



Viscosity and structure of hydrous silicate liquids: Constraints on oceanic hydrothermal circulation

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ABSTRACT

Despite its central role in cooling the oceanic lithosphere, the efficiency and geometry of hydrothermal circulation remain debated. While ridge-scale simulations typically adopt pure-water transport properties, circulating liquids at the magmatic-hydrothermal transition can dissolve appreciable silicate at elevated temperatures and pressures. These silicate-bearing liquids likely mediate heat and mass transfer, yet their mobility is poorly constrained. Here we present viscosity measurements of hydrous $\text{Na}_2\text{Si}_2\text{O}_5$ liquids (HNS2) at 130–250 °C using in-situ falling-sphere viscometry in a hydrothermal diamond-anvil cell and characterize liquid structure with Raman spectroscopy up to 350 °C and 440 MPa. Measured HNS2 viscosities increase by $\sim$ one order of magnitude (7.35×10^{-4} to 9.60×10^{-3} Pa·s) as $\text{Na}_2\text{Si}_2\text{O}_5$ increases from 25.4 to 69.7 wt.%. Raman spectra indicate systematic shifts toward less connected Si-O environments with increasing H_2O and temperature, consistent with the lower viscosity in H_2O -rich liquids. Incorporating these data into two-dimensional porous-flow simulations of hydrothermal circulation shows that compared to pure water, introducing modest dissolved silicate weakens distal off-axis circulation while preserving a ridge-proximal primary cell that reaches the brittle-ductile transition depth. This suggests that pure-water models are likely to overestimate the lateral extent of off-axis advection required to match conductive backgrounds. More generally, silicate-bearing liquids can remain sufficiently mobile to contribute to lower-crustal heat extraction and may help account for pervasive low-grade (greenschist-facies) hydrothermal alteration in gabbroic and sheeted-dike sections. Silicate contents in deep liquids thus should be incorporated into quantitative models of crustal cooling.

1. Introduction

Seafloor hydrothermal systems serve as critical conduits for exchanging energy and materials among the oceanic crust, the underlying mantle, and the ocean. In a general view of black-smoker systems, cold seawater penetrates the crust, is heated as it approaches the magma body, reacts with surrounding rocks and ascends as buoyant fluid that precipitates hydrothermal minerals upon mixing and cooling (Rona, 1984). However, the connection of this process to cooling the oceanic crust and lower crust in particular remains debated. Hydrothermal convection and conduction in oceanic crust are estimated to account for $\sim$ 70% of Earth's surface heat loss (Parsons and Sclater, 1977; Sclater et al., 1980), yet conductive heat flow observed at ridge axes ($135 \pm 100 \text{ mW}\cdot\text{m}^{-2}$) is markedly lower than that predicted by plate-cooling models

($>1000 \text{ mW}\cdot\text{m}^{-2}$) (Fisher and Harris, 2010). This “missing heat” discrepancy has motivated vigorous near-ridge circulation models (e.g., Barreyre et al., 2025; Harris et al., 2017; Hasenclever et al., 2014; Hutnak et al., 2008; Kardell et al., 2021; MacLennan et al., 2005; Price et al., 2023). For instance, MacLennan et al. (2005) propose that substantial lower-crustal cooling can be achieved by the existence of hydrothermal fluids within 20 km of the ridge, removing $\sim$ 50–70 kW of heat per meter length of ridge axis at crustal ages younger than 0.1–0.4 Ma. In contrast, fluid-rock Sr isotope constraints imply a time-integrated fluid flux of $1.5\text{--}3.2 \times 10^6 \text{ kg}\cdot\text{m}^{-2}$ (Harris et al., 2017), suggesting that upper crustal hydrothermal circulation alone may account for only 20–60% of the heat potentially available from lower-crustal cooling. These uncertainties underpin competing views of crustal accretion and thermal structure (Früh-Green et al., 2022), e.g., the “gabbro glacier”

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model (Morgan and Chen, 1993) versus “multiple sills” model (Boudier et al., 1996; Kelemen et al., 1997), and highlight the need for improved constraints on the properties of deep circulating liquids near the magmatic-hydrothermal transition.

Heat-flow studies have traditionally emphasized black-smoker systems because they are the most visible hydrothermal circulation at middle ocean ridges (MORs) and provide direct constraints on discharge temperatures and fluxes. Indeed, Earthquake cracking events on axis are directly tied to black smoker system changes (Sohn et al., 1998), indicating a causal relationship of fresh seawater directly reacting with hot rock to produce black smoker output. Geophysical constraints further imply efficient lower-crustal cooling (Dunn et al., 2000), motivating models in which deep hydrothermal circulation extracts heat near the magmatic-hydrothermal transition (Cherkaoui et al., 2003; Hasenclever et al., 2014). In particular, Hasenclever et al. (2014) propose that deep off-axis circulation can approach the magmatic-hydrothermal transition and account for up to ~70% of the thermal energy released at the ridge axis. If deep circulation operates in this regime, the common assumption in the existing models (e.g., Barreyre et al., 2025; Fisher and Becker, 2000; Hasenclever et al., 2014; Kardell et al., 2021) using pure-water (or seawater) transport properties becomes a potentially important limitation. Silicate solubility in aqueous fluids increases strongly with temperature (T) and pressure (P) (Manning, 2018), and dissolved silicate can increase liquid viscosity and modify density relative to pure water (Audéat and Keppler, 2004). Because viscosity and density control buoyancy forcing and Darcy velocities, ignoring dissolved silicate may bias modeled circulation geometry and the heat-flow patterns.

Geological observations and experiments support the plausibility of silicate-bearing liquids in oceanic crust. Most of the oceanic crust section of the Troodos ophiolite has been greenschist metamorphosed with abundant chlorite, actinolite, epidote, albite and quartz interspersed to various degrees even in the gabbros and sometimes ultramafic rocks (e.g., Coogan et al., 2006). This verifies that significant amounts of water do get to Moho depths but also tends to indicate it resides in a liquid that equilibrates with this mineral assemblage. Experimental constraints in simplified Na-Al-Si-O-H systems further show that quartz-albite-saturated liquids can evolve continuously with cooling, from rhyolite-like compositions at ~800 °C to Na₂Si₂O₅-rich compositions carrying up to 37 wt.% H₂O at 330 °C (Lundstrom, 2020). While much work remains to understand the phase equilibria of more complex systems, experiments show that NS2-H₂O compositions are completely miscible (a single liquid forms) below 400 °C at 100 MPa (Mustart, 1972). These compositions are not well represented as “water with dilute solutes”; rather, they span a continuum of silicate/H₂O ratios where changes in molecular structure can plausibly translate into large changes in mobility. Accordingly, we refer to NS2-H₂O composition as ‘liquids’ rather than ‘fluids’ throughout this paper.

A key missing constraint is the viscosity of silicate-bearing liquids at sub-magmatic conditions, which governs liquid’s mobility relevant to the magmatic-hydrothermal transition. In silicate systems, water acts as a depolymerizing agent by breaking bridging oxygen bonds and forming terminal hydroxyl groups bonded to tetrahedrally coordinated cations such as Si and Al. As a result, even minor additions of water can reduce silicate melt viscosity by several orders of magnitude (e.g., Dingwell et al., 1996; Giordano et al., 2004; Whittington et al., 2000). Most existing viscosity data focus on the silicate melts endmember at magmatic T (> 600 °C) and comparatively low water contents (< 10 wt. %) (e.g., Giordano et al., 2008; Le Losq et al., 2025). One previous work by Audéat and Keppler (2004) measured viscosities of aqueous liquids containing 10–80 wt.% dissolved silicates and found values ranging from 10⁻⁴ to 10⁻⁵ Pa·s at 800 °C. However, their findings do not directly constrain relatively low-T oceanic circulation and require caution when extrapolated to sub-magmatic conditions where potential melt-fluid immiscibility may occur in the investigated systems such as albite-H₂O and leucite-H₂O. Furthermore, because viscosity is tightly linked to liquid structure, namely degree of polymerization (Mysen et al., 1980;

Wang et al., 2014) and abundance of network modifiers such as Na₂O (Cassetta et al., 2023; Mysen and Cody, 2005; Stopponi et al., 2025), coupled constraints on both viscosity and structure are needed to evaluate mobility in this low-temperature regime.

Here we present measurements of viscosity and molecular structure for hydrous sodium disilicate liquids (HNS2) with varying silicate/H₂O ratios under temperature and pressure conditions up to 350 °C and 440 MPa, by using a combination of in-situ falling-sphere viscometry and Raman spectroscopy in a hydrothermal diamond anvil cell (HDAC). By focusing on this simplified system, we isolate the first-order effect of dissolved silicates on liquid viscosity and structure, providing a baseline for evaluating more complex natural compositions. Motivated by proposals that silicate-bearing hydrothermal liquids may participate in deep crustal circulation and potentially influence near-axis thermal structure (Lundstrom, 2020), we further examined the HNS2 mobility and established a two-dimensional convection model to simulate potential convection patterns in young oceanic crust within a porous-flow framework.

2. Methods

2.1. High pressure and high temperature liquids

The starting sodium disilicate glass (Na₂Si₂O₅, NS2) was synthesized from reagent-grade Na₂CO₃ and SiO₂ powder. The mixture was dry-ground, decarbonated by slowly heating, and molten in a platinum crucible at 1100 °C for 1 h. The glass was re-ground and re-melted twice for homogeneity and stored dry. Prior to each experiment, a piece of NS2 glass was doubly polished to a 120–150 µm plate, then rinsed in acetone. This preparation minimized potential surface alteration and enabled volume-based mass estimates of bulk silicate/H₂O ratio for each run (Audéat and Keppler, 2004; Steele-MacInnis and Schmidt, 2014).

Six experiments were performed in an externally heated Bassett-type HDAC-VT (Bassett et al., 1993). The experimental conditions are summarized in Table 1 with detailed composition given in Table S1. To minimize fluorescence background in Raman spectra, two type IIa ultra-low fluorescence diamonds (1.6 mm culets) were employed. Rhenium gaskets were used as sample chambers with an inner diameter of ~1.0 mm. The gaskets were pre-indented to a final thickness of 227–239 µm from an initial thickness of ~251 µm. Temperature was measured with two K-type thermocouples attached to both the upper and lower diamond anvils. Pressure was primarily generated by heating in the confined sample chamber, as the sample followed an approximately isochoric path defined by its bulk density (Shen et al., 1993). Each run contained an NS2 plate, 2–3 Ti spheres (10–24 µm diameter),

Table 1
Experimental conditions.

Run	Starting materials	wt. % solids	T* (°C)	P* (MPa)	Measurements
Cell1	w, gl, Ti	69.7 (9.1)	250	131 ^a	viscosity
Cell2	w, gl, Ti	27.2 (4.1)	300	380 ^a	viscosity + Raman
Cell3	w, gl, zr	41.2 (5.7)	280	270 ^b	Raman
Cell4	w, gl, zr	41.2 (5.7)	280	440 ^b	Raman
Cell5	w, gl, Ti	25.4 (3.5)	230	152 ^a	viscosity
Cell6	w, gl	79.8 (10.4)	350	140 ^a	Raman

* represents maximum experimental P-T conditions.

^a : pressure calculated from EOS.

^b : pressure calculated based on the wavenumber shift of the ν³(SiO₄) Raman band of zircon. The values in parentheses are standard deviations. w: water; gl: NS2 glass; Ti: titanium sphere; zr: zircon.

and a drop of distilled H₂O (Fig. S1). A small vapor bubble was intentionally included to constrain the starting bulk density. In two runs (Cell3 and Cell4), a zircon chip was added as a pressure sensor (Schmidt et al., 2013).

To determine the silicate/H₂O ratio in the sample chamber, the three dimensions of the glass chip, diameter of Ti spheres and sample chamber were measured using a high-resolution Leica stereomicroscope, combined with the room-condition density of NS2 glass (2.50 g/cm³, Pant and Cruickshank, 1968). The uncertainty in each length measurement was typically within $\pm 5 \mu\text{m}$ for the glass and $\pm 1 \mu\text{m}$ for the spheres. The thickness of the gasket was measured by a Mitutoyo digital micrometer (uncertainty is $\pm 0.5 \mu\text{m}$). The chamber dimensions remained unchanged within uncertainties before and after the experiments, because of the low-pressure conditions. Water volume was determined by subtracting the volumes of the glass and titanium spheres from the measured volume of the sample chamber. Water density during heating was estimated from the vapor-bubble dissolution temperature using the IAPWS-95 equation of state for pure water (Wagner and Pruß, 2002). Typical errors associated with the silicate/water weight ratio determination were $\sim 10\%$.

2.2. Falling sphere experiments

In-situ falling-sphere viscometry in DAC has a long history (e.g., Audetat and Keppler, 2004; Munro et al., 1979; Piermarini et al., 1978). The HDAC was initially oriented with the compression axis vertical, while the sample was heated to the target temperature and held for 10 min. After observing the complete dissolution of the glass, the cell was tilted by 90°, making the compression axis horizontal. Thus, spheres fell in the direction parallel to the chamber diameter with a total length (L) of $\sim 1 \text{ mm}$, through a layer of liquid whose thickness (H) was equal to the thickness of the gasket. The spheres fell within several seconds due to gravity (Supplementary Video1), recorded with a high-speed camera (124 frames/s). Because the chamber was fully liquid-filled, spheres remained separated from the anvils by a liquid layer. That the spheres were not rolling along the diamond surface can be established by the

following evidence: (1) All falling paths were vertical with lateral drift below the 1-pixel resolution (Fig. 1); (2) in a single experiment, after the compression axis of the HDAC was rotated 180° (up-side-down), sphere terminal velocities were indistinguishable from those determined before the rotation.

Sphere locations over time were tracked using MTrackJ in ImageJ (Meijering et al., 2012), to generate a distance/time curve, from which the terminal velocity of the sphere was determined. The long path length L relative to the small sphere diameter d (ca. 0.02 mm) essentially ensured that terminal velocity was reached in all our measurements. Two examples of the distance versus time curves are shown in Fig. 1. The terminal velocities were then used to calculate the viscosity using Stokes' law:

$$\eta_R = \frac{gd^2 \Delta\rho}{18U_\infty} \quad (1)$$

where η_R is raw viscosity, d is diameter of the sphere, g is gravitational acceleration, $\Delta\rho$ is density difference between the sphere and the liquid, and U_∞ is terminal velocity of the sphere. HNS2 density was estimated from the measured silicate/H₂O ratio assuming ideal mixing with zero excess volume, as a first-order approximation for this system. The density of Ti sphere was determined using the high T-P Birch-Murnaghan thermal equation of state (Zhang et al., 2008). Due to the assumptions used in estimating HNS2 density, we assign $\Delta\rho$ an additional uncertainty of $\pm 5\%$.

The raw viscosity η_R needs to be corrected for the wall effect (W) (Faxen, 1922):

$$\eta_w = \eta_R \cdot W \quad (2)$$

where

$$W = 1 - 2.104 \cdot (d/H) + 2.09 \cdot (d/H)^3 - 0.95 \cdot (d/H)^5 \quad (3)$$

where H is thickness of the gasket.

There is also a correction for drag due to the end of the sample

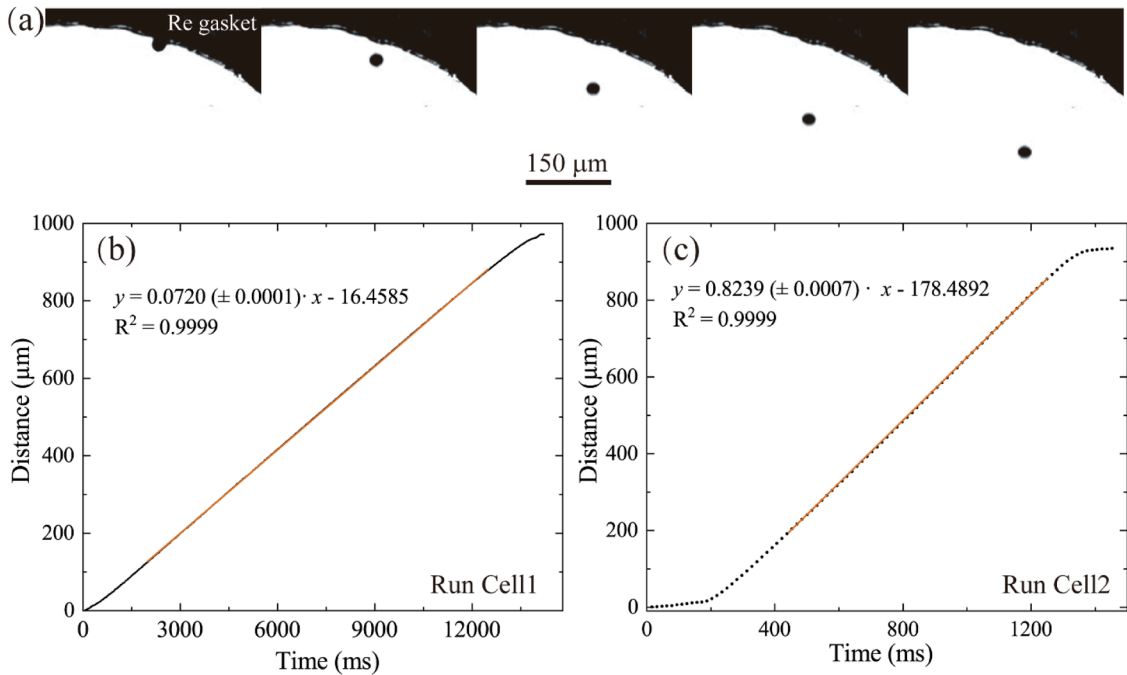


Fig. 1. (a) A sequence of images illustrating the descent of Ti spheres in HNS2 at 150 °C for Run Cell2 (27.2 wt. % NS2). The time interval between each snapshot is 125 ms. (b) and (c) show the falling distance versus time for Ti spheres in Run Cell1 (69.7 wt.% NS2) at 210 °C and Run Cell2 at 150 °C, respectively. The initial acceleration and the landing of the spheres are visible before and after reaching terminal velocity. The slopes of the plotted red lines have been used to determine the terminal velocities of the spheres through linear regression of the falling distance versus time data.

chamber, known as the end effect (E), given by (Ladenburg, 1907):

$$\eta_E = \eta_R \cdot E = \frac{\eta_R}{1 + 3.3 \frac{d}{2L}} \quad (4)$$

where L is the total height of the capsule.

In our experiments, H is 227–239 μm and d is 12–24 μm (Table 2). The d/H is typically < 0.1 , leading to a small uncertainty in the correction for viscosities. Measurement reproducibility was evaluated by using different sphere sizes at the same temperature in selected runs (e.g., $\sim 200^\circ\text{C}$ in run Cell1 and Cell2, Table 2). An evaluation of uncertainty in viscosity determination is given in Section 3.1, and the sensitivity to various combined confinement correction schemes is evaluated in Supplementary Note S1 and Fig. S4.

2.3. In-situ Raman spectroscopy

Raman spectra were collected on the GSECARS Raman System using a 532 nm laser. A Princeton Instruments Acton Series 2500 spectrograph (focal length: 500mm, gratings: 1200 lines/mm) was coupled with a UV-enhanced PIXIS100 (200–1000nm) detector. A $20 \times$ objective and 50 μm confocal pinhole yielded a 4.5 μm spot size on the sample. Spectra of glass or liquids (Fig. 3) were collected with a laser power of 25 mW (measured after exiting the microscope objective) for silicate-rich experiments (e.g., run Cell6) and 40 mW for H_2O -rich experiments (e.g., run Cell2), with 3 accumulations of 420–900 s in the wavenumber range of 200–1200 cm^{-1} and of 90–300 s in the wavenumber range of 2700–4000 cm^{-1} .

Spectra were normalized following Steele-MacInnis and Schmidt (2014), with linear baseline subtraction using Fitky software. Deconvolution and symmetric peak shapes (Gaussian + Lorentzian area model) fitting were established in PeakFit v4.12. For experiments with zircon pressure calibration (Table 1), the antisymmetric stretching mode of zircon at 1008 cm^{-1} was monitored during both heating and cooling processes and corrected for systematic offsets by referencing the ambient-position shift of the $\nu^3(\text{SiO}_4)$ band, following Schmidt et al. (2013).

2.4. Two-dimensional hydrothermal flow model

We employed a two-dimensional (2D) hydrothermal convection model to evaluate circulation of silicate-bearing liquids based on a finite difference scheme (Supplementary Note S2). The model is based on the open MATLAB code from Kardell et al. (2021) and configured for the extensively investigated fast-spreading East Pacific Rise at 9°N in this study. The code was run in hydrothermal mode, with thermal structure (Fig. S7) adopted from Dunn et al. (2000), to isolate the hydraulic influence of liquid properties on circulation geometry. Instead of assuming pure water, we prescribed transport properties for silicate-bearing

liquids using the new viscosity constraints from this study. The computational domain is 25 km wide and 6 km deep, with a uniform grid of 37.5 m in both directions. The brittle-ductile transition was assumed to be at the 800 $^\circ\text{C}$ isotherm (Hasenclever et al., 2014), so that no significant liquid flow occurs through rocks hotter than 800 $^\circ\text{C}$. The permeability structure was adopted from Hasenclever et al. (2014). Porosity was held constant at 2%. Liquids were permitted to enter or exit the domain through the top boundary. Viscosity beyond present experimental temperature range was obtained by extrapolating the fitted viscosity-composition relation. Over the pressure range spanned by our experiments, the pressure effect is small relative to the order-of-magnitude viscosity variations driven by composition across the dataset. Accordingly, we treat composition as the primary control for the extrapolated viscosities (Fig. S5). The associated uncertainty was evaluated with sensitivity tests, which showed that the relative circulation differences among cases are robust (Supplementary Note S2). We ran the models for 20 kyr with 4 yr time steps. A quasi-steady state convergence was typically reached after approximately 16 kyr (Fig. S8).

3. Results

3.1. Viscosity measurements

Example images of the falling sphere in experiment Cell2 (150 $^\circ\text{C}$ and 85 MPa) are shown in Fig. 1a. The displacement of the sphere in each snapshot represents its falling distance. Three distinct trends were observed during each experiment (Fig. 1b and c): (1) An acceleration stage, where the falling distance gradually increases over time; (2) a linear falling stage, where the falling distance increases proportionally with time, indicating that the sphere has reached terminal velocity as the buoyancy force and viscous drag force are balanced; (3) a sharp inflection stage, as the spheres approach the bottom of the gasket during landing. The long segment of the linear falling distance indicates that we can easily obtain a robust terminal velocity. The influence of the channel ends (end effect) is minor ($< 1\%$) in our configuration with d/L of 0.01–0.02 (Fig. S4b).

Table 2 summarizes the experimental results. The HNS2 viscosities as a function of temperature are plotted in Fig. 2. The raw Stokes viscosity (η_R) is reduced by 15–20% after applying wall corrections, yielding corrected viscosities η_W in Table 2. The uncertainty in optical measurements for the sphere diameter ($\pm 1 \mu\text{m}$) introduces an additional 7–12% uncertainty in η_W . Along with the uncertainty in the calculated density mentioned above, the overall uncertainty in η_W remains within 18%. An additional contribution to the uncertainty arises from the choice of confinement correction scheme (finite cylinder and wall formulations), which brackets the viscosity within 7% over our d/H range (Note S1; Fig. S4); including this term by root-sum-square combination increases the total uncertainty to $< 19\%$ ($< 25\%$ by linear addition). This uncertainty is comparable to, or lower than, previously reported DAC

Table 2
Experimental data and calculated viscosities of HNS2.

Run	T ($^\circ\text{C}$)	P (MPa)	d_{sphere} (μm)	$\Delta\rho$ (g/cm^3)	velocity ($\mu\text{m}/\text{ms}$)	H_{channel} (μm)	η_R (Pa·s)	η_W (Pa·s)
Cell1	200	46*	23.1	2.87	0.069	239	1.20×10^{-2}	9.60×10^{-3}
	210	63*	24.7	2.87	0.086	239	1.11×10^{-2}	8.77×10^{-3}
	210	63*	23.1	2.87	0.072	239	1.16×10^{-2}	9.24×10^{-3}
Cell2	130	49	22.4	3.35	0.600	239	1.53×10^{-3}	1.23×10^{-3}
	150	85	24.5	3.35	0.824	239	1.38×10^{-3}	1.05×10^{-3}
	200	179	24.5	3.35	0.721	239	1.52×10^{-3}	1.19×10^{-3}
	200	179	22.4	3.35	0.586	239	1.56×10^{-3}	1.25×10^{-3}
	250	278	10.8	3.35	0.164	239	1.32×10^{-3}	1.19×10^{-3}
Cell5	150	12	17.4	3.40	0.638	227	8.76×10^{-4}	7.36×10^{-4}
	210	116	12.0	3.40	0.323	227	8.27×10^{-4}	7.35×10^{-4}

Pressures were calculated from EoS of water.

* represents minimum P values for high NS2/ H_2O ratio experiments, true pressure might be 100 MPa higher (cf., Section 3.3). d_{sphere} is the diameter of the sphere, $\Delta\rho$ is the density difference between the sphere and the liquid, η_R is the raw viscosity before correction, η_W is the viscosity after wall correction.

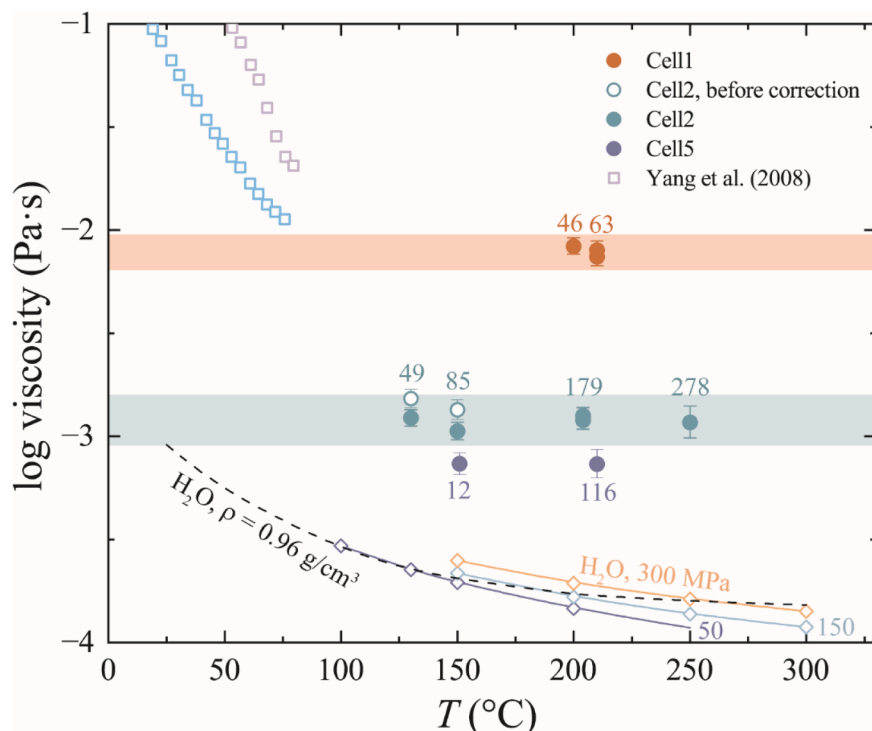


Fig. 2. Viscosities of HNS2 (hollow circle: no correction for wall effect; solid circle: after correction) plotted as a function of temperature. Orange, green and violet circles represent 69.7, 27.2, and 25.4 wt.% NS2 in the liquids, respectively. Numbers next to the symbols are pressures in MPa. Also shown as square are viscosities of hydrous sodium silicate (molar ratio of SiO_2 to Na_2O is 3.3, NS3.3) at 1 atm and up to 80 °C (Yang et al., 2008), pink and blue squares represent 34.2% and 17.1% NS3.3, respectively. The calculated viscosities of pure H_2O are based on the equations from the IAPWS Formulation 2008 for the Viscosity of Ordinary Water Substance (Huber et al., 2009). Dashed line: viscosity change of pure H_2O with constant density of 0.96 g/cm^3 ; diamond and solid line: viscosity change of pure H_2O with three different sets of constant pressure, 50 MPa, 150 MPa, and 300 MPa.

falling-sphere uncertainties (e.g., up to 50% in Piermarini et al., 1978 and 20–30% in Audétat and Keppler, 2004), underlining the importance of small d/H ratio adopted in falling sphere experiments. The viscosities obtained from spheres of different sizes in Cell1 and Cell2 show good agreement (Fig. 2). The viscosity of the silicate-rich system is systematically 7–13 times higher than that of the H_2O -rich system. For comparison, previous viscosity measurements by Yang et al. (2008) on hydrous sodium silicate (NS3.3), as well as calculated viscosities of pure H_2O based on the IAPWS Formulation 2008 (Huber et al., 2009), are also plotted.

3.2. Raman spectroscopy

The Raman spectra (Fig. 3) display broad and overlapping bands in the $250\text{--}1200 \text{ cm}^{-1}$ wavenumber range, attributable to silicate species (Mysen, 2009; Mysen and Cody, 2005). Peaks between low frequency (LF) region (450 and 600 cm^{-1}), which are observed in all spectra, correspond to the bending vibrations of bridging oxygen in Si-O-Si groups (Hunt et al., 2011; Le Losq et al., 2022; Steele-MacInnis and Schmidt, 2014). However, precise assignment of these peaks is challenging, as their positions are highly sensitive to variations of the Si-O-Si bond angle (Furukawa et al., 1981). The high-frequency (HF) region ($\sim 700\text{--}1200 \text{ cm}^{-1}$) is commonly discussed in relation to the average number of bridging oxygen (BO) atoms, or Q^n ($n = 0$ to 4) species (Brawer and White, 1975; Mysen and Cody, 2005). The evolution of Q^n species provides qualitative indications for melt polymerization and corresponding viscosity behavior. Thus, this study focuses on the HF spectral range.

The spectra from run Cell6 indicates a transition from NS2 glass signal below 100 °C to homogeneous HNS2 signal by 150 °C (Fig. 3a), consistent with optical observations of complete glass dissolution by 120 °C (Fig. S1). Raman spectra collected across the sample chamber display

identical features, suggesting that the system is homogeneous after full glass dissolution. During the glass dissolution, the peak near 950 cm^{-1} diminishes, while a weak peak emerges at around 850 cm^{-1} , and the peak at 1100 cm^{-1} shifts to 1050 cm^{-1} (Fig. 3a). The signal for the Si-O region is markedly stronger from the silicate-rich sample (Fig. 3a) than that of the H_2O -rich sample (Fig. 3b). A noticeable increase in the peak intensity at $\sim 770 \text{ cm}^{-1}$ is observed for both runs with increasing temperature, accompanied by a decrease in intensity at 1050 cm^{-1} . Remarkably, the 770 cm^{-1} band is dominant in the aqueous liquids of the H_2O -rich sample.

In the $2700\text{--}4000 \text{ cm}^{-1}$ region, a broad asymmetric Raman intensity peak appears (centered $\sim 3450 \text{ cm}^{-1}$) at room temperature with an extended tail toward lower frequencies (Fig. 3). This band is assigned to the O-H stretching modes of hydroxyl groups and molecular H_2O (Steele-MacInnis and Schmidt, 2014). As T increases, this Raman band narrows, becomes more symmetric, and adopts a Lorentzian shape. Additionally, the maximum intensity shifts gradually toward higher frequencies.

3.3. Pressure calibration

It has been pointed out that in the $\text{H}_2\text{O}\text{--}\text{Na}_2\text{Si}_3\text{O}_7$ system, pressures derived from the EoS of H_2O tend to be lower than the calibrated P by zircon at relatively high bulk densities, whereas both methods yield consistent results up to 550 °C in the H_2O -albite system (Schmidt et al., 2013). To assess the accuracy of P calculations using the EoS of H_2O in HNS2, we conducted two experiments using zircon as a P calibrant with different liquid densities (Fig. 4). Comparing pressures determined from zircon and H_2O , we conclude that application of the EoS of H_2O (Wagner and Pruß, 2002) provides reliable pressure estimates under our experimental conditions up to 300 °C , with $\sim 41.2 \text{ wt.}\%$ NS2 dissolved in the liquids. However, at very high sodium silicate concentrations (Run Cell1

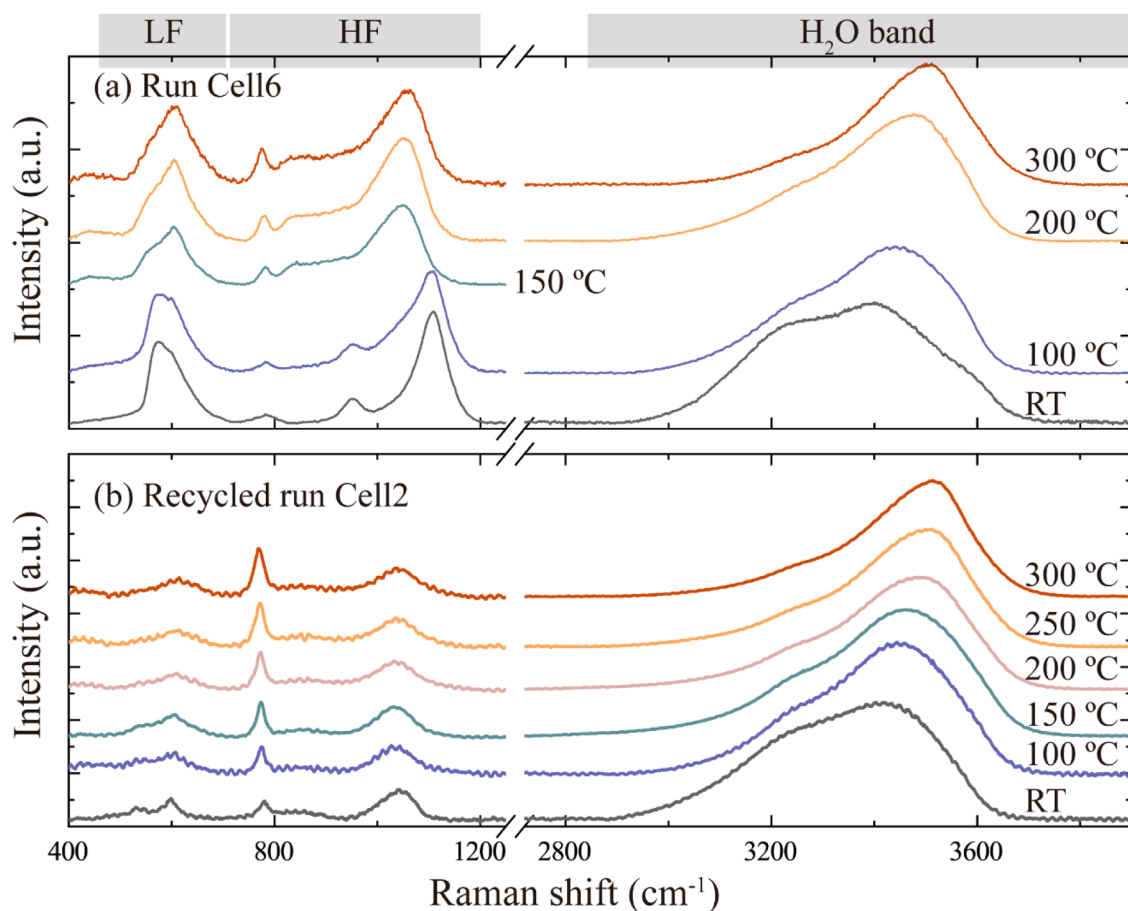


Fig. 3. In-situ Raman spectra of H₂O—NS2 liquids upon heating. (a) Run Cell6 with 79.8 wt.% NS2; (b) Run Cell2 with 27.2 wt.% NS2. Note that (a) was conducted with a piece of NS2 glass and H₂O at room temperature, where the glass dissolved after 120 °C. In contrast, (b) was a recycled run following viscosity measurements, with the initial composition consisting of a single HNS2 and an air bubble. All spectra are vertically offset for clarity.

and 6, 69.7 and 79.8 wt.% NS2, respectively), isochores may deviate toward lower pressures compared to those predicted by the EoS of H₂O. Consequently, the actual P could be 100 MPa higher than the calculated values in these experiments (Schmidt et al., 2013).

4. Discussion

4.1. Raman spectral evolution in hydrous sodium silicate liquids

Raman spectroscopy was used to characterize structure evolution in HNS2 as a function of temperature and bulk composition. Raman spectra were fitted with five main Gaussian-Lorentzian components in the Si-O stretching region and three components in the O—H stretching region (Fig. 5) to enable consistent comparison among experiments. Raman band intensities depend on overlapping vibrational contributions and Q^n -dependent cross sections, with depolymerized units (low n) often producing disproportionately strong signals (Furukawa et al., 1981; Le Losq et al., 2015a). Therefore, the fitted band areas are interpreted comparatively rather than as absolute measures of structural abundance.

NS2 glass at room temperature exhibits distinct features near 950 and 1100 cm^{-1} (Fig. 3a and S2a), consistent with previously reported vibrational environments in alkali silicate glasses (e.g., Brawer and White, 1975; Furukawa et al., 1981; Le Losq et al., 2015a). Upon heating and dissolution, the Raman features evolve into broader and partially overlapping spectral envelopes (Fig. 3). In particular, the spectra show a reproducible redistribution of intensity from higher-Raman-shift contributions (~ 950 – 1150 cm^{-1}) toward lower-Raman-shift contributions

($<950 \text{ cm}^{-1}$) with increasing temperature in both silicate-rich and H₂O-rich liquids (Fig. 3 and 5).

The feature near 770 cm^{-1} has been reported in hydrous silicate systems and commonly attributed to Si-O stretching vibrations in highly hydroxylated/molecular silicate species (Zotov and Keppler, 2002). Its presence in NS2 glass spectra at room T (Fig. 3a and S2a) may reflect near-surface hydration and/or glass-water reaction at ambient conditions. Spectral evolution in the 850–950 cm^{-1} interval is characterized by attenuation of the $\sim 950 \text{ cm}^{-1}$ feature and development of a shoulder toward $\sim 850 \text{ cm}^{-1}$ (Fig. 3). In silicate glasses, ~ 850 and $\sim 950 \text{ cm}^{-1}$ band are commonly attributed to Q^1 and Q^2 units (Brawer and White, 1975; Furukawa et al., 1981). However, in sodium silicate melts these features can redistribute and shift to lower-Raman-shift with increasing temperature (Mysen and Frantz, 1994), and in hydrous compositions the 860–900 cm^{-1} region can additionally include contributions from T-OH ($T = \text{Si, Al}$) vibrations (Le Losq et al., 2015a, 2022). We therefore treat the 850–950 cm^{-1} interval as a composite region demonstrating overlapping vibrational environments rather than uniquely resolved Q^n species. Similarly, in the higher-Raman-shift region, the 1100 cm^{-1} band characteristic of NS2 glass becomes absent after dissolution; instead, a broader contribution centered at $\sim 1050 \text{ cm}^{-1}$ emerges (Fig. 3 and S2). Because multiple assignments have been proposed for 1000–1050 cm^{-1} interval in hydrous alkali silicate systems (e.g., Le Losq et al., 2015a, 2022; Mysen, 2009; Spiekermann et al., 2012; Steele-MacInnis and Schmidt, 2014 and references therein), we label the components descriptively in the deconvolution (Fig. 5) and focus on reproducible shifts in relative spectral intensity rather than unique structural interpretations. Notably, although individual components in the Si-O

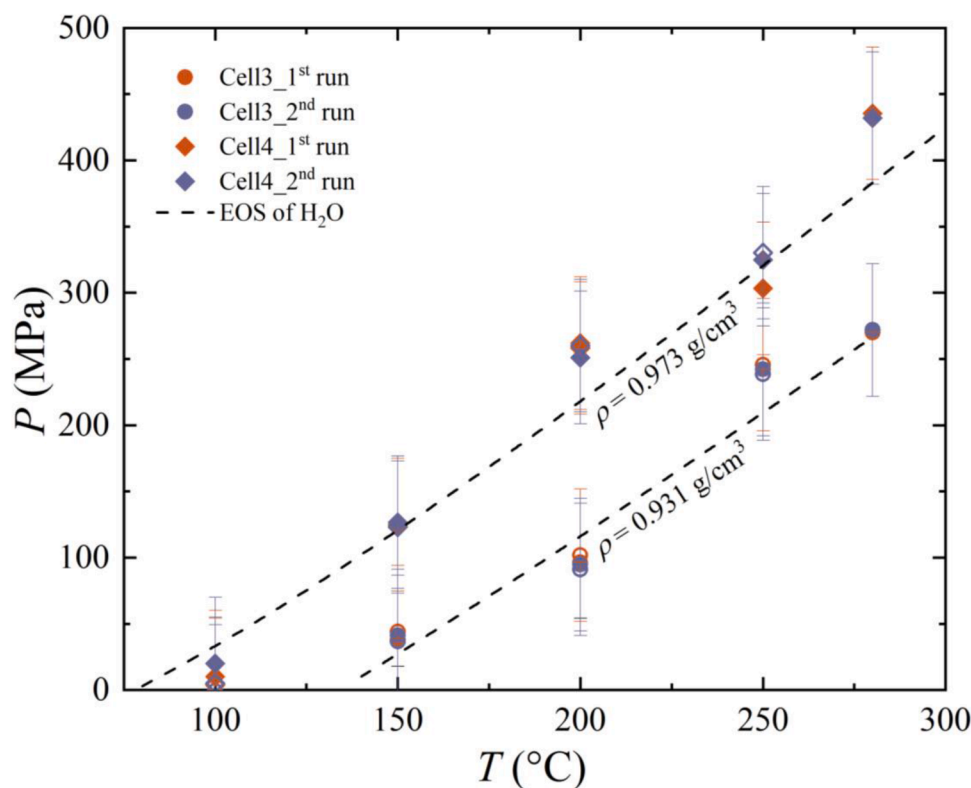


Fig. 4. Isochores determined using the zircon pressure sensor, with solid symbols representing heating and hollow symbols representing cooling process, based on the calibration by Schmidt et al. (2013). The densities and corresponding isochores (dashed lines) are derived from the EoS of H₂O (Wagner and Pruß, 2002). The estimated pressure uncertainty from the Raman band of zircon is ± 50 MPa (Schmidt et al., 2013).

stretching envelope are not uniquely attributable in hydrous liquids, the relative intensity of lower- (<950 cm^{-1}) versus higher-Raman-shift (~ 1000 – 1150 cm^{-1}) Si–O stretching contributions is commonly interpreted as reflecting shifts between less connected and/or hydroxylated environments and more connected environments (Le Losq et al., 2015a, 2022; Steele-MacInnis and Schmidt, 2014). We therefore use the fitted components to track relative changes in Si–O bonding environments (Fig. 5 and S2).

The O–H stretching region (Fig. 5b) is fitted following Brubach et al. (2005) by: ν^6 (~ 3300 cm^{-1}) corresponding to strongly H-bonded “network” water, ν^7 (~ 3450 cm^{-1}) to intermediate H bonding, whereas ν^8 (~ 3600 cm^{-1}) to weakly H-bonded species. With increasing temperature, ν^8 increases while ν^6 decreases (Fig. 5d; Fig. S2d), indicating progressive weakening of hydrogen bonding. This temperature effect has also been reported in water-rich systems (Frantz et al., 1993) and in hydrous silicate melts and coexisting aqueous fluids (Le Losq et al., 2017).

Additionally, at comparable temperatures, the integrated Si–O stretching intensity relative to the O–H stretching envelope is substantially smaller for Cell2 (H₂O-rich) than for Cell6 (silicate-rich). Furthermore, Cell2 exhibits relatively stronger lower-Raman-shift contributions, whereas Cell6 retains substantially stronger higher-Raman-shift components (Fig. 5c and S2c). This compositional contrast indicates that increasing dissolved silicate content is accompanied by an increasing contribution from higher-Raman-shift Si–O stretching environments, consistent with increased connectivity of the dissolved silicate component.

In alkali silicate systems, dissolved water can occupy multiple proton environments, and several reaction pathways may operate depending on composition and polymerization state (Le Losq et al., 2015a, 2015b). In the present low-temperature HNS2 liquids, interpretation is therefore restricted to comparative spectral trends within the Si–O and O–H stretching region, without quantitative inference of molecular versus

hydroxyl water speciation.

4.2. Viscosity changes in HNS2

The strongest manifestation of viscosity variability emerges when comparing experiments with differing initial silicate/H₂O ratios (Table 2 and Fig. 2). Increasing the silicate fraction from 25 to 70 wt.% results in approximately one order-of-magnitude rise in viscosity, highlighting the dominant influence of dissolved silicate concentration and associated network structure in governing liquid viscosity (Mysen and Richet, 2005). Nonetheless, the maximum viscosity measured here (9.6×10^{-3} Pa·s for 70 wt.% NS2 at 200 °C) remains significantly lower than that of anhydrous NS2 melt at magmatic temperatures (~ 46.7 Pa·s at 1350 °C; Dingwell, 1986). Although the absolute values are not directly comparable across such disparate temperatures, the contrast illustrates that additions of water will drastically alter the NS2 network, producing low-viscosity HNS2 at sub-magmatic conditions. Consistent with this, Fig. S2c shows a distinct redistribution of spectral intensity between 100 and 150 °C following glass dissolution. The composition-viscosity pattern is consistent with that of albite–H₂O system (Audetat and Keppler, 2004). They proposed viscosity exponentially decreases with the first 10–20 wt.% H₂O, followed by a more modest linear decline at higher H₂O contents. This is because at elevated H₂O concentrations, additional H₂O predominantly remains molecular rather than actively participating in network modification (Stolper, 1982).

Another significant observation is the weak apparent temperature dependence of viscosity in H₂O-rich liquids (< 30 wt.% NS2) along the experimental paths, as shown by runs Cell2 and Cell5 (Fig. 2). This differs from the behavior of typical silicate melts, which usually exhibit a pronounced decrease in viscosity as T increases (Giordano et al., 2008). The negative T effect is also observed in hydrous NS3.3 liquids at 1 atm (Yang et al., 2008; Fig. 2). Instead, the observed flat trend in our study can be interpreted within the context of our near-isochoric experimental

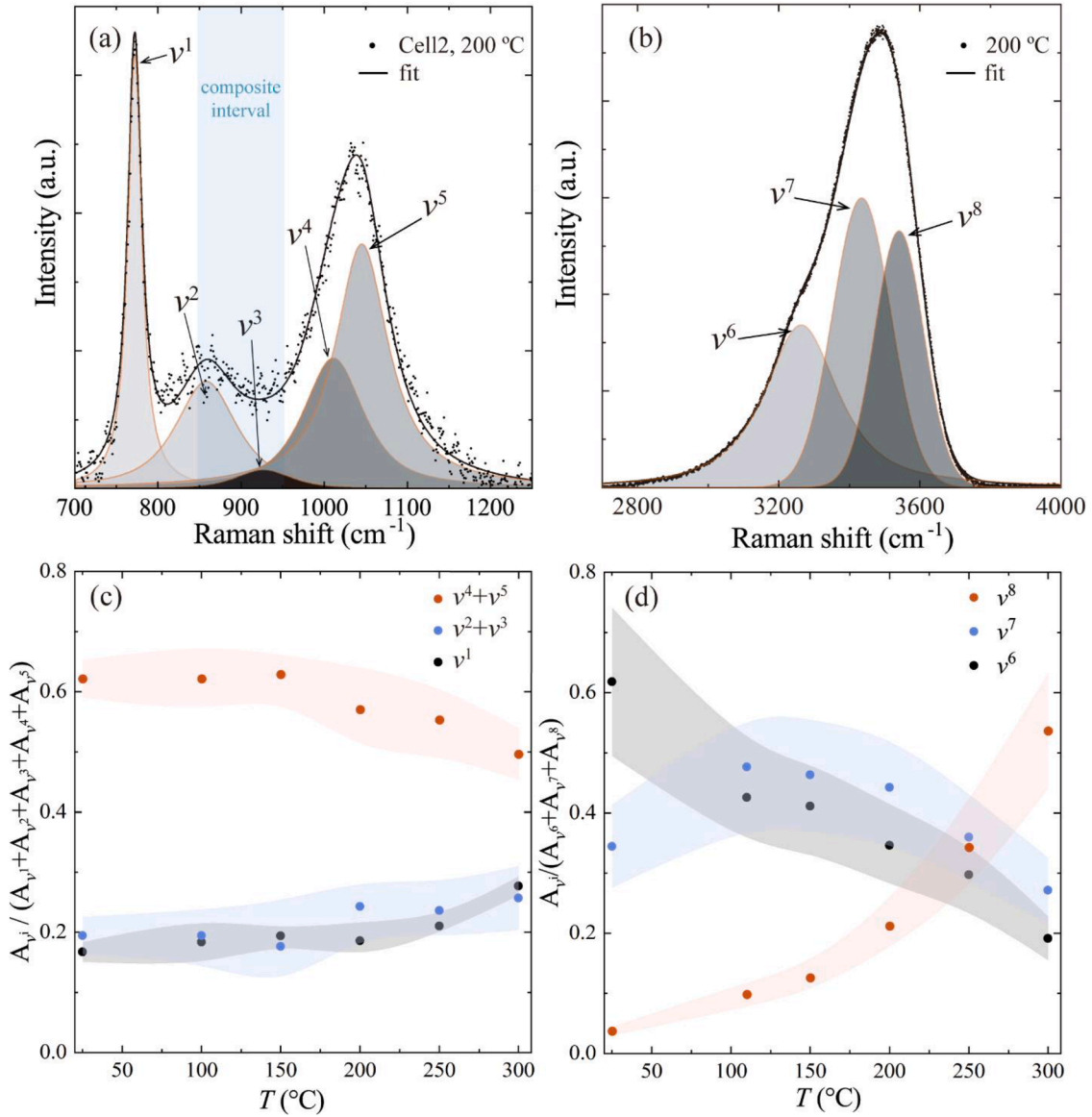


Fig. 5. Representative deconvolution of Raman spectra for Cell2 at 200 °C in the (a) 700–1200 cm⁻¹ and (b) 2700–4000 cm⁻¹ spectral ranges, together with the relative integrated areas for (c) grouped Si-O components and (d) the O-H components as a function of T . In (a), ν^1 – ν^5 denote neutral fitted components used to compare the Si-O stretching envelope and are combined into three groups in (c): 770 cm⁻¹ as ν^1 ; the 850–950 cm⁻¹ composite interval as $\nu^2+\nu^3$; and the 1000–1150 cm⁻¹ contribution as $\nu^4+\nu^5$. The shaded 850–950 cm⁻¹ interval is treated as a composite region that may include overlapping contributions from Si-O stretching environments and T-OH ($T = \text{Si, Al}$) vibrations (Le Losq et al., 2015a, 2022). The O-H region is parameterized by three components (ν^6 – ν^8) representing strongly, intermediately, and weakly hydrogen-bonded species, respectively (Brubach et al., 2005). Bands in (c–d) represent the propagated uncertainties derived from the deconvolution and integration. See main text for further discussion.

setup in HDAC (Shen et al., 1993), where pressure increases concurrently with temperature. We thus propose that this trend reflects two competing effects: a temperature-driven viscosity decrease and a pressure-driven viscosity increase, rather than anomalous liquid behavior. This interpretation is supported by comparing runs Cell2 and Cell5 (Fig. 2), which share a similar NS2/H₂O ratio (Table 1): at similar temperatures (e.g., ~200 °C), the higher-pressure run (Cell2) exhibits a systematically higher (~0.1 log unit) viscosity. Raman spectra further indicate that, in these H₂O-rich compositions, the dissolved silicate occurs predominantly as less connected Si-O environments (Fig. 3 and 5), suggesting that bonding characteristics are similar to smaller polymeric units mixing with H₂O in dilute liquids (Hack and Thompson, 2011). Consistently, pure water reference curves (Fig. 2) exhibit viscosity increases with elevated P at fixed T . A similar positive pressure dependence has also been reported for 2 wt.% NaCl bearing H₂O up to 1 GPa

(King et al., 1992). Moreover, at constant density comparable to that in run Cell2, pure-water viscosity shows a minor decrease in viscosity between 130 and 250 °C. Consequently, the dominant factor controlling the overall viscosity trend across the entire compositional range is the concentration of dissolved silicate.

4.3. Mobility and percolation velocity of HNS2

To evaluate flow rates of hydrothermal liquids within a porous oceanic crust, we applied Darcy's law to calculate the percolation velocity, using HNS2 as a proxy for magmatic-to-hydrothermal liquids under typical oceanic crust conditions:

$$V_{l-m} = \left(\frac{k}{\phi}\right) \cdot \left(\frac{\Delta\rho_{s-l}}{\eta_l}\right) \cdot g \quad (6)$$

where $V_{l,m}$ is the percolation velocity; k , the permeability; ϕ , interconnected porosity in the solid and fixed at 2%; $\Delta\rho_{s-l} = \rho_s - \rho_l$, density contrast between surrounding solid crust (ρ_s) and the liquid (ρ_l); η_l , liquid viscosity. We adopted an average oceanic crust density (ρ_s) of 2900 kg/m³ (Hasenclever et al., 2014) with HNS2 densities calculated from the NS2/H₂O ratios by assuming ideal mixing. In Eq. (6), one important parameter is $\Delta\rho_{s-l}/\eta_l$, which is commonly referred to as liquid mobility.

Our HNS2 viscosity data at 200 °C is fitted to a linear dependence on H₂O content over the studied range and extended to the pure H₂O endmember (Fig. 6a). This linear model is a justified approximation for this compositional range (> 30 wt.% H₂O), consistent with that reported by Audetat and Keppler (2004) for the albite-H₂O system, which also exhibits a linear viscosity decline at H₂O concentrations above 20 wt.%. With H₂O added to the liquid from 30 to 80 wt.%, the mobility of HNS2 increases monotonically by one order of magnitude, despite the decrease in density (Fig. 6b). Without detailed knowledge of permeability, we calculated percolation velocities based on a range of constant permeability, from 10⁻¹⁶ to 10⁻¹³ m². The resulting percolation velocities increase with decreasing NS2/H₂O nearly linearly on a logarithmic scale (Fig. 6c). With assumed permeability in the 10⁻¹⁶ to 10⁻¹⁵ m² range, the calculated percolation velocities are comparable in order of magnitude to Darcy-scale porous-flow velocities predicted in vent-constrained hydrothermal circulation simulations (5–158 m/yr) for Wocan-1 Hydrothermal Field, Indian Ocean (Zeng et al., 2023). Our scaling provides a physical basis for such rapid ascent where permeability permits and indicates that silicate enrichment would damp ascent speeds unless compensated by permeability.

5. Implications for oceanic crustal heat flow

To assess how dissolved silicate may influence deep hydrothermal circulation, we conducted 2-D porous-flow simulations under three end-member liquid scenarios: (1) pure H₂O (reference model; Fig. S9), (2) a solubility-constrained silicate-bearing case (Fig. 7) with temperature-dependent NS2 contents prescribed from experimental silicate solubility in H₂O (Manning, 2018; Fig. S6), and (3) an NS2-enriched end-member (Fig. 8) in which the liquid contains 20 wt.% NS2 at the 800 °C isotherm and decreases linearly to 0 wt.% at 2 °C. Scenario (2) provides a thermodynamically constrained and conservative baseline for testing

whether modest dissolved silicate can measurably alter circulation, even if silicate solubility is expected to be significantly higher in the per-alkaline HNS2 system investigated experimentally (Lundstrom, 2020).

In the reference model (pure H₂O, Fig. S9), the most intense upward mass flow flux is focused at the ridge axis (Fig. S9b) and supplied by two coupled components: a shallow on-axis convection above the melt lens and a ridge-proximal deep off-axis recharge limb (i.e., downwelling and lateral inflow several kilometers off-axis that returns toward the ridge and feeds axial discharge) that penetrates to ~5–6 km depth below seafloor, approximately to the brittle-ductile transition depth (Fig. S9). Within the near-axis (0–8 km) window highlighted by Hasenclever et al. (2014), this pure-water model reproduces the first-order “hybrid” circulation pattern. We use this near-axis agreement as a baseline check on model behavior close to the ridge, while noting that our simulations extend farther off-axis region (to 25 km) and therefore provide additional “distal off-axis recirculation” expressed as low-flux return flow and more diffuse discharge toward the outer part of the domain.

Introducing dissolved silicate does not significantly suppress or reorganize the ridge-proximal hybrid structure (Fig. 7 and 8). The shallow on-axis cell and the deep off-axis recharge limb feeding axial discharge remain present with broadly similar geometry. Instead, dissolved silicate mainly affects the distal recirculation. In the solubility-constrained silicate-bearing case (Fig. 7), flux magnitudes decrease modestly relative to the pure-water reference, and distal off-axis recirculation becomes weaker and less continuous. In the NS2-enriched case (Fig. 8), distal circulation is further weakened and broadened laterally, producing more diffuse return flow and a reduced tendency for vigorous far-field circulation. Quantitatively, we define the distal domain occurring at distance > 8 km and compute the distal fraction that occurs in this domain as the portion of total upward advective heat flux released through the seafloor boundary (Fig. S10). The distal fraction decreases systematically from the pure-H₂O case to the solubility-constrained silicate-bearing case and is nearly halved in the NS2-enriched case (Fig. S10). This trend is consistent with the combined effects of higher liquid viscosity (greater viscous resistance) and density (reduced buoyancy forcing), which preferentially impede low-gradient convection in the far field compared to ridge-proximal flow.

A key control not treated here is porosity and permeability evolution driven by cracking, alteration, and mineral precipitation, due to the poor knowledge of the changes in permeability with silica deposition. In particular, silica precipitation and albitization near the gabbro-sheeted

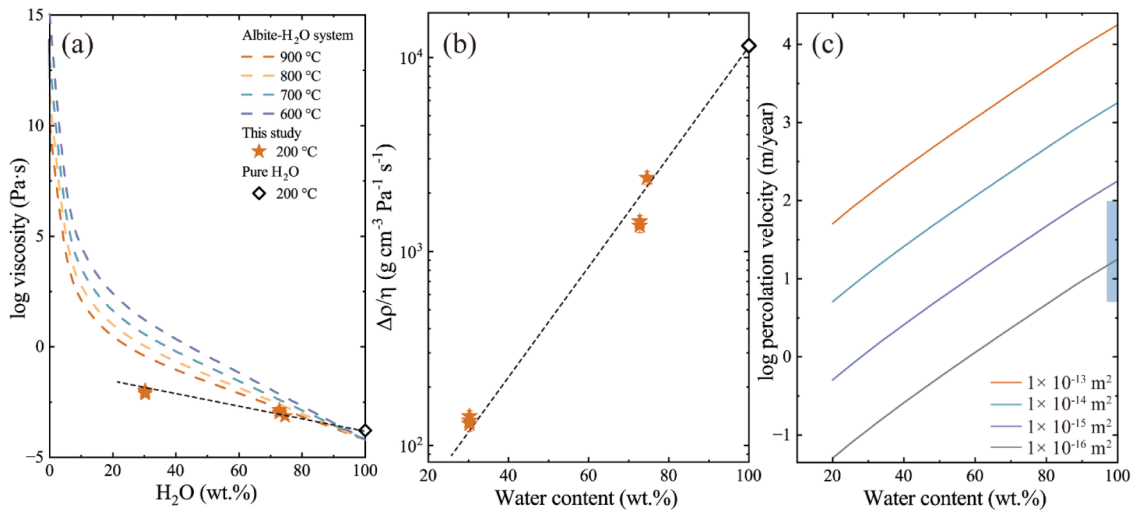


Fig. 6. (a) Viscosities of HNS2 (stars) as a function of water content at 200 °C, black dashed line is the linear fit to the data. Albite-H₂O viscosities at 600–900 °C are also shown as dashed lines (Audetat and Keppler, 2004). Open diamond is pure H₂O at 200 °C and 150 MPa, calculated based on the equations from the IAPWS Formulation 2008 for the Viscosity of Ordinary Water Substance (Huber et al., 2009). (b) Mobility of HNS2 at 200 °C. (c) Darcy percolation velocity as a function of water content calculated using Darcy's law. Colored lines represent different permeability of oceanic crust adopted in the simulations, blue band indicates the range of modeled Darcy-scale porous-flow velocities inferred from Wocan-1 Hydrothermal Field (Zeng et al., 2023).

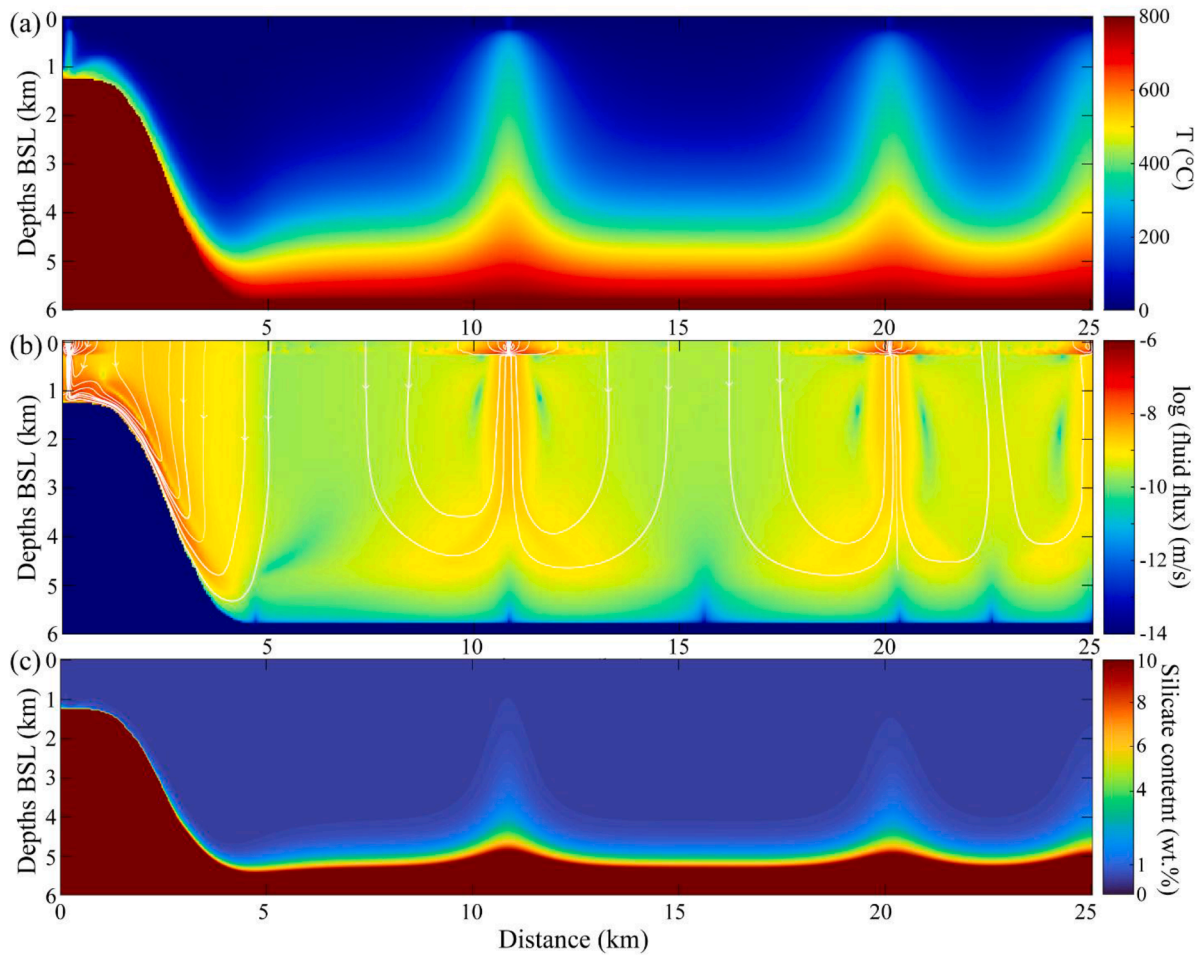


Fig. 7. 2-D hydrothermal flow model for a solubility-constrained silicate-bearing liquid, with temperature-dependent dissolved silicate contents prescribed from experimental silicate solubility in H_2O (Manning, 2018; Fig. S6) for a 6 km-deep, 25 km-wide section beneath the EPR. (a) Temperature distribution; (b) liquid flux magnitude map with white streamlines indicating flow direction; (c) dissolved silicate (NS2) content in wt.%. The left boundary represents the ridge-axis heat source; the seafloor is held at 2°C and open to outflow. Closure of pore space associated with the brittle-ductile transition is represented by reducing permeability to 10^{-18} m^2 at 800°C . Accordingly, no significant hydrothermal flow occurs at temperatures above 800°C . Vertical up-flow zones at ~ 11 and 21 km arise spontaneously from the modeled hydrothermal convection.

dike transition could locally modify permeability and promote lateral redistribution of liquids. Geological observations motivate this possibility: the near-complete geologic exposure of the gabbro-sheeted dike boundary in Troodos (Lundstrom, 2020) finds abundant plagiogranites, rocks composed dominantly of plagioclase and quartz (the two minerals in equilibrium with HNS2). Fluid-inclusion analyses from Troodos samples further reveal that quartz in plagiogranites contains fluid inclusions that homogenize at temperatures between 250 and 540°C (Kelley et al., 1992). A conceptual extension is that reaction-driven permeability reduction could introduce lateral flow components not captured by our fixed-permeability framework. Lundstrom (2020) proposed that a thin plagiogranite layer may exist ubiquitously between gabbros and sheeted dikes, with quartz deposition and albitization producing low-permeability rocks that could promote lateral diversion of liquids away from the ridge. In this setting, lateral transport by low-viscosity silicate-bearing liquids could be relevant to interpretations of geophysical and field observations that point to laterally redistributed heat and fluid flow, including off-axis melt anomalies at the EPR (Han et al., 2014) and ridge-flank heat loss at basement outcrops (Hutnak et al., 2008).

Nevertheless, our temperature fields indicate where such permeability evolution would be most effective. In all models we performed, $< 100^\circ\text{C}$ temperatures exist in the upper $1\text{--}2$ km (near the sheeted dike gabbro depth). This reflects the fact that cold seawater can easily

penetrate the volcanic and even sheeted dike layers via off axis cracking. Upward-moving HNS2 will therefore experience their strongest cooling across this shallow transition, particularly just off axis where ridge-proximal upflow encounters cold recharge. Over time, this is the depth interval where silica supersaturation and quartz precipitation are most likely, potentially reducing permeability and reorganizing discharge. To illustrate the dynamical leverage of shallow permeability, we performed a schematic sensitivity test using the NS2-enriched setup of Fig. 8 but artificially reducing permeability in the upper $0\text{--}1.2$ km near axis (Fig. S11). This imposed shallow “sealing” localizes discharge and strongly alters the distribution of venting to off-axis region, highlighting that even simple, imposed permeability reduction can reorganize circulation. We emphasize that Fig. S11 is not intended as a quantitative precipitation model (nor a representation of black-smoker permeability). Rather, it demonstrates that any realistic treatment of silica deposition will require a coupled, spatially heterogeneous porosity and permeability evolution.

It is also important to acknowledge limitations associated with necessary simplifications made in this study. Our experiments focus on the simple HNS2 system, whereas natural silica-bearing hydrothermal liquids will contain additional components (e.g., Al, Ca, Mg, Cl, S, Fe). The influence of dissolved Al on network polymerization and viscosity is composition- and charge balance-dependent and can be non-linear (Giuliani et al., 2025). However, quartz-albite saturated liquids in

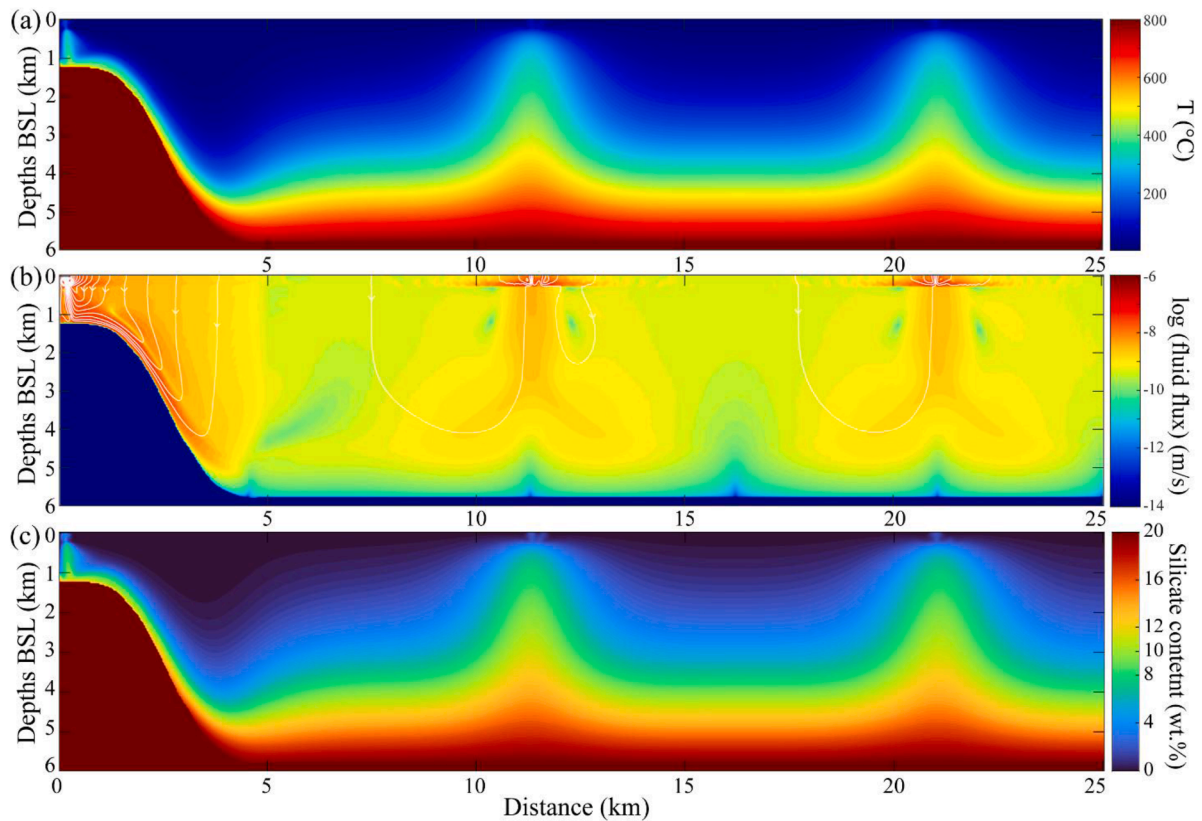


Fig. 8. Same model configuration as Fig. 7 but NS2-enriched endmember, where the liquid contains 20 wt.% NS2 at the 800 °C and decreases linearly to 0 wt.% at 2 °C. (a) Temperature distribution; (b) liquid flux magnitude map; (c) silicate (NS2) content in wt.%. Boundary conditions and the 800 °C permeability cap are identical to Fig. 7.

Lundstrom (2020) suggest Al_2O_3 contents at 330–400 °C can be <1 wt.%. On the other hand, chloride-bearing species and dissolved sulfur can occur at substantial concentrations in liquids and modulate their chemistry and properties (Manning, 2018 and references therein). Network-modifying cations (e.g., Ca^{2+} , Fe^{2+} , and Mg^{2+}) will also influence viscosity and density. These effects cannot be addressed in our current model. Extending viscosity constraints to multi-component liquids represents an important direction for future experimental work. Nevertheless, the first-order trend emphasized here, that increasing dissolved silicate raises viscosity and density and therefore damps off-axis convection should be robust.

Despite these simplifications, our simulations show a first-order outcome: incorporating dissolved silicate preserves the ridge-proximal hybrid circulation but systematically reduces distal off-axis (> 8 km) advection. Consequently, heat-flow models that adopt pure water as the circulating fluid may overestimate the intensity and spatial extent of far-field advective upflow required to match conductive background. Besides, in silicate-bearing cases, the remaining off-axis discharge is preferentially expressed near boundaries between circulation cells (Figs. 7c and 8c), where warm liquids rise and cool rapidly. If such liquids approach silica saturation during cooling, these zones would also be favorable sites for mineral precipitation and localized permeability modification. Kilometer-scale spatial organization of hydrothermal alteration in gabbroic sections, such as the regularly spaced alteration zones described in the Semail ophiolite (Coogan et al., 2006), is qualitatively consistent with this style of localization. In addition, proposed upward migration of REE-bearing fluids from plagiogranitic environments in Troodos (Anenburg et al., 2015) suggests that silica-bearing liquids may enhance trace-element transport. In that sense, incorporating HNS2 into quantitative models may be important not only for heat extraction but also for chemical fluxes in oceanic crust

hydrothermal systems. Thus, the potential roles of HNS2 in both elemental transport and heat transfer should be incorporated into future, more fully coupled models that treat permeability evolution and more realistic liquid compositions.

6. Conclusion

We quantified the rheology and structure of hydrous sodium-disilicate liquids (HNS2) under hydrothermal to sub-magmatic conditions up to 350 °C and 440 MPa in a hydrothermal diamond-anvil cell and linked those measurements to a model of crust-scale porous-flow circulation. Falling-sphere experiments show that viscosity is controlled primarily by composition: H_2O -rich liquids are nearly an order of magnitude less viscous than silicate-rich liquids at the same temperature. Raman spectroscopy indicates systematic shifts toward less connected Si-O environments with increasing H_2O and temperature. Between 200 and 250 °C, the viscosity of HNS2 decreases by about an order of magnitude from silicate-rich to H_2O -rich compositions at fixed permeability and porosity, resulting in large variations in calculated percolation velocity of hydrothermal liquids.

Two-dimensional porous-flow models show that adding modest dissolved silicate to deep circulating liquids modifies circulation while preserving the ridge-proximal primary cell: off-axis circulation weakens at depth and becomes more laterally diffuse. Consequently, hydrothermal heat-flow models that assume pure water likely overestimate the vigor and lateral extent of off-axis advective heat transport required to match background conductive heat flow. Together, these results indicate that silicate-bearing, low-viscosity liquids can contribute meaningfully to lower-crustal heat extraction and should be incorporated into quantitative models of oceanic-crust thermal evolution.

CRediT authorship contribution statement

Qi Chen: Writing – original draft, Visualization, Software, Methodology, Investigation, Formal analysis, Conceptualization. **Craig Lundstrom:** Writing – review & editing, Resources, Methodology, Funding acquisition. **Yanchong Li:** Writing – review & editing, Software, Methodology. **Young Jay Ryu:** Writing – review & editing, Methodology, Formal analysis. **Tony Yu:** Writing – review & editing, Methodology, Formal analysis. **Stella Chariton:** Writing – review & editing, Methodology, Formal analysis. **Dongzhou Zhang:** Writing – review & editing, Methodology, Formal analysis. **Vitali Prakapenka:** Writing – review & editing, Methodology, Formal analysis. **Yanbin Wang:** Writing – review & editing, Supervision, Resources, Funding acquisition, Formal analysis, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Supplementary materials

Supplementary material associated with this article can be found, in the online version, at [doi:10.1016/j.epsl.2026.120059](https://doi.org/10.1016/j.epsl.2026.120059).

Data availability

Data will be made available on request.

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