

Key Points:

- Thermal conductivity of fcc-Fe varies monotonically with pressure and temperature
- Thermal conductivity values are consistent with a Sommerfeld-like Lorenz number for fcc-Fe
- Core thermal conductivity of Mercury estimated to be between $38 \text{ W m}^{-1} \text{ K}^{-1}$ and $23 \text{ W m}^{-1} \text{ K}^{-1}$ at 25 GPa for realistic bulk core compositions

Supporting Information:

Supporting Information may be found in the online version of this article.

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Citation:

Edmund, E., Geballe, Z. M., Rivoldini, A., Tosi, N., Chariton, S., Prakapenka, V. B., & Goncharov, A. F. (2026). Thermal conductivity of iron at the conditions of small planetary cores. *Journal of Geophysical Research: Planets*, 131, e2025JE009205. <https://doi.org/10.1029/2025JE009205>

Received 23 MAY 2025

Accepted 18 SEP 2026

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Thermal Conductivity of Iron at the Conditions of Small Planetary Cores

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Abstract Planetary interiors cool and evolve over geologic time, and their thermal evolution depends on heat transport through their constituent layers. Here, the thermal conductivity of iron has been measured between 19 and 65 GPa and between 1700 and 2500 K in order to constrain its value at the conditions of small planetary cores. Our results are comparable to the low end of electrical resistivity experiments on fcc-Fe at comparable conditions, suggesting that the relationship between electrical and thermal conductivity described by the Wiedemann-Franz law provides a reasonable description of heat transport in fcc-Fe at planetary core conditions. The thermal conductivity of fcc-Fe at high pressures and temperatures places an upper bound on the thermal conductivity of Mercury's core, with a value of 52(7) W/m/K at 40 GPa, which corresponds to the pressure at the center of the core. Accounting for the effects of silicon and sulfur alloying reduces Mercury's core conductivity by 10%–40% for realistic core compositions, which satisfy constraints on Mercury's core density.

Plain Language Summary The thermal conductivity of iron places constraints on the magnitude of heat flow within planetary cores. Here, we have investigated the thermal conductivity of iron at the pressure and temperature conditions of small rocky planets such as Mercury and Mars. We provide new insight into discrepancies between direct measurements of thermal conductivity with inferred values based on electrical measurements for face-centered cubic (fcc) iron. We find that the thermal conductivity of fcc iron is 52(7) W/m/K at the central pressures of Mercury and provide estimates for the thermal conductivity values of possible core-forming alloys at Mercury's core conditions.

1. Introduction

Mercury has a magnetic field in the present-day, which is generated by convective flow of liquