-CO. Data were obtained at 50-100 °C intervals with ~three acquisitions at each temperature which take a few seconds, and 1 minute between acquisitions. Heating rates varied from 8 to 18 K per minute, and stability was reached before each acquisition, as confirmed by repeatability among the 3 data points. Runs typically take a total of 6 to 8 hours. These rates are not important due to the rapidity of the data collection (see discussion of T_{12} vs $T_{g,LFA}$ below).

Data were processed using the algorithm of Mehling et al. (1998) to extract thermal diffusivity from the time-dependent emissions. This model accounts for radiative surface losses to the surroundings and spurious radiative transfer through the sample between the top and bottom graphite coats, and allows for absorbance being frequency dependent, although the detailed values of optical properties are not needed. The measured shape of the laser pulse is accounted for (Blumm and Opfermann 2002).

Thermal diffusivity is accurate to 2%, verified against opaque reference materials (see Pertermann et al. 2008). We emphasize that this uncertainty arises from uncertainties in thickness and in slight departures from parallel faces. The precision within any given run is much higher, so the temperature dependence of D is better constrained than the absolute values of D .

Heat Capacity

Isobaric heat capacities (C_p) were measured between room temperature and 1770 K using a Netzsch DSC 404 Pegasus F1 Differential Scanning Calorimeter (DSC) with a Pt-furnace. A 20 ml min⁻¹ nitrogen gas flow was used to stabilize atmospheric conditions within the measurement cell. Pt-Rh alloy was used as sample and reference pan materials.

Each heat capacity measurement consisted of three run segments: (1) a blank run with empty sample and reference pans, (2) a standard run with empty reference pan but using a pure 1 mm thick sapphire (α -Al₂O₃) disk with a weight of ~115 mg, and (3) a sample run with an empty reference pan. Each run segment consisted of a 15 minute isothermal equilibration segment at 325 K, data collection while the cell was continuously heated at a constant rate of 20 K min⁻¹ up to some maximum temperature, a 5 minute isothermal equilibration segment, after which the furnace was turned off and the measurement cell cooled ballistically.

The α -Al₂O₃ disk yielded values within 1% of reference values (Ditmars et al., 1982) for measurements up to ~1200 K, but differed by ~ 2% toward higher temperatures up to 1770 K. These differences indicate experimental uncertainties.

RESULTS

Chemical compositions, NBO/T, and density

Chemical analyses are given in Table 2. The remelted glasses have compositions consistent with the stated rock types. Our remelted Chengwatana, WI rift basalt is enriched in silica compared to analyses of other batches (Wirth et al. 1997). Water contents are uniformly low, <200 ppm, from the spectroscopic measurements of transparent samples in Fig. 1. Because this behavior pertains to all samples prepared by us, and such low water contents do not affect *D* values, we do not pursue this aspect further.

Values of NBO/T, calculated from the chemical compositions span a wide range, from 0 to 1 (Table 2). Andesite has a low NBO/T for its silica content, due to its relatively high Al content, whereas OIB2 has a high NBO/T for its silica content, due to its relatively low Al content.

Measured density of glass increases across our series, from rhyolite to basalt (Table 1). Calculated ρ for the liquids (Table 2, after Lange, 1997) show a similar trend. Liquid densities are lower than glass densities across the compositional spectrum, especially for the basalts, whose liquidus temperatures are much higher than those of silicic to intermediate melts.

Oxidation state

Visible spectra (Fig. 1a) show that all samples strongly absorb above 300 nm, attributable to metal-oxygen charge transfer. All Fe-bearing samples have bands near 1000 nm due to Fe^{2+} transitions (Fig. 1b). Brown Fe-bearing samples (including leucogranite and rhyolite: Hofmeister et al. 2014a) absorb strongly near 400 nm, attributable to Fe^{3+} transitions and to Fe^{2+} - Fe^{3+} charge transfer. Green dacite-andesite appears to lack Fe^{3+} , despite its substantial total Fe content of 3.26 wt % (as $\text{FeO}_{\text{total}}$). Comparing peak heights suggests FeO near 0.03 wt % for synthetic andesite-basalt and 0.01 wt% for basalt. Low amounts are resolved via use of thick sections. The synthetic andesites have negligible amounts of iron cations.

We did not attempt to quantify Fe concentrations from spectra with mixed valances because d-d peak heights can be affected by the presence of charge transfer and can depend on the concentration (see Rossmann, 1988). However, qualitatively, $\text{Fe}^{3+}/\text{Fe}^{2+}$ is similar for dacite, basalt-andesite, and also for the arc basalt, which was measured by wet chemistry as ~ 1 (Robert et al. 2014). The dolerite has lower ferric content, given its strong Fe^{2+} peak, which is consistent with $\text{Fe}^{3+}/\text{Fe}^{2+} = 0.45$ from our wet chemical analyses. Spectra were not obtained for the iron rich samples. Wet chemistry gave $\text{Fe}^{3+}/\text{Fe}^{2+} = 2.1$ for OIB2 (Sehlke et al. 2014) and 1.7 for continental, so it is likely that MORB and P-MORB with high total Fe also have $\text{Fe}^{3+}/\text{Fe}^{2+} \sim 1.7$ or larger.

Viscosity

Near the glass transition, viscosity is close to a linear function of inverse temperature for all samples. The most silica-rich samples have the highest viscosity and show shallower slopes than basaltic melts (Fig. 2; Table 3). Both viscosity and the viscometric glass transition temperature, taken to be the temperature of the 10^{12} Pa s isokom (T_{12}), decrease with decreasing melt polymerization (Table 4). At high temperatures, above their liquidii, basaltic melts remain more fluid than silicic melts but they now have the shallower slope on an Arrhenian diagram.

The datasets for individual compositions were fitted with the non-Arrhenian Vogel-Fulcher-Tamman (VFT) equation:

$$\log \eta = A + B/(T-C) \quad (2)$$

where A, B and C are adjustable parameters. Best-fit values are given in Table 4, excluding OIB1 due to the presence of crystallites. It is generally accepted that all silicate melts tend towards a common value of A of about -4.5 at very high temperatures (Russell et al. 2003), and most of our A values were in the range -4.5 ± 0.5 . Leucogranite and rhyolite are more negative (-6 to -7) whereas MORB and P-MORB are more positive (~ -2.6). The viscosity dataset and fitted VFT equations allow interpolation of viscosity with a high degree of confidence from the glass transition up to the highest temperatures at which samples were analyzed by LFA, which is useful in interpreting changes in D observed at high temperatures.

One way to quantify the degree of non-Arrhenian behavior, or “fragility”, is to calculate the gradient of the viscosity curve at the glass transition temperature on a reduced temperature scale (Plazek and Ngai 1991); see Table 4. Melt fragility can be calculated from the VTF coefficients:

$$m = B/[T_{12}(1-C/T_{12})^2] \quad (3)$$

Calculated values of fragility range from 22 for leucogranite to 48 for P-MORB, with all the basaltic samples ($NBO/T \geq 0.5$) having a fragility near 45 ± 3 .

Fragility is directly related to configurational heat capacity, C_P^{conf} , which is the difference between C_P of the glass at T_g and C_P of the melt (Richet et al. 1986). From the temperature dependence of relaxation times in viscous liquids (Adam and Gibbs 1965), the viscosity of a liquid can be related to its configurational entropy (S^{conf}) through:

$$\log \eta = A_e + \frac{B_e}{TS^{conf}} \quad (4)$$

where A_e and B_e are constants differing from A , B and C above. These reflect the Gibbs free-energy barriers hindering co-operative rearrangements in the liquid (Richet 1984). Melts with small values of C_P^{conf} have small changes in S^{conf} with temperature, and hence exhibit near-Arrhenian or “strong” behavior. Melts with large values of C_P^{conf} have correspondingly large changes in S^{conf} with temperature, and hence exhibit distinctly fragile, non-Arrhenian, behavior.

Thermal Diffusivity

Thermal diffusivity of the glasses is highest at room temperature and decreases as temperature increases from 298 to about 1000 K (Fig. 3; Table 5). By about 800 K, D becomes nearly constant, as observed for other silicate glasses (e.g., Hofmeister et al. 2009). In some cases (leucogranite, dacite, arc basalt, OIB2) D reaches a minimum near 600 to 800 K and then clearly increases at higher T . The positive, high-temperature slopes are discussed below after presenting fits to the $D(T)$ data.

As discussed earlier, OIB1 with minor amounts of crystallites has a straight trend that is not seen in the homogeneous glasses. P-MORB had some tiny crystallites in some chips. The trend in D is slightly less bowed, which may be due to small amounts of crystallites, but

similarities to other trends, suggest that this variation could equally well be due its particular chemical composition and structure. We do not see evidence for crystallites in the response of the dacite glass to temperature. Tiny amounts of crystallites having a small effect is consistent with behavior of the natural rhyolites (Romine et al. 2012).

As in our previous studies, D decreases substantially upon crossing the glass transition (Fig. 3). Behavior above the glass transition is complicated (discussed below).

Our measurements probe relaxed melt, starting at T corresponding to a viscosity of $\sim 2 \times 10^8$ Pa s. As detailed in Pertermann et al. (2008) and Hofmeister et al. (2009), the temperature of the transition in LFA ($T_{g,LFA}$) is higher than that in viscometric experiments (T_{12}) due to the time-scales of the experiments. At low T , a viscous melt appears frozen (glassy) on the short timescale of the laser pulse, although it can still flow over the longer (minutes to hours) timescale of a viscosity measurement. Although the state immediately above the glass transition is supercooled (metastable) liquid, thermal diffusivity above $T_{g,LFA}$ represents that of melt up to and including superliquidus conditions, and is thus denoted D_{melt} . For brevity, we refer to this state as liquid or melt.

Roughly, siliceous lavas have $D_{melt} = 0.52 \text{ mm}^2\text{s}^{-1}$, intermediate lavas have D_{melt} near $0.40 \text{ mm}^2\text{s}^{-1}$, and basalts have D_{melt} of >0.20 to $0.37 \text{ mm}^2\text{s}^{-1}$ (Table 6). In detail, for the synthetics with very little Fe, D_{melt} was observed over a few hundred K, and was found to be constant (Fig. 3). Recovered synthetic samples retained their original thickness and appearance. In contrast, for most Fe-bearing samples, observing the temperature dependence of D_{melt} was precluded by these low viscosity samples flowing, outgassing, and sometimes partially crystallizing after melting (Fig. 3), so Table 6 only reports D at the completion of melting, but prior to the above behaviors. Recovered samples of OIB1 and P-MORB had flowed and warped. In these two cases, because

the true thickness during the highest T acquisitions is larger than the initial thickness used in calculations by the instrument software, and because D goes as $1/L^2$ (Parker et al. 1961), the reported low values of D (0.17 to 0.25 mm²s⁻¹) likely underestimate D_{melt} , as indicated in Table 6. Due to similar limitations, a few samples provide upper limits on D_{melt} (Table 6). For the 2nd run on arc basalt, the sample melted and fell in the heating interval between temperature set points so data could not be collected on D_{melt} .

The Si-rich natural samples (Fig. 3a) did not crystallize, but with time flowed and degassed slightly, as indicated by the slightly increased thickness and bumpy texture of the recovered samples. The increase in thickness due to flow precludes accurately determining $\partial D_{\text{melt}}/\partial T$. For these Si-rich compositions (Fig. 3a), D for the melt can only be obtained reliably from the lowest T data on the melt.

The data for the dacite-andesite melt indicate positive $\partial D/\partial T$ (Fig. 3b). This positive slope is not due to crystallization or degassing because the recovered sample remained transparent, with few bubbles. Although the thickness on one edge had increased slightly, the center was unchanged, so positive $\partial D/\partial T$ for dacite-andesite is intrinsic. If the edge thickening were the source of the T dependence of D , then the slope would have been negative, not positive as observed.

For MORB, the two sections examined have different melting temperatures (Table 6) due to variations in initial water content. Because water contents are not near the >0.1 wt.% level needed to alter D -values (Hofmeister et al. 2006), the data instead constrain the T dependence of D_{melt} for MORB (Fig. 3d). The slope for MORB is similar to that found for dacite-andesite melt (Table 6, footnotes). The strong dependence of melting temperature on water content will be quantified elsewhere.

We considered several different formulae in fitting D of glass to a function of T . The T^{-1} form that is frequently used to describe phonon scattering of crystalline material (e.g., Julian 1965) is obviously too steep for the flat lying trends of Fig. 3. Nor did a simple power law accurately reproduce most trends. This finding, coupled with observed upturns at high T for SiO_2 glass (Hofmeister and Whittington 2012), Fe-bearing high-silica glasses (Hofmeister et al. 2014a), and Fe-rich pyroxene glasses (Hofmeister et al. 2014b), suggests that the following form best describes the temperature dependence of D :

$$D = FT^{-G} + HT \quad (5)$$

Note that parameters F and G are not independent due to trade-offs in the fitting. Also, because precision during the run is much better than the 2% uncertainty in the absolute values of D arising from L , the T dependence is well-constrained.

Equation 4 is not limited to glasses but was originally developed to describe data on crystals (Hofmeister et al. 2014c), where G depends on the crystal structure and cation disorder. The crystalline material (of those studied) most similar to glasses are feldspars, which are highly disordered, and have low G (0.30 to 0.85). For our remelted lavas, fitted values of G are even lower, ranging from 0.07 to 0.52 (Table 6), which is consistent with glasses being more disordered than feldspars.

Regarding the physical basis of Eq. (4), the FT^{-G} term describes phonon-phonon scattering whereas the HT term is attributed to radiative transfer at infrared frequency (Hofmeister et al. 2014c). More exactly, the proposed radiative mechanism involves polaritons, which, in essence, are a blend of infrared and acoustic modes.

Heat Capacity

Table 7 lists heat capacity data. For all glasses, C_P first gradually increases as T increases, following the Maier-Kelly (1932) formulation (Fig. 5; Table 8). The glass transition is represented by a lambda-shaped curve and an abrupt increase in C_P that represents the configurational heat capacity. The very narrow range of C_P at the glass transition reflects the Dulong-Petit harmonic limit of $3R \text{ J mol}^{-1} \text{ K}^{-1}$, where R is the universal gas constant (Neuville and Richet 1991, Richet 1984).

Regarding the liquid phase, for low silica, low Fe samples (Fig. 4a) and the dacite-andesite, C_P of liquid can be confidently fitted as linear with temperature. For most low-Fe liquids, C_P of liquid is nearly constant, but for rhyodacite and dacite-andesite liquids, that lack Fe^{3+} , C_P increases as T increases. For the high Fe liquids, good data could be collected over a limited T -range. Samples with significant Fe cations first crystallize upon reaching the glass transition (e.g., P-MORB and Contl.), and then remelt as T rises, with the associated latent heat resulting in large variations in apparent C_P vs T . Upon reaching the relevant liquidus T , the whole sample becomes stable, crystal-free melt. These high- T data (Fig. 4bc) are accurate and are included in the fits, as shown. However, because we excluded regions of crystallization and remelting from the fits, $\partial C_P / \partial T$ for liquids of basalt-andesite FU-18 and the other Fe-rich (and Si-poor) samples is somewhat uncertain.

DISCUSSION

Relationships among physical properties and dependence on composition

Values of fragility show a positive but non-linear correlation with NBO/T (Fig. 5a), whereby highly polymerized melts have low fragility, low C_P^{conf} , and show near-Arrhenian

viscosity behavior, in contrast to depolymerized melts which are fragile, have high C_P^{conf} , and show highly non-Arrhenian behavior. Our viscosity data and calculated fragility values are consistent with previous data for similar compositions (e.g. Giordano et al. 2008). We find a roughly linear correlation between fragility and density (Fig. 5b), and between fragility and silica content (Fig. 5c). Scatter in these plots reflects variations in FeO and MgO content for example, which nominally play the same role in calculating NBO/T but have very different partial molar densities.

A direct dependence of D on more readily measured physical properties is sought in this section, in order to provide a means of interpolating or extrapolating our measurements to lavas of other chemical compositions. Because correlations exist among the various physical properties for our series of glasses (Fig. 5bc), demonstrating a direct dependence of D on any one of the various physical properties is likely equivocal.

The linear dependence of D on Si content (Fig. 6a) is probably most useful, although Si content cannot possibly represent all the changes in chemical composition from leucogranites to ultramafic lavas. The trend at 298 K is associated with phonon-phonon scattering because the high- T radiative mechanisms contribute negligibly to D at 298K per Eq. (5) and Planck's blackbody curve. Similarly, linear trends of D exist with fragility (Fig. 6b) and also density and NBO/T (not shown), in part due to the physical properties and chemical compositions being correlated (Fig. 5). The scatter for the glass trends (exemplified by Fig. 5a) can be attributed in part to uncertainties in D and in part due to low amounts of both crystallinity for the naturally occurring samples (at the percent level), and strain, which also affects thermal diffusivity values of glasses (Hofmeister and Whittington, 2012; Romine et al. 2012). However, the single largest contributor to scatter is probably the variations in glass and melt structure and density at a given

value of NBO/T. For example, melt and glass structure depend on variables such as the ratio of alkali to alkaline-earth cations, the Al/Si ratio, and Fe content and redox state, none of which are addressed in the formulation for NBO/T.

Melts are less dense than the corresponding glasses, so from Fig. 5b, one might expect D to increase during the transition: instead, D decreases across the glass transition. This behavior is due to an increase in disorder: melts have a configurational heat capacity which glasses do not, and consequently the entropy of the melts rises rapidly above T_g (Richet 1984). Dimensionally, thermal diffusivity of phonons is the product of the compressional velocity (u_p) with the mean free path (λ), e.g. (Hofmeister 2010). The smallest λ can be is the cube root of atomic volume. We use molar volume (Table 2) and compute up from a linear fit with volume to the P-wave velocity data in Whittington et al. (2012). As generally occurs (Shankland 1972), speed increases with density. The resulting trend (Fig. 6c) is opposite that expected from $D = u_p \lambda$, suggesting that another, compositionally dependent, factor must pertain to λ .

In previous studies we have noted that more polymerized melts tend to have higher D than less polymerized melts (Hofmeister et al. 2009; 2014a; Romine et al. 2012). In the present study, our samples span a wide range of polymerization, with NBO/T varying from 0 to 1. Thermal diffusivity of glass and melt decrease as fragility (Fig. 7b) and NBO/T increase (not shown). One possible explanation is that as the melt becomes less polymerized, fewer high frequency vibrations exist (e.g., the O-Si-O bending modes), so the average frequency decreases. This interpretation, where D depends on vibrational frequency, is consistent with anisotropy in chain silicates whereby D along the chains of Si tetrahedra (which involve high frequency tetrahedral modes) is ~ 1.5 times D across the chains (which involve low frequency modes of the

divalent cations: see e.g., Hofmeister 2012). Essentially, spatially extensive Si-O-Si linkages provide a “faster track” for heat transfer in polymerized glasses.

The strongest correlation of D is with fragility (Fig. 7b). We suggest that the best representation of D_{melt} is the least squares fit:

$$D_{\text{melt}} (\text{mm}^2\text{s}^{-1}) = 0.694 - 0.0082 m, \quad (6)$$

This can be used to estimate D_{melt} to within $\sim 0.04 \text{ mm}^2\text{s}^{-1}$, where the uncertainty approximates the fit, and does not address the outliers. If only chemical data are available, then the following correlation with NBO/T can be used:

$$D_{\text{melt}} (\text{mm}^2\text{s}^{-1}) = 0.44 - 0.168 \text{ NBO/T}, \quad (7)$$

which fit is accurate to within $\sim 0.06 \text{ mm}^2\text{s}^{-1}$. Again, many factors affect D , and also the chemical composition of the glass does not uniquely constrain its structure, so the above relationships are representative, rather than exact.

Five of our basalts (P-MORB, Arc bas, MORB, OIB2) have a narrow range of silica content (49.9-50.8 wt% SiO_2), which potentially permits examination of the effects of Fe, Ca and Al. The difference in D between the two Hawai’ian basalts is greater than the difference between our arc basalt and MORB samples, but OIB1 has crystallites, which affect its trend at high T and likely perturbs its ambient value. Although the effect of particular cations could be elucidated in our study of silica-rich melts (Hofmeister et al. 2014a), this was not possible for the basalts studied here, among which D varies little. The most important observation is that D is low for basalts, $\sim 0.3 \text{ mm}^2\text{s}^{-1}$, compared to $\sim 0.5 \text{ mm}^2\text{s}^{-1}$ for rhyolitic melts.

Heat capacities of glass and liquid depend linearly on density (Fig. 7) Similar relationships also exist with the other parameters (not shown) which is consistent with simple mixing models of oxide components representing C_p (e.g. Stebbins et al. 1984; Richet 1987).

Heat capacity is largely a harmonic property (it is described by the distribution of heat energy among the available states per Debye and Einstein) whereas thermal diffusivity is entirely anharmonic (it arises through phonon interactions per Debye). This fundamental difference in behavior permits reasonably accurate determination of C_P from composition or from vibrational frequencies whereas more complicated models are required for D , as follows.

Temperature dependence of D : three microscopic mechanisms

The initial decrease in thermal diffusivity for crystals with increasing temperature has been attributed to phonon scattering (e.g., Liebfried and Schlömann 1954). At ~500 K, increases in thermal diffusivity have been previously observed due to spurious radiative transfer between the heater and thermocouple, which results from the high transparency of many materials in the near-IR (see e.g., Hofmeister et al. 2007). In our LFA measurements, this unwanted ballistic radiative transfer has been removed through use of coatings and mathematical models. In our previous studies of SiO_2 glasses and high silica glasses, we found that D of the glass increases linearly with T at above ~1000 K, for samples that remained glassy at these high temperatures (Fig. 3; also see Hofmeister and Whittington 2012; Hofmeister et al. 2014a). For Fe-bearing pyroxene glasses, we found an obvious linear increase at a much lower temperature of only 700 K for $\text{Fs}_{10}\text{En}_{90}$ and the ternary $\text{Fs}_{15}\text{Wo}_{40}\text{En}_{45}$, and more subtle increases for several other samples (Hofmeister et al. 2014b). For the remelted lavas investigated here, obvious increases in D occur for the high silica content samples, in part due to the higher temperatures reached, and also for OIB2 and the arc basalt. Decreases are seen for Fe-free synthetic glasses and for OIB1 with minor crystallites (per electron microprobe). The changes are resolved due to precision within a

given run being high (precision regards relative values of D), although in most cases the change across the T -range observed is greater than the experimental uncertain in absolute values of D .

Our previous study on glasses with structures close to those of tectosilicates and with less than 4 wt % FeO showed a correlation of the fitting constant H with Fe^{2+} content and/or charge transfer between Fe^{2+} and Fe^{3+} (Hofmeister et al. 2014a). This study also showed that the existence and magnitude of positive $\partial D/\partial T$ observed for the glass was repeated for the melt. The pyroxene glasses did not have a positive correlation of F , G or H with Fe content or with the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio (Hofmeister et al., 2014b). Neither did we find such a correlation for the lava glasses.

In our previous study of low Fe glasses for which the structure is highly polymerized, we found that the HT term in Equation 3 positively correlated with Fe^{2+} content in the presence of Fe^{3+} . This behavior is consistent with a radiative process, because the high frequency electronic transitions of Fe^{2+} are coupled with vibronic transitions (e.g. Rossman, 1988). The data on pyroxene composition glasses show more variable behavior (Hofmeister et al. 2014ab) which could have been caused by small but varying amounts of crystallites in these black samples. Regarding the remelted lavas, a low value of H is seen for dacite-andesite: unlike the other remelts, this sample is green with no indication of Fe^{3+} whereas the brown and black samples have significant Fe^{3+} in addition to Fe^{2+} and high absorbance throughout the visible indicating energy exchange (Fig. 1; cf. Hofmeister et al. 2014ab). Green dacite-andesite has the lowest measurable value of H (Table 6). Electronic-vibronic coupling does not explain the high values of H for the high silica glasses with little Fe (Table 6): however, for basalts with similar compositions, the synthetics with only a trace of Fe^{2+} have lower H .

The data on remelted lava glasses, coupled with our previous datasets, suggest that three mechanisms for heat transfer exist in glasses. Existence of three mechanisms would produce the variety of behaviors discussed above. One mechanism is phonon scattering, as evident in the decrease in D above ~ 298 K. Two other mechanisms are connected with the increase in D with T at high temperature. The second mechanism is the IR polariton mechanism proposed for crystals (Hofmeister et al. 2014c), as described above, which operates in the low Fe samples, and apparently weakens as Si content decreases. The third mechanism involves electronic-vibronic coupling of charge transfer complexes ($\text{Fe}^{3+}\text{-O-Fe}^{2+}$), which operates in the high Fe, oxidized samples. Because the juxtaposition of Fe cations is sample specific, this mechanism is present to varying degrees in Fe-bearing glasses. Quantifying the role of Fe cations in heat transfer of glasses will require a suite of synthetic samples with controlled $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratios at constant total Fe content, in the absence of crystallites.

Thermal Conductivity

Thermal conductivity is the product of D , density, and C_p (Eq. 1), permitting inference of this important property from our measurements, utilizing some previous findings. However, as a consequence of this convolution, k is necessarily complicated and sample specific.

For glasses, our computations (Eq. 1) incorporate the fits to D and C_p (Tables 6 and 7). We used linearized density $\rho = \rho_{298}[1-\alpha(T-298)]$ where α is the typical, constant thermal expansivity of $25 \times 10^{-6} \text{ K}^{-1}$ (e.g. Bouhifd et al. 2015). This procedure produces a smooth curve for k vs T for each sample (Fig. 8). Uncertainties for k of glass are $<4\%$, after Bevington (1969), considering experimental errors in C_p and D , which dominate those in ρ . Again we note that

changes in k are less uncertain than the absolute values of k , due to uncertainties in D arising from sample thickness and parallelism, which are constant over each run.

For most liquids, we calculated k at the liquidus temperature, using ρ calculated for that temperature (Table 2), the fit predicting liquid C_P at that temperature, and the best measurement of D_{melt} measured above the calorimetric or viscometric glass transitions but not at liquidus temperatures. We also provide k for the liquid calculated for T_g , which is similar in most cases. For dacite-andesite and MORB only, we used their trends of D with T (Table 6) and liquid thermal expansivity of $75 \times 10^{-6} \text{ K}^{-1}$ (Bouhifd et al. 2015) to calculate $k(T)$ in Table 9. Uncertainties for k of liquid are 5-8%, possibly increasing to 10% at the liquidus, based on estimating α and the limitations in our direct measurements of D_{melt} .

Although these uncertainties in k may appear high for a modern method, our results offer considerable improvement over conventional measurements of k , which are uncertain by ~20% at 298 K, and greatly increase with T , depending on transparency (Hofmeister 2007; Hofmeister et al. 2007). Conventional measurements of thermal conductivity rarely reach temperatures above 1200 K and systematically err due to unwanted radiative transfer.

The resulting smooth trends for k with T (Fig. 8) are less regular than the trends for D (Fig. 3) or C_P (Fig. 4), in part because uncertainties of a few % for each parameter are multiplied in the convolution of Eq (1). More importantly, both D and C_P vary little with T , and in opposite directions, making the variation of k with T weak but with the sign of the slope varying. There is no indication of different mechanisms operating below and above ~400K, which was proposed based on trends for k alone (Cahill and Pohl 1987).

The results for k of glass, despite the variety of trends (Fig.8), are best fit by:

$$k = A + BT + CT^{-2} + DT^{1/2}, \quad (8)$$

where parameters are given in Table 9. This form was deduced from Maier-Kelly fits to heat capacity (reviewed by Robie et al. 1987). This discovery emphasizes that heat capacity largely controls thermal conductivity in glasses, as a consequence of thermal diffusivity depending weakly on temperature.

Trends of k with various material properties are poorly defined (see Table 9), largely because the variations in k are small. Small variations in k are due to the different dependence of C_P , D , and ρ on chemical composition and scatter (cf. Figs. 6-8). The most regular trend was seen for k with fragility, but it was unimpressive and is not shown. We did find that glass k at 298 K depends oppositely of melt k , regardless of the parameter chosen, whereas glass k at T_g is roughly independent of any parameter. From our data, k appears to depend on a combination of Si content, oxidation state of Fe, and hydration, but this remains to be tested by independently varying the parameters.

We emphasize that the fundamental parameter for heat transport is thermal diffusivity (Hofmeister et al. 2014c), and that basalts clearly have lower D than silica-rich melts (Table 6). However, thermal conductivity is important for modelling. It would be best to use our trends in D and estimate density and heat capacity in order to model compositions other than those presented here. However, rough numbers for k_{melt} are called for, since a very rough number of $1 \text{ mm}^2 \text{ s}^{-1}$ is almost universal in geophysical models. Whereas silicic to intermediate melts have k between 1.8 and $1.3 \text{ W m}^{-1} \text{ K}^{-1}$, values for basaltic melts are more restricted, $1.3\text{-}1.5 \text{ W m}^{-1} \text{ K}^{-1}$, with $1.4 \text{ W m}^{-1} \text{ K}^{-1}$ reasonably representing any temperature.

Implications for partially molten zones

The difference in thermal properties between silicic and mafic liquids is smaller than the variability between different minerals, although minerals also become more insulating at high

temperatures. Quartz and olivine have relatively high values of k and D , whereas pyroxenes and feldspars are more insulating (Branlund and Hofmeister, 2007; Branlund and Hofmeister, 2012; Hofmeister, 2012; Hofmeister and Pertermann 2008; Pertermann and Hofmeister 2006). Consequently, the thermal properties of crustal rocks depend strongly on mineralogy (Merriman et al. 2013). Many crustal rocks such as pelitic schists become very insulating at high temperatures, with $D \sim 0.56 \text{ mm}^2\text{s}^{-1}$ and $k \sim 1.84 \text{ Wm}^{-1}\text{K}^{-1}$ at typical crustal melting temperatures of $\sim 1000 \text{ K}$ (Whittington et al. 2009c). Leucogranite melts have D and k values only slightly lower, $\sim 0.52 \text{ mm}^2\text{s}^{-1}$ and $\sim 1.84 \text{ Wm}^{-1}\text{K}^{-1}$, so the effect of partial melting on thermal properties of the crust is negligible.

Taking forsterite as a proxy for mantle peridotite, D at 1500 K is $\sim 0.64 \text{ mm}^2\text{s}^{-1}$ and k is $2.83 \text{ Wm}^{-1}\text{K}^{-1}$, while values for MORB are $\sim 0.30 \text{ mm}^2\text{s}^{-1}$ and $\sim 1.25 \text{ Wm}^{-1}\text{K}^{-1}$. Consequently, partially molten zones in the mantle become more insulating as melting progresses (Fig. 9). This feedback relationship potentially increases the melt fraction produced by decompression melting under mid-ocean ridges and may contribute to the generation of large melt volumes that comprise large igneous provinces (e.g. Saunders 2005). Conversely, crystallizing magmas will become more effective conductors of heat as crystallization progresses, allowing an efficient pathway for the removal of latent heat of crystallization.

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REFERENCES CITED

- Adam, G. and Gibbs, J.H. (1965) On the temperature dependence of relaxation phenomena in glass-forming liquids. *Journal of Chemical Physics*, 43,139-146.
- Armstrong, J.T. (1995) CITZAF, A package of correction programs for the quantitative electron microbeam X-ray analysis of thick polished materials, thin films, and particles. *Microbeam Analysis*, 4, 177-200.
- Avard, G. and Alan G. Whittington (2012) Rheology of arc dacite lavas: experimental determination at low strain rates. *Bull Volcanol* (2012) 74:1039–1056
- Bevington, P.R. (1969) Data reduction and error analysis for the physical sciences. McGraw-Hill Book Co., New York.
- Blumm, J. and Opfermann, J. (2002) Improvement of the mathematical modeling of flash measurements. *High Temperature High Pressure*, 34, 515-521.
- Bouhifd, M.A., Besson, P., Courtial, P., Gérardin, C., Navrotsky, A., and Richet, P. (2007) Thermochemistry and melting properties of basalt. *Contrib. Mineral. Petrol.* 153, 689-698.
- Bouhifd, M.A., Whittington, A.G., and Richet, P. (2015) Densities and volumes of hydrous silicate melts: New measurements and predictions. *Chem. Geol.* 418, 40-50.
- Branlund J.M. and Hofmeister A.M. (2007) Thermal diffusivity of quartz to 1000 degrees C: Effects of impurities and the α - β phase transition. *Phys Chem. Minerals.* 34, 581-595.
- Branlund J.M. and A.M. Hofmeister. (2012) Heat transfer in plagioclase feldspars. *American Mineralogist*, 97, 1145-1154.
- Cahill, D.G. and Pohl, R.O. (1987) Thermal conductivity of amorphous solids above the plateau. *Phys. Rev. B* **35**, 4067 (1987).

639 Degiovanni, A., Andre, S., and Maillet, D. (1994) Phonic conductivity measurement of a semi-
 640 transparent material. In, Tong TW (ed) Thermal conductivity 22. Technomic, Lancaster, PA,
 641 pp 623-633

642 Dingwell, D.B. 1995. Viscosity and anelasticity of melts. AGU reference shelf 2, p. 209-217.

643 Ditmars DA, Ishihara S, Chang SS, Bernstein G, West ED (1982) Enthalpy and Heat-Capacity
 644 Standard Reference Material: Synthetic Sapphire (α -Al₂O₃) from 10 to 2250 K. J Res Natl
 645 Bur Standards 87: 159-163

646 Fei Y. (1995) Thermal Expansion. In Mineral Physics and Crystallography. A Handbook of
 647 Physical Constants (ed. T. J. Ahrens), 29-44. American Geophysical Union, Washington D.C.

648 Getson, J.M., and Whittington, A.G. (2007) Liquid and magma viscosity in the anorthite-
 649 forsterite-diopside-quartz system and implications for the viscosity temperature paths of
 650 cooling magmas, Journal of Geophysical Research, 112, B10203, DOI:
 651 10.1029/2006JB004812

652 Giordano, D., Russell, J.K., and Dingwell, D.B. (2008) Viscosity of magmatic liquids, a model.
 653 Earth and Planetary Science Letters, 271, 123–134.

654 Hellwig, B.M. (2006) The viscosity of dacitic liquids measured at conditions relevant to
 655 explosive arc volcanism: determining the influence of temperature, silicate composition, and
 656 dissolved volatile content. M.Sc. Thesis, University of Missouri – Columbia, 145pp.

657 Hofmeister, A.M. (2007) Pressure dependence of thermal transport properties. Proceedings of
 658 the National Academy of Science 104, 9192-9197.

659 Hofmeister, A.M. (2010) Scale aspects of heat transport in the diamond anvil cell, in
 660 spectroscopic modeling, and in Earth's mantle. *Physics of the Earth and Planetary Interiors*
 661 180, 138-147.

662 Hofmeister, A.M. (2012) Thermal diffusivity of orthopyroxenes at elevated temperature.
 663 European Journal of Mineralogy, 24, 669-681.

664 Hofmeister, A.M. and Criss R.E. (2013) Earth's interdependent thermal, structural, and chemical
 665 evolution, *Gondwana Research*. **24** 490–500

666 Hofmeister A.M. and Pertermann, M. (2008) Thermal diffusivity of clinopyroxenes at elevated
 667 temperature. European Journal of Mineralogy, 20, 537-549.

668 Hofmeister, A.M. and Whittington, A.G. (2012) Effects of hydration, annealing, and melting on
 669 heat transport properties of fused quartz and fused silica from laser-flash analysis. Journal of
 670 Non-Crystalline Solids 358, 1072-1082.

671 Hofmeister A.M., Pertermann, M., Branlund, J. and Whittington, A.G. (2006) Geophysical
 672 implications of reduction in thermal conductivity due to hydration. *Geophysical Research*
 673 *Letters* 33, doi: 10.1029/2006GL026036

674 Hofmeister, A.M., Pertermann, M., and Branlund, J.M. (2007) Thermal Conductivity of the Earth.
 675 In, Schubert G (ed). Treatise in Geophysics (Schubert G, Ed. In Chief) V. 2 Mineral Physics
 676 (Price GD, ed.). Elsevier, The Netherlands, pp 543-578

677 Hofmeister, A.M., Whittington, A.G., and Pertermann, M. (2009) Transport properties of high albite
 678 crystals and near-endmember feldspar and pyroxene glasses and melts to high temperature.
 679 Contributions to Mineralogy and Petrology, 158, 381-400.

680 Hofmeister, A.M. Jonas Goldsand, Alan G. Whittington, and Reinhardt G. Criss. (2014a)
 681 Effects of chemical composition and temperature on transport properties of silica-rich
 682 glasses and melts. *American Mineralogist* **99**, 564-577

683 Hofmeister, A.M., Sehlke, A. and Whittington, A.G. (2014b) Thermal diffusivity of Fe-rich
 684 pyroxene glasses and their melts. *Chemical Geology* 384, 1-9.

685 Hofmeister, A.M., Dong, J., Branlund, J.M., (2014c). Thermal diffusivity of electrical insulators
 686 at high temperatures: evidence for diffusion of phonon-polaritons at infrared frequencies
 687 augmenting phonon heat conduction, *Journal of Applied Physics* **115**, #163517;
 688 <http://dx.doi.org/10.1063/1.4873295>
 689 Julian, C.L. (1965) Theory of heat conduction in rare-gas crystals. *Physical Review A*, 137, 128-137.
 690 Lange, R.A. (1997) A revised model for the density and thermal expansivity of K₂O-Na₂O-CaO-
 691 MgO-Al₂O₃-SiO₂ liquids from 700 to 1900 K, extension to crustal magmatic temperatures.
 692 *Contributions to Mineralogy and Petrology*, 130, 1-11.
 693 Lev, E., Marc Spiegelman, Robert J. Wysocki, Jeffery A. Karson (2012) Investigating lava flow
 694 rheology using video analysis and numerical flow models. *Journal of Volcanology and*
 695 *Geothermal Research* 247–248, 62–73.
 696 Liebfried, G. and Schlömann, E. (1954) Warmleitund in elektrische isolierenden Kristallen.
 697 *Nachrichten von der Gesellschaft der Wissenschaften zu Göttingen Mathematik und Physik*
 698 K1,71–93.
 699 Maier, C. G. and Kelly, K. K. (1932) An equation for the representation of high temperature heat
 700 content data. *J. Am. Chem. Soc.* 54, 3243– 3246.
 701 Mehling, H., Hautzinger, G., Nilsson, O., Fricke, J., Hofmann, R., and Hahn, O. (1998) Thermal
 702 diffusivity of semitransparent materials determined by the laser-flash method applying a new
 703 mathematical model. *International Journal of Thermophysics*, 19, 941-949.
 704 Merriman, J.D., Whittington, A.G., Hofmeister, A.M., Nabelek, P.I., and Benn, K. (2013) Thermal
 705 transport properties of major Archean rock types to high temperature and implications for
 706 cratonic geotherms. *Precambrian Research* 233, 358-372.

707 Nabelek, P.I., Whittington, A.G., and Hofmeister, A.M. (2012) The influence of temperature-
708 dependent thermal diffusivity on the conductive cooling rates of plutons and temperature-
709 time paths in contact aureoles. *Earth and Planetary Science Letters*, 317-318, 157-164.

710 Neuville, D.R., and Richet, P. (1991) Viscosity and mixing in molten (Ca, Mg) pyroxenes and
711 garnets. *Geochimica et Cosmochimica Acta* 55, 1011–1019.

712 Parker, J.W., Jenkins, J.R., Butler, P.C., and Abbott G.I. (1961) Flash method of determining
713 thermal diffusivity, heat capacity, and thermal conductivity. *Journal of Applied Physics*. 32,
714 1679–1684.

715 Pertermann. M., and Hofmeister, A.M. (2006) Thermal diffusivity of olivine-group minerals at high
716 temperatures. *American Mineralogist*, 91, 1747-1760.

717 Pertermann. M., Whittington, A.G., Hofmeister, A.M., Spera, F.J., and Zayak, J. (2008) Thermal
718 diffusivity of orthoclase glasses and single-crystals at high temperatures. *Contributions to*
719 *Mineralogy and Petrology*, 155, 689-702. doi, 10.1007/s00410-007-0265-x

720 Plazek. D.J. and Ngai, K.L. (1991) Correlation of polymer segmental chain dynamics with
721 temperature dependent time-scale shifts. *Macromolecules*, 24, 1222–1224.

722 Richet, P. (1984) Viscosity and configurational entropy of silicate melts. *Geochimica Cosmochimica*
723 *Acta*, 48, 471-483.

724 Richet, P. (1987) Heat capacity of silicate glasses. *Chemical Geology* 62, 111-124.

725 Richet, P., Robie, R.A., and Hemingway, B.S. (1986) Low-temperature heat capacity of diopside
726 glass (CaMgSi₂O₆), a calorimetric test of the configurational entropy theory applied to the
727 viscosity of liquid silicates. *Geochimica Cosmochimica Acta*, 50, 1521-1533.

728 Robert, G. (2014) The effects of volatiles on the viscosity and heat capacity of calc-alkaline basaltic
729 and basaltic andesite liquids. Ph.D. Dissertation, University of Missouri – Columbia, 197pp.

730 Robert, G., Whittington, A.G., Stechern, A., and Behrens, H. (2013) The effect of water on the
 731 viscosity of a synthetic calc-alkaline basaltic andesite. *Chemical Geology*, 346, 135-148.
 732 Robert, G., Whittington, A.G., Stechern, A., Behrens, H. (2014) Heat capacity of hydrous glasses and
 733 liquids. *Journal of Non-Crystalline Solids*, 390, 19-30.
 734 Robert, G., Knipping, J.L., Scherbarth, S., Robertson, T.E., Stechern, A., Behrens, H. and
 735 Whittington, A.G. (2015) Heat capacity and viscosity of basaltic melts with $\text{H}_2\text{O} \pm \text{F} \pm \text{CO}_2$,
 736 *Chemical Geology*, 418, 51-65.
 737 Robie, R.A., Hemingway, B.S., and Fisher, J.R. (1978) Thermodynamic properties of minerals and
 738 related substances at 298.15 K and 1 bar (10^5 pascals) and at higher temperatures. U.S.
 739 Geological Survey Bulletin, 1452, 456 pp.
 740 Romine, W.L. and Whittington, A.G. (2015) A simple model for the viscosity of rhyolites as a
 741 function of temperature, pressure and water content. *Geochim Cosmochim Acta*, 170, 281-
 742 300, doi:10.1016/j.gca.2015.08.009
 743 Romine, W.L., Whittington, A.G., Nabelek, P.I., and Hofmeister, A.M. (2012) Thermal
 744 diffusivity of rhyolitic glasses and melts, effects of temperature, crystals and dissolved water.
 745 *Bull Volc* 74, 2273-2287, doi:10.1007/s00445-012-0661-6
 746 Russell, J.K., Giordano, D., and Dingwell, D.B. (2003) High-temperature limits on viscosity of non-
 747 Arrhenian silicate melts. *American Mineralogist*, 88, 1390–1394.
 748 Saunders, A.D. (2005) Large Igneous Provinces: Origin and environmental consequences.
 749 *Elements*, 1, 259-263
 750 Schuessler J. A., Botcharnikov R. E., Behrens H., Misiti V., Freda C. (2008) Oxidation state of
 751 iron in hydrous phonotephritic melts, *American Mineralogist* 93, 1493–1504.
 752 Sehlke, A., Whittington, A.G., Robert, B., Harris, A., Gurioli, L., and Médard, E. (2014)

Pahoehoe to 'a'a transition of Hawaiian lavas: an experimental study. *Bulletin of Volcanology*, 76, 876-896

Shankland, T.J. (1972) Velocity-density systematics: derivation from Debye theory and the effect of ionic size. *Journal of Geophysical Research*, 77, 3750-3758.

Whittington, A.G., Bouhifd, M.A., and Richet, P. (2009a) The viscosity of hydrous NaAlSi₃O₈ and granitic melts, configurational entropy models. *American Mineralogist* 94, 1-16.

Whittington AG, Hellwig BM, Behrens H, Joachim B, Stechern A, Vetere F (2009b) The viscosity of hydrous dacitic liquids: implications for the rheology of evolving silicic magmas. *Bull Volcanol* 71:185–199.

Whittington AG, Hofmeister AM, Nabelek P (2009c) Temperature-dependent thermal diffusivity of Earth's crust and implications for magmatism. *Nature*, 458, 319-321

Whittington A.G., Richet P., Polian A. (2012) Water and the compressibility of silicate glasses: a Brillouin spectroscopic study. *Amer Mineral* 97: 455-467.

Wirth, K.R., Vervoort, J.D., Naiman, Z.J., 1997. The Chengwatana Volcanics Wisconsin and Minnesota: petrogenesis of the southernmost volcanic rocks exposed in the Midcontinent Rift. *Can J Earth Sci* 34: 536-548.

FIGURE CAPTIONS

Fig. 1. Spectra of various electronic transitions in the UV-visible region (a) and of O-H and Fe²⁺ d-d transitions visible-IR region (b). Curves as labeled. Vertical lines indicate peak assignments. Samples have oxidation states that are not dictated by total FeO. Original thicknesses in mm are 9.41 (rhyodacite), 0.47 and 1.22 (dacite), 0.36 (dacite-andesite), 8.35 (synthetic andesite-basalt,

SBA), 0.21 (basalt-andesite, Fu18), 0.2 (dolerite), 3.55 (synthetic basalt, SB), and 0.040 (arc basalt, Fu06).

Fig. 2. Viscosity data for supercooled liquids near the glass transition and melts. Symbols are as listed where circles are silica-rich melts, squares are intermediate lavas, and diamonds are basic. Solid lines = VFT curves. Dashed line = the 10^{12} Pa s isokom.

Fig. 3. Thermal diffusivity vs. temperature showing fits to $D=FT^{-G}+HT$. Melt values (not included in the fits) are shown with larger symbols. Except for part (a), symbols are described in the legend. **(a)** High silica glasses. From top to bottom: plus sign and thick line = leucogranite (after Hofmeister et al. 2014a). Various circles and fine line = rhyolite (Romine et al. 2012). Squares and dotted line = rhyodacite. Grey triangles and grey line = synthetic dacite. Vertical lines indicate T_{12} , which differs from the glass-to-liquid conversion in LFA due to rapidity of measurements (see text or Pertermann et al. 2009). **(b)** Intermediate Si contents. For andesite-basalt, data were collected from a slab before and after concentric cylinder (cc) measurements. The post-cc measurements were single data points rather than averaging 3 collections, except at 298K, where 13 measurements were averaged. Thin line = linear fit to the andesite-dacite melt data. **(c)** Basalts from island and arc environments. **(d)** Mid-ocean ridge and continental basalts.

Fig. 4. Heat capacity vs. temperature showing Maier-Kelly fits for the glass and linear fits for the liquid. Insets tie symbols to compositions. An increase in C_p with T was clearly observed for rhyodacite, and dacite-andesite.

Fig. 5. Relationships of fragility with various physical properties. Data from Tables 1, 2 and 4. Least squares fits are listed in each panel. **(a)** Non-linear dependence of fragility on NBO/T, fitted with a third-order polynomial. **(b)** Fragility and NBO/T vs. density. Fragility varies approximately linearly with density, whereas NBO/T is better fitted with a quadratic equation.

(c) Dependence of fragility (triangles), NBO/T (plotted as $\times 100$, circles), and density (squares) on silica content, with linear fits.

Fig. 6. Dependence of thermal diffusivity on various physical properties. Least squares fits are shown. Filled triangles and solid line = glass data. Open diamonds and dotted line = melt data. Vertical dotted line (or grey area) separates basalts from the other remelts. Vertical arrow indicates that a minimum value was observed. (a) D vs SiO_2 content. (b) D vs. fragility. (c) D vs parameters expected to contribute to heat transfer, velocity of carriers and a representative distance.

Fig. 7. Dependence of heat capacity on density for each of glass and melts. Least squares fits are listed.

Fig. 8. Thermal conductivity vs. temperature, computed using Eq. 1. In each part, different patterns are used for each composition, as labeled, and liquid is shown as a heavy line with the same pattern as the glass. The computed values are smooth because the inputs are smooth. Fits to Eq. (7) are indistinguishable from the data for most samples. A few examples of fits to glass are shown (grey curves).

Fig. 9. Thermal diffusivity and conductivity in partially molten zones. “Mantle” uses forsterite for solid end-member and MORB for liquid end-member. “Crust” uses schist for solid end-member and rhyolite for liquid end-member. Dashed curves are for “resistors in series” model, while solid curves are for “resistors in parallel” model, representing upper and lower bounds depending on melt geometry. The two are indistinguishable for crustal melting.

824 Table 7: Isobaric heat capacities of glasses and liquids measured by differential scanning
825 calorimetry (DSC) every 30 degrees
826

Leucogranite Harney Peak (HP96)		Rhyolite Mono Craters (ANCAR)		Rhyodacite synthetic (SRD)		Dacite Mt. St. Helens (SH305)	
Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)
338	0.961	338	1.023	338	0.834	337	0.977
361	0.917	361	0.959	361	0.850	360	0.928
389	0.930	388	0.956	388	0.873	388	0.919
419	0.960	418	0.969	418	0.899	417	0.933
450	0.988	450	0.986	450	0.925	449	0.950
483	1.016	482	1.012	482	0.949	481	0.967
516	1.048	515	1.043	515	0.974	514	0.991
549	1.079	548	1.066	548	0.996	547	1.011
582	1.105	581	1.085	581	1.015	580	1.026
614	1.124	613	1.101	614	1.031	612	1.045
647	1.147	646	1.114	646	1.043	645	1.063
679	1.168	678	1.124	678	1.055	677	1.080
710	1.181	709	1.133	710	1.071	709	1.093
742	1.194	741	1.141	742	1.084	740	1.107
773	1.212	772	1.155	773	1.094	771	1.124
804	1.223	803	1.168	804	1.108	803	1.141
835	1.234	834	1.169	836	1.117	834	1.154
866	1.246	865	1.173	866	1.125	865	1.171
897	1.252	896	1.178	897	1.137	895	1.187
928	1.259	926	1.182	928	1.151	926	1.198
958	1.277	957	1.183	958	1.160	957	1.210
989	1.302	987	1.188	989	1.186	987	1.221
1019	1.330	1018	1.213	1020	1.222	1018	1.259
1049	1.352	1048	1.270	1050	1.331	1048	1.396
1080	1.411	1079	1.343	1080	1.490	1079	1.550
1110	1.540	1109	1.386	1111	1.398	1109	1.429
1140	1.551	1139	1.353	1141	1.379	1139	1.421
1171	1.481	1170	1.341	1171	1.383	1170	1.420
1201	1.489	1200	1.338	1201	1.382	1200	1.414
1231	1.495	1230	1.348	1231	1.381	1230	1.429
1261	1.496	1261	1.354	1261	1.414	1260	1.430
1291	1.472	1291	1.346	1291	1.401	1291	1.454
1321	1.481	1321	1.342	1321	1.419	1321	1.452
1351	1.483	1352	1.367	1352	1.424	1351	1.440
1381	1.497	1382	1.368	1382	1.428	1381	1.438
1412	1.479	1412	1.416	1412	1.427	1412	1.424
1442	1.481	1442	1.519	1442	1.430	1442	1.424
1472	1.461	1472	1.593	1472	1.415	1472	1.433
1502	1.429	1503	1.615	1502	1.337	1502	1.437
1532	1.417	1533	1.632	1532	1.264	1533	1.436
1562	1.403	1564	1.612	1562	1.207	1563	1.400
1592	1.401	1594	1.571	1592	1.191	1593	1.369
1622	1.406	1624	1.535	1622	1.173	1624	1.300
1653	1.393	1654	1.507	1653	1.170	1654	1.261
1683	1.388	1685	1.446	1683	1.165	1684	1.240
1713	1.372	1715	1.365	1713	1.158	1715	1.225

1744	1.344	1746	1.351	1744	1.149	1745	1.198
1761	1.311	1763	1.356	1760	1.145	1762	1.228

Table 7: continued

Dacite-andesite Santiaguito (SA05-14)		Andesite-basalt* SBA		Basalt-andesite* Fuego (Fu18)		Dolerite Chengwatana (CHW)		Basalt* SB	
Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)
337	1.087	386	0.943	388	0.905	338	0.755	386	0.962
360	1.043	406	0.941	413	0.910	361	0.833	406	0.958
387	1.031	430	0.957	442	0.929	388	0.877	430	0.973
417	1.047	455	0.975	474	0.953	418	0.911	455	0.991
449	1.069	482	0.993	506	0.975	450	0.938	482	1.011
481	1.086	509	1.010	539	0.992	482	0.966	509	1.029
514	1.105	536	1.027	571	1.007	515	0.993	536	1.045
547	1.123	564	1.042	604	1.026	547	1.023	563	1.056
580	1.139	591	1.057	636	1.040	580	1.044	591	1.066
612	1.156	618	1.073	668	1.055	613	1.059	618	1.076
645	1.177	645	1.090	700	1.072	645	1.071	644	1.089
677	1.196	671	1.099	732	1.084	677	1.089	671	1.095
708	1.212	698	1.107	763	1.098	709	1.104	698	1.112
740	1.229	724	1.118	794	1.104	741	1.110	724	1.122
771	1.248	750	1.131	825	1.109	772	1.121	750	1.136
802	1.261	776	1.140	856	1.114	803	1.135	776	1.151
833	1.274	802	1.146	887	1.110	834	1.145	802	1.159
864	1.289	828	1.155	917	1.108	865	1.149	828	1.165
895	1.309	854	1.165	948	1.167	896	1.148	854	1.167
926	1.334	880	1.164	978	1.397	927	1.208	879	1.175
956	1.389	905	1.160	1009	1.497	957	1.481	905	1.176
987	1.561	931	1.153	1039	1.463	988	1.565	931	1.165
1017	1.690	956	1.153	1070	1.447	1018	1.524	956	1.165
1048	1.609	981	1.180	1100	1.437	1049	1.511	981	1.211
1078	1.609	1007	1.326	1130	1.419	1079	1.496	1007	1.415
1108	1.615	1032	1.581	1160	1.364	1109	1.484	1032	1.652
1139	1.618	1057	1.567	1190	1.271	1140	1.475	1057	1.610
1169	1.649	1082	1.553	1221	0.752	1170	1.454	1082	1.598
1199	1.649	1108	1.552	1251	0.135	1200	1.247	1108	1.604
1229	1.640	1133	1.550	1281	0.353	1231	0.267	1133	1.604
1260	1.636	1158	1.549	1311	0.828	1261	-1.171	1158	1.606
1290	1.663	1183	1.551	1341	1.132	1291	0.170	1183	1.600
1320	1.684	1208	1.553	1371	1.112	1321	1.573	1208	1.568
1351	1.670	1233	1.508	1401	1.612	1352	2.169	1233	1.505
1381	1.672	1258	1.399	1431	1.923	1382	2.938	1258	1.352
1411	1.668	1283	1.243	1461	1.904	1412	3.695	1283	1.117
1441	1.689	1308	1.067	1491	2.178	1442	2.933	1308	0.826
1472	1.745	1334	0.880	1521	2.364	1473	1.708	1333	0.450
1502	1.799	1359	0.665	1551	2.529	1503	1.489	1358	0.036
1532	1.808	1384	0.584	1582	1.486	1533	1.458	1384	-0.097
1562	1.793	1409	0.847	1612	1.299	1563	1.473	1409	0.142
1593	1.743	1434	1.400	1642	1.304	1594	1.486	1434	0.642
1623	1.694	1459	2.368	1672	1.294	1624	1.478	1459	1.248
1653	1.673	1484	3.487	1702	1.291	1654	1.498	1484	1.999

1684	1.672	1509	2.403	1732	1.290	1684	1.520	1509	3.333
1714	1.674	1534	2.093	1752	1.287	1715	1.527	1534	5.050
1744	1.685	1559	1.444			1745	1.524	1559	2.046
337	1.087	1584	1.396					1584	2.157
		1609	1.390					1609	2.271
		1634	1.381					1634	1.817
		1659	1.378					1659	1.430
		1684	1.374					1684	1.404
		1710	1.370					1709	1.396
		1735	1.359					1735	1.399

Table 7: continued

Arc basalt* Fuego (Fu06)		MORB E. Pacific Rise (14Nov04)		P-MORB Laki		OIB Mauna Ulu (MU-SQ-4)		Continental* Nyiragongo (NYI)	
Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)	Temp. (K)	Cp (J g ⁻¹ K ⁻¹)
388	388	338	0.858	338	1.029	336	1.005	336	0.966
413	413	361	0.873	361	0.971	359	0.973	359	0.913
442	442	389	0.897	388	0.951	386	0.985	387	0.921
474	474	419	0.924	418	0.950	416	1.003	417	0.937
506	506	450	0.951	450	0.958	447	1.023	448	0.957
539	539	483	0.975	482	0.964	479	1.042	481	0.973
572	572	516	1.000	515	0.975	512	1.060	514	0.983
604	604	549	1.021	547	0.986	545	1.080	547	0.998
636	636	581	1.035	580	0.999	578	1.098	580	1.008
668	668	614	1.056	613	1.010	611	1.115	612	1.013
700	700	646	1.082	645	1.019	643	1.125	645	1.026
732	732	679	1.101	677	1.029	675	1.138	677	1.031
763	763	710	1.119	709	1.042	707	1.154	708	1.037
794	794	742	1.144	741	1.047	739	1.172	739	1.048
825	825	773	1.172	772	1.056	770	1.190	771	1.058
856	856	805	1.193	803	1.071	801	1.205	802	1.060
887	887	836	1.215	834	1.075	832	1.218	833	1.054
918	918	866	1.228	865	1.087	863	1.225	864	1.038
948	948	897	1.241	896	1.096	894	1.224	894	1.045
979	979	928	1.263	927	1.128	925	1.250	925	1.167
1009	1009	958	1.420	957	1.232	956	1.500	955	1.290
1039	1039	988	1.680	988	1.425	986	1.763	986	1.038
1070	1070	1019	1.670	1018	1.419	1017	1.688	1016	1.032
1100	1100	1049	1.668	1049	1.393	1047	1.644	1047	1.261
1130	1130	1079	1.659	1079	1.311	1077	1.615	1077	1.428
1161	1161	1109	1.676	1109	0.969	1107	1.188	1108	1.213
1191	1191	1139	1.682	1140	-0.114	1138	-1.809	1138	-2.178
1221	1221	1170	-0.098	1170	-0.149	1168	-2.902	1168	0.920
1251	1251	1201	-2.753	1200	1.013	1198	1.227	1199	1.124
1281	1281	1230	0.995	1231	1.327	1229	1.374	1229	1.338
1311	1311	1260	1.544	1261	1.504	1259	1.438	1259	1.501
1341	1341	1290	1.322	1291	1.732	1289	1.563	1290	1.666
1371	1371	1320	1.252	1321	2.120	1320	1.810	1320	1.995
1401	1401	1351	1.343	1352	2.844	1350	2.238	1350	2.492
1431	1431	1381	1.925	1382	3.619	1380	2.749	1380	3.867

1461	1461	1411	3.309	1412	3.357	1410	3.600	1411	3.453
1491	1491	1441	6.118	1442	2.102	1441	4.602	1441	1.636
1521	1521	1471	5.222	1473	2.080	1471	3.110	1472	1.678
1552	1552	1501	1.664	1503	1.969	1501	2.115	1502	1.790
1582	1582	1531	-0.121	1533	1.557	1532	1.569	1532	1.851
1612	1612	1562	-0.199	1563	1.544	1562	1.575	1563	1.570
1642	1642	1592	-0.352	1594	1.569	1592	1.572	1593	1.568
1672	1672	1622	-0.154	1624	1.558	1623	1.552	1623	1.564
1702	1702	1652	0.006	1654	1.524	1653	1.556	1653	1.559
		1682	0.027	1684	1.489	1683	1.549	1684	1.550
		1713	0.005	1715	1.463	1714	1.535	1714	1.512
		1743	0.076	1745	1.426	1744	1.531	1745	1.417
		1760	0.089						

Note: Glassy starting materials always showed an increase in C_p at the glass transition temperature, followed by a large negative apparent C_p , that corresponds to latent heat released during crystallization (ΔH^{xtl}) of the sample. At higher temperatures we observed an abrupt increase in the heat flow signal, corresponding to melting of the crystals, followed by a return to the liquid heat capacity of $\sim 1.5 \text{ Jg}^{-1}\text{K}^{-1}$.

*Data below 1200 K from Robert et al. (2014) whereas data above 1200 K from Robert (2014)

Table 1. Sample descriptions

Lava type	Abbrev.	Density ^a kg m ⁻³	Color ^b	Type and Origin	Previous Sample No.	Reference
Leucogranite	leuco	2338	pale brown	Remelted Harney Peak, SD	Leuco	<i>D</i> (Hofmeister et al., 2014a); η (Whittington et al., 2009a)
Rhyolite	rhyo	2333	lt. brown	Remelted Mono Craters, CA	NCAR	<i>D</i> (Romine et al., 2012); η (Romine and Whittington, 2015)
Rhyodacite	rhyo-dac	2415	colorless	Synthetic	SRD	-
Dacite	dac	2632	brown	Remelted Mt. St. Helens	SH305	-
Dacite-andesite	dac-and	2630	green	Remelted Santiaguito, Guatemala	SA05-14	-
Andesite	and	2459	colorless	Synthetic	BD	-
Andesite-basalt	and-bas	2611	colorless	Synthetic	SBA	η (Robert et al., 2013)
Basalt-andesite	bas-and	2655	brown	Remelted Fuego, Guatemala	Fu18	-
Dolerite	dolerite	2740	brown	Remelted Chengwatana, WI ^c	Lab. Expt.	-
Basalt	bas	2679	colorless	Synthetic	SB	η at $T < 1000\text{K}$ (Robert et al., 2015); η at $T > 1000\text{K}$ (Robert, 2014)
Arc basalt	arc bas	2770	black	Remelted Fuego, Guatemala	Fu06	η at $T < 1000\text{K}$ (Robert et al., 2015); η at $T > 1000\text{K}$ (Robert, 2014)
Ocean island	OIB1	2822	black	Remelted Mauna Kea quarry, HI	WU31-108	-
Ocean island	OIB2	2816	black	Remelted Mauna Ulu, HI	MU-SQ-4	η (Sehlke et al., 2014)
MORB	MORB	2783	black	Glassy rim, East Pacific Rise ^d	14Nov04	-
Icelandic	P-MORB	2753	black	Remelted lava tube roof, Iceland	LAKI	-
Continental basalt	contl	2873	black	Remelted Mt. Nyiragongo, DR. Congo ^e	NYI001	-

^a At $\sim 298\text{ K}$. The last digit is uncertain, except for MORB for which three different pieces with varying water contents gave a range from 2765 to 2798 kg m⁻³. Details on MORB will be presented elsewhere.

^b From $\sim \text{mm}$ thick sections

^c Sample described by Lev et al. (2012). Successive remelts of a continental basalt incorporated silica sand: the composition provided here is for the material studied here.

^d Collected in an Alvin dive by Mitch Schulte

^e Sample provided by Benoît Smets and Matthieu Kervyn

Table 2. Chemical compositions determined by electron microprobe (in wt %)

	Leuco ^a	Rhyo ^b	Rhyo-dac	Dacite	Dac-and	And ^c	And-bas ^d	Bas-and ^d	Dolerite
SiO ₂	77.72	76.53	68.41	62.16	64.04	61.05	54.42	53.08	53.02
TiO ₂	0.07	0.07	0.49	0.58	0.57	0.71	1.37	0.88	1.69
Al ₂ O ₃	13.80	12.92	18.38	19.83	18.80	22.19	20.49	20.42	15.68
Cr ₂ O ₃	bld	nm	bld	0.01	nm	bld	nm	nm	0.02
FeO	1.00	0.99	bld	3.83	3.29	bld	bld	7.50	9.86
MnO	0.06	0.06	bld	0.05	0.14	bld	bld	bld	0.17
MgO	0.04	0.02	1.57	1.70	2.17	1.46	8.40	3.21	6.40
CaO	0.46	0.97	4.47	4.46	5.77	7.82	10.39	8.94	9.04
Na ₂ O	3.26	3.81	4.43	4.46	4.45	4.33	3.76	3.93	2.55
K ₂ O	4.59	4.69	2.59	1.38	1.60	1.48	1.06	0.83	0.68
P ₂ O ₅	nm	nm	nm	0.16	nm	nm	nm	0.20	0.12
Total	101.09	100.06	100.34	98.69	100.85	99.04	99.90	98.99	99.24
NBO/T	0	0.02	0.05	0.09	0.20	0.06	0.40	0.33	0.58
V _{molar} , cm ³	27.81	27.70	27.08	26.49	26.33	26.57	24.76	25.25	23.93
ρ_{liquid} , kg m ⁻³	2335	2338	2402	2487	2472	2479	2538	2592	2646
T _{liquid} , K	1173	1173	1223	1223	1273	1273	1423	1423	1473

Notes: bld = below limit of detection; nm = not measured. Compositions are the average of between 5 and 10 analyses for each sample. Analytical totals below 100% in part reflect that some iron is present as Fe₂O₃. Totals are low only for PMorb, which has a very small amount of crystallites. NBO/T calculated after Mysen et al. (1982a) approximates glass structure. NBO/T uncertainties (2 σ) are ± 0.01 based on chemical analysis standard deviations. Density of the liquid calculated from the composition at the indicated liquidus temperature using the formulation of Lange (1997).

^aAnalysis by Hofmeister et al. (2014a)

^bAnalysis by Romine et al. (2012)

^cAnalysis of andesite from sample BD of Hellwig (2006). Similar values of D were also measured for sample DF-05 of Hellwig (2006), which is the same base composition containing 0.03 wt% F.

^dAnalysis from Robert et al. (2013)

^eAnalysis from Robert et al. (2015)

Table 2 continued.

[illegible]

Table 3. Viscosity measurements

Rhyo-dac			Dacite			Dac-and		Bas-and ^c		Dolerite	
T (K)	$\log \eta$ (Pa s)		T (K)	$\log \eta$ (Pa s)		T (K)	$\log \eta$ (Pa s)	T (K)	$\log \eta$ (Pa s)	T (K)	$\log \eta$ (Pa s)
1021	12.38	a	1009	12.22	a	967.8	11.97	932.8	12.75	935.9	12.00
1030	12.19	a	1030	11.66	b	974.4	11.74	942.8	12.22	950.1	11.37
1041	11.86	a	1040	11.39	b	985.2	11.37	953.0	11.71	959.8	10.97
1051	11.60	b	1046	11.20	b	990.8	11.19	962.7	11.35	963.6	10.76
1061	11.27	a	1050	11.08	b	994.6	11.06	972.7	10.99	969.8	10.50
1071	11.06	b	1050	11.01	a	1006	10.72	982.9	10.65	975.1	10.32
1082	10.70	a	1055	10.87	a	1010	10.61	993.0	10.32	982.4	10.07
1091	10.55	b	1061	10.78	b	1021	10.28	1003	10.00	983.9	9.98
1096	10.33	a	1065	10.67	b	1025	10.16	1013	9.71	989.5	9.83
1111	10.06	b	1070	10.49	a	1031	10.00	1025	9.38	994.5	9.63
1114	9.92	a	1071	10.47	a	1035	9.88	1518	2.41 c	994.7	9.65
1115	9.89	a	1076	10.39	b	1045	9.62	1544	2.24 c	998.8	9.49
1121	9.82	b	1081	10.26	b	1065	9.12	1570	2.09 c	1005	9.29
1124	9.69	a	1086	10.12	b	1066	9.13	1595	1.94 c	1013	9.03
1126	9.73	b	1090	10.01	b	1431	3.93	1620	1.81 c	1013	8.87
1129	9.59	a	1095	9.88	a	1451	3.74	1669	1.55 c	1023	8.70
1135	9.50	a	1100	9.76	b	1470	3.57	1720	1.32 c	1033	8.16
1136	9.51	b	1106	9.64	a	1490	3.40	1745	1.21 c	1478	2.00
1136	9.51	b	1108	9.57	a	1514	3.21	1770	1.10 c	1502	1.84
1139	9.39	a	1116	9.42	a	1539	3.03	1796	0.99 c	1548	1.55
1144	9.28	a	1120	9.32	a	1563	2.86	1803	0.97 c	1595	1.29
1146	9.30	b	1120	9.30	b	1587	2.70	1827	0.88 c	1643	1.06
1155	9.08	a	1403	4.93		1612	2.55			1692	0.85
1162	8.94	a	1413	4.80		1636	2.41			1741	0.65
1489	4.53		1432	4.59		1660	2.27			1789	0.47
1508	4.37		1452	4.40		1684	2.15			1838	0.31
1528	4.17		1471	4.17		1685	2.14			1857	0.25
1530	4.25		1491	4.01		1709	2.01				
1548	4.06		1510	3.87		1733	1.88				
1567	3.91		1530	3.70		1757	1.76				

1575	3.89	1549	3.54	1782	1.66
1587	3.77	1569	3.39		
1606	3.63	1588	3.25		
1625	3.53	1627	2.99		
1626	3.49	1647	2.87		
1645	3.36	1685	2.63		
1665	3.23	1705	2.53		
1684	3.11	1724	2.42		
1684	3.12	1743	2.33		
1704	2.99	1773	2.19		
1723	2.88	1774	2.16		
1734	2.81				
1743	2.77				
1762	2.67				
1773	2.63				
1774	2.56				

Notes: Uncertainties discussed in the text.

a,b indicate two different cores in the parallel plate viscometer.

x indicates unexpectedly high values that suggest crystallization.

c Samples FU18 (bas-and) and FU06 (arc basalt) from Robert (2014)

Table 3, continued.

Arc basalt ^c			OIB1		MORB		P-MORB			Continental		
<i>T</i> (K)	log η (Pa s)		<i>T</i> (K)	log η (Pa s)	<i>T</i> (K)	log η (Pa s)	<i>T</i> (K)	log η (Pa s)		<i>T</i> (K)	log η (Pa s)	
912.4	13.13		978.8	12.48	959.1	11.22 x	903.1	12.79	a	934.4	12.13	a
922.9	12.49		993.6	11.49	964.9	10.87 x	912.2	12.16	a	943.2	11.76	b
932.5	11.93		1003	10.90	973.5	10.59	922.4	11.76	a	947.5	11.55	a
947.6	11.22		1013	10.36	975.3	10.56	925.4	11.85	b	953.5	11.19	a
958.0	10.80		1018	9.97	984.9	10.14	927.0	11.54	a	954.6	11.22	a
968.0	10.40		1033	9.55	987.3	10.04	931.3	11.21	a	968.0	10.76	b
978.2	10.04				995.9	9.72	941.7	10.94	a	974.8	10.50	b
984.0	9.80				998.4	9.59	945.8	10.93	b	980.3	10.29	b
994.3	9.48				1001	9.49	946.5	10.68	a	980.7	10.21	a
1467	2.03	c			1006	9.38	951.1	10.37	a	985.3	10.09	b
1493	1.80	c			1016	8.91	961.4	10.09	a	990.3	9.89	b
1519	1.63	c			1024	8.76	966.0	10.01	a	995.1	9.69	b
1544	1.49	c			1579	2.19	969.6	9.96	b	995.8	9.64	a
1569	1.35	c			1580	2.19	970.6	9.96	b	1001	9.45	b
1596	1.21	c			1580	2.19	974.8	9.42	a	1006	9.42	b
1621	1.09	c			1621	1.95	985.0	9.21	a	1008	9.13	b
1647	0.97	c			1622	1.95	985.1	9.42	b	1012	9.06	a
1671	0.86	c			1669	1.70	988.6	9.22	b	1026	8.49	a
1696	0.76	c			1670	1.70	990.4	9.19	b	1038	7.93	a
1722	0.66	c			1675	1.70	992.1	8.79	a	1477	1.78	
1747	0.56	c			1698	1.57	1002	8.79	b	1502	1.44	
1773	0.47	c			1698	1.57	1003	8.74	b	1502	1.45	
1798	0.39	c			1699	1.57	1009	8.23	a	1502	1.42	
1807	0.36	c			1718	1.47	1394	3.89	x	1548	1.11	
1828	0.30	c			1718	1.48	1443	2.98	x	1549	1.11	
					1719	1.49	1465	2.85	x	1550	1.12	
					1767	1.27	1492	2.34		1596	0.86	
					1767	1.28	1514	2.23		1597	0.86	
					1767	1.29	1541	1.98		1597	0.86	
					1784	1.20	1590	1.72		1645	0.62	

1785	1.20	1614	1.57	1646	0.64
1785	1.20	1638	1.49	1646	0.64
1816	1.10	1687	1.29	1694	0.40
1816	1.08	1715	1.16	1694	0.42
1816	1.09	1736	1.11	1694	0.40
1858	0.97	1774	1.02	1742	0.20
1858	0.96	1775	1.03	1743	0.23
1861	0.93			1743	0.21
1861	0.94			1790	0.02
				1791	0.00
				1825	-0.09

Table 4. Fits to viscosity data of the form $\log(\eta)=A+B/(T-C)$

sample	Leuco ^a	Rhyo ^b	Rhyo-dac	Dacite	Dac-and	And-bas ^c	Bas-and	Dolerite	Basalt ^d	Arc-bas ^d	OIB2 ^e	P-MORB	Contl
A	-7.06	-6.00	-4.43	-4.68	-4.50	-4.81	-3.96	-4.09	-5.20	-3.89	-5.10	-2.54	-4.81
B	18159	11452	9029.8	8655.5	7967.9	6826.7	6077.6	5293.6	6597.5	5018.7	6230.3	4054.9	5685.1
C	141.1	222.0	486.2	500.1	482.4	582.2	567.2	607.5	599.3	616.7	554.7	638.3	601.2
No. Pts.	43	22	46	41	31	24	22	27	33	25	27	25	41
rmsd	0.02	0.06	0.03	0.04	0.02	0.04	0.03	0.06	0.07	0.04	0.11	0.09	0.08
T_{12} (K) ^f	1093.7	1080.6	1035.8	1019.2	965.3	988.3	948.0	936.5	982.9	932.5	919.0	917.2	939.4
m ^g	21.9	22.7	31.0	32.7	33.0	40.9	39.7	45.8	44.1	46.9	43.1	47.8	46.7

^a Whittington et al. (2009a)

^b Romine and Whittington (2015)

^c Sample SBA from Robert et al. (2013)

^d Fits to data on dry samples of basalt SB and arc basalt Fu06 from Robert (2014)

^e Sehlke et al. (2014)

^f The digits reported for T_{12} are needed to accurately ascertain fragility.

^g Fragility, $m = B/[T_{12}(1-C/T_{12})^2]$, with parameters B, C and T_{12} obtained from the VTF fit.

Table 5. Thermal diffusivity values (T in K; D in mm^2s^{-1})

rhyodacite		dacite		dacite-andesite		andesite ^a		and-basalt ^b		basalt-andesite		dolerite #3 ^c		dolerite#4 ^b	
T	D	T	D	T	D	T	D	T	D	T	D	T	D	T	D
294.1	0.592	295.4	0.535	293.5	0.569	294.9	0.567	294.4	0.549	294.2	0.536	294.6	0.544	294.5	0.512
349.6	0.540	344.2	0.514	337.7	0.557	347.4	0.555	345.8	0.528	338.1	0.513	348.9	0.529	346.7	0.484
404.4	0.517	405.9	0.500	388.8	0.546	411.1	0.536	410.9	0.513	390.2	0.496	403.2	0.512	410.7	0.476
475.5	0.504	477.7	0.485	448.0	0.535	475.3	0.524	485.2	0.499	444.5	0.486	479.9	0.495	475.0	0.463
553.7	0.491	557.8	0.480	515.6	0.522	555.0	0.516	554.2	0.484	518.2	0.474	554.5	0.487	555.7	0.456
644.7	0.472	653.0	0.479	598.8	0.516	646.7	0.506	643.5	0.478	598.4	0.469	645.9	0.484	645.9	0.451
749.7	0.479	753.3	0.470	689.5	0.509	748.9	0.500	750.5	0.466	693.4	0.461	745.1	0.482	748.4	0.448
855.3	0.470	860.9	0.470	794.0	0.502	859.6	0.495	859.6	0.460	783.1	0.460	872.1	0.476	862.5	0.445
974.6	0.466	998.0	0.466	902.1	0.497	960.3	0.490	970.8	0.438	882.9	0.456	1031	0.446	959.5	0.44
1126	0.443	1118	0.464	1027	0.490	1008	0.490	1069	0.395	981.3	0.447	1164	0.377	1062	0.345
1154	0.446	1177	0.472	1072	0.487	1041	0.487	1090	0.358	1010	0.446	1202	xtls.	1095	0.341
1184	0.406	1235	0.433	1146	0.440	1077	0.489	1115	0.356	1039	0.444			1125	0.345
1214	0.391	1293	0.427	1182	0.419	1082	0.484	1134	0.352	1068	0.426			1156	0.358
1244	0.387	1351	0.417	1219	0.426	1106	0.485	1176	0.362	1096	0.388			1181	xtls
1274	0.395	1405	0.429	1329	0.431	1131	0.480	1203		1152	0.383				
1324	0.380	1459	0.413	1439	0.436	1156	0.454			1209	0.455				
						1180	0.428								
						1232	0.418								
						1282	0.415								

Notes: Uncertainties are $\sim\pm 2\%$, except at room temperature where uncertainties are $\pm 3\text{-}5\%$ per standardization.

^a Results for sample DF-05 (andesite with 0.03 wt% F from Hellwig 2006) were identical within uncertainties and provided $D_{\text{melt}}=0.416 \text{ mm}^2\text{s}^{-1}$ from 1175 to 1428 K.

^b Data from andesite-basalt (SBA) before concentric cylinder measurements.

^c Dolerite #3 was the sample as received. Dolerite #4 was measured after remelting.

Table 5, continued.

Basalt (synthetic)		Arc basalt		OIB1		OIB2		MORB		P-MORB		Continental	
<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>	<i>T</i>	<i>D</i>
294.3	0.542	293.9	0.520	293.5	0.461	296.6	0.529	293.2	0.556	295.2	0.555	293.2	0.472
347.0	0.516	336.6	0.487	377.0	0.445	350.0	0.501	337.1	0.536	337.2	0.543	337.2	0.446
411.6	0.496	384.7	0.471	474.2	0.434	413.1	0.477	378.3	0.519	390.5	0.530	388.9	0.430
474.2	0.485	447.1	0.454	574.4	0.429	478.7	0.465	429.9	0.502	445.4	0.513	451.1	0.418
555.2	0.475	515.9	0.443	699.4	0.429	562.9	0.455	478.0	0.493	515.7	0.506	517.7	0.408
646.0	0.467	596.1	0.439	824.7	0.425	654.2	0.453	553.8	0.483	597.0	0.489	602.6	0.403
747.5	0.461	688.6	0.440	949.6	0.416	759.5	0.444	625.2	0.479	687.9	0.479	693.4	0.394
859.1	0.451	779.9	0.440	1074	0.404	869.4	0.449	725.2	0.474	789.0	0.474	793.6	0.389
957.7	0.442	879.5	0.439	1223	0.458	992.8	0.456	825.0	0.468	900.0	0.460	902.1	0.388
1062	0.426	991.2	0.442	1348	0.402	1096	0.364	901.3	0.464	1020	0.450	1030	0.346
1096	0.342	1051	0.386	1374	0.388	1095	0.402	949.7	0.455	1140	0.477	1073	0.390
1126	0.335	1077	0.370	1398	0.407	1095	0.399	974.5	0.449	1261	0.453	1113	0.407
1155	0.331	1104	0.398	1424	0.170	1147	0.494	997.8	0.405	1382	0.399	1145	0.480
1179	0.330	1161	0.428			1202	0.485	1023	0.309	1482	0.205	1186	0.480
1231	xtls.					1343	0.472	1048	0.289				
								1073	0.369				
								1097	0.376				
								1124	0.370				
								1470	0.310				

Table 6. Thermal diffusivity initial values and fitting parameters.

Sample	Thickness mm	T °C	D_{298} mm ² s ⁻¹	$D = FT^{-G} + HT$			T_{melt} K	D_{melt} mm ² s ⁻¹
				F	G	$H \times 10^5$		
Leucogranite ^a	0.77	24.3	0.67	5.1196±1.0	0.37323±0.036	17.06±1.8	1370	0.52
Rhyolite ^b	0.4-0.9	21.0	0.63	2.9587±0.6	0.28256±0.034	12.647±1.9	1330	0.52
Rhyo-dac (syn.)	0.70	21.9	0.59	4.4338±1.7	0.36867±0.067	10.776±3.1	1200	0.39
Dacite	0.389	22.1	0.535	2.2637±0.3	0.2655±0.024	10.482±1.3	1300	0.42
Dacite-andesite	0.38	20.2	0.57	1.7065±0.12	0.19834±0.013	5.8235±0.9	>1180	0.42 ^g
Andesite (syn.)	0.886	21.6	0.567	2.682±0.37	0.28213±0.025	11.05±1.5	>1175	0.42h
And-bas ^c (syn.)	0.875	21.0	0.54	1.5661±0.27	0.18983±0.031	2.9777±2.0	1150	0.348
Bas-and	0.527	20.9	0.57	2.9178±0.62	0.31099±0.039	11.241±2.1	1120	0.39
Dolerite#3d ^a	0.885	21.2	0.543	2.7491±0.56	0.29749±0.038	12.639±2.3	1100	<0.37
Dolerite#4	0.895	21.1	0.512	2.5599±0.61	0.29707±0.043	11.546±2.3	1100	0.34
Basalt (syn.)	0.976	21.0	0.542	2.5776±0.58	0.28395±0.041	8.3044±2.5	1100	0.33
Arc basalt	0.71	20.6	0.54	8.5283±2.3	0.51659±0.048	20.721±1.8	1080	0.36
OIB1	0.935	20.4	0.46	0.67838±1.0	0.0704±0.022		1150	~0.25
OIB2#1	1.045	22.8	0.53	7.5374±1.2	0.488±0.027	19.784±1.1	1070	<0.36 ⁱ
OIB2#2	1.088	22.8	0.49				1400	0.34
MORB#1 ^e	0.71	20.1	0.56	3.5728±0.8	0.34001±0.04	12.108±2.2	1050	0.29 ^j
MORB#2 ^f	0.663						1470	0.31 ^j
P-MORB	0.846	21.9	0.56	1.4342±0.4	0.16726±0.004		1500	>0.2
Continental	0.748	19.9	0.47	4.4707±1.0	0.41267±0.041	13.284±1.8	1030	<0.34 ⁱ

Notes: Syn. indicates synthetic, which lacked Fe³⁺ associated with many natural samples. Designations #1 and #2 indicate runs were performed on different sections of the same sample. Except as footnoted, values of D immediately upon melting are presented, because several samples crystallized shortly thereafter. The fits provide D in mm²s⁻¹ using T in K. Coefficient G has no units. Units of coefficient F depend on the value of G. Coefficient H has units of mm²s⁻¹K⁻¹. χ^2 ranged from 0.0001 to 0.0005. High uncertainties in coefficient F are due to tradeoffs in fitting exponent G. To reproduce the data, all digits that are shown for F, G, and H are needed. These fits can be projected up to the melting temperature less ~100 K.

^aData on D from Hofmeister et al. (2014)

^bData on D from Romine et al. (2012)

^cFit to and-bas data on chips both before and after concentric cylinder measurements. D_{melt} constrained by multiple runs.

^dDolerite #1 had color zonation and is not reported. Dolerite#3 is the sample as received, whereas dolerite#4 was remelted.

^eFrom glassy rim which is wet compared to the core, as indicated by its low melting temperature.

^fFrom the interior, which is dry as indicated by having the nominal melting temperature of basalts.

^gThe temperature dependence of D_{melt} for dacite-andesite was resolved as $0.3505 + 5.9955 \times 10^{-5}T$ in mm²s⁻¹ from 1180-1430 K.

^hConstant $D_{\text{melt}}=0.416$ was observed from 1180-1430 K for the andesite with 0.03 wt% F, which equals D of F-free andesite upon melting.

ⁱMelting may have been incomplete due to outgassing and crystallization.

^jThe temperature dependence of wet and dry MORBs combined gives $D_{\text{melt}} = 0.23681 + 4.9782 \times 10^{-5} T$ in mm^2s^{-1} from 1000-1470 K.

**TABLE 7 is vertical and is after the references in the
ms file**

Table 8. Fits to heat capacity in ($\text{J g}^{-1}\text{K}^{-1}$) against temperature in K.

	Leuco	Rhyolite	Rhyodac	Dacite	Dac-and	And-bas ^a	Bas-and ^a	Dolerite	Bas ^a	Arc bas ^a	MORB	P-MORB	OIB2	Contl.
Glass														
a	1.07	1.19	0.95	0.79	0.85	1.00	1.06	1.036	0.92	0.91	0.66	0.83	0.919	0.95
b	0.00027	0.00005	0.00025	0.00045	0.00051	0.00023	0.00012	0.00019	0.00032	0.0003	0.00066	0.00030	0.00037	0.00016
$c \times 10^{-4}$	-4.13	-4.63	-2.76	-0.915	-0.315	-2.80	-3.85	-3.66	-1.56	-2.45	-0.170	-0.168	-1.20	-1.35
$C_{p,g}$ at 298K	0.68539	0.68353	0.71370	0.82106	0.96651	0.75324	0.66222	0.68048	0.8397	0.732	0.83754	0.90048	0.89413	0.84566
T_g (K)	1080	1047	1044	1042	978	1013	968	947	1008	948	964	965	953	947
$C_{p,g}$ at T_g	1.33	1.203	1.183	1.253	1.353	1.23	1.14	1.18	1.22	1.17	1.29	1.12	1.26	1.09
Liquid														
d	1.49	1.31	1.17	1.38	1.54	1.90	1.71	1.49	1.98	1.67	1.81	1.39	1.77	1.63
e	0	-0.00003	0.00018	0.00004	0.00008	-0.00031	-0.00025	6.49×10^{-6}	-0.00034	-0.00012	-0.00014	0.00007	-0.00013	-0.00004
2σ	0.02	0.01	0.02	0.03	0.09	0.01	0.02	0.04	0.02	0.02	0.01	0.11	0.02	0.02
$C_{p,l}$ at T_g	1.49	1.34	1.36	1.42	1.62	1.58	1.30	1.50	1.63	1.55	1.68	1.46	1.65	1.59
$C_{p,conf}$	0.162	0.138	0.178	0.167	0.271	0.375	0.17	0.321	0.41	0.38	0.384	0.340	0.391	0.499
$C_{p,l}$ at T_{liquid}	1.49	1.28	1.39	1.43	1.64	1.46	1.35	1.50	1.48	1.49	1.60	1.49	1.58	1.57
T_{max}	1400	1550	1550	1750	1750	1750	1750	1750	1750	1650	1100	1750	1750	~1600

Notes: All fits are to data collected with the Netzsch calorimeter. For Maier-Kelly (1932) glass fits, $C_p = a + bT + cT^{-2}$ with $2\sigma < 0.01$ and $T < T_g$. For linear fits to liquid data, $C_p = d + eT$ with 2σ and T_{max} as reported. For T_{liquid} see Table 2. Experimental accuracies are 1% for glass and 2% for melt based on repeated measurements on $\alpha\text{-Al}_2\text{O}_3$.

^aRobert et al. (2014) provide data on and-bas (labeled SBA therein); bas-and (Fu18); bas (SB); and arc basalt (Fu06).

Table 9. Fits to glass thermal conductivity = $A+BT+CT^{-2}+DT^{1/2}$ and selected liquid k (in $\text{W m}^{-1}\text{K}^{-1}$) against temperature (in K).

	Leuco	Rhyolite	Rhyodac	Dacite	Dac-and	And-bas	Bas-and	Dolerite	Basalt	Arc bas	MORB	P-MORB	OIB2	Contl
Glass														
A	2.8683	2.7578	2.3706	1.7444	1.7973	1.9808	2.2837	2.2931	2.2099	2.5728	1.8148	1.5682	2.3872	1.9863
$B \times 10^4$	11.609	12.292	14.002	13.104	10.577	6.6136	10.735	12.6983	11.064	22.7959	17.759	4.046	21.658	12.735
C	-64532	-71174	-40254	-12497	-2953.7	-40598	-54011	-52463	-21139	-34545	1768.9	3189.7	8296.5	7263.5
$D \times 10^2$	-8.7197	-7.6603	-7.7977	-4.926	-3.4101	-3.8249	-6.1831	-6.607	-5.3125	-10.661	-6.0827	-1.8849	-9.273	-6.4536
k at 298K	0.98	1.00	0.99	1.14	1.41	1.06	0.93	0.94	1.21	1.02	1.28	1.37	1.33	1.13
k at T_g	1.20	1.50	1.28	1.51	1.76	1.39	1.34	1.41	1.44	1.41	1.63	1.38	1.50	1.20
Liquid														
k_{liquid} at T_g	1.81	1.55	1.27	1.48	1.64	1.40	1.31	1.36	1.41	1.52	1.29	1.46	1.52	1.3-1.5
k_{liquid} at T_{liquid}	1.80	1.55	1.30	1.49	1.73	1.29	1.36	1.36	1.28	1.44	1.34	1.22	1.46	~1.4

Notes: Data on glass thermal conductivity are accurate to better than 4%, whereas those for the liquid range are ~5% and increasing to ~8% at the liquidus, see text. For T_{liquid} , see Table 2. Parameters in fits are not rounded to ensure reproduction of the calculations. For the dacite-andesite and MORB liquids, $D(T)$ is known, providing higher accuracy; for the other samples D is assumed to be constant. Assuming $\alpha = 7.5 \times 10^{-6} \text{ K}^{-1}$ (Bouhifd et al. 2015) provides $k_{\text{liquid}} = 1.3298 + 0.00031647T$ for dacite-andesite and $k_{\text{liquid}} = 1.2005 + 9.2508 \times 10^{-5}T$ for MORB.

Figure 1a

Figure 1b

Figure 2

Figure 3a

Figure 3b

Figure 3c

Figure 3d

Figure 4a

Figure 4b

Figure 4c

Figure 5a

Figure 5b

Figure 5c

Figure 6a

Figure 6b

Figure 6c

Figure 7

Figure 8a

Figure 8b

Figure 8c

Figure 9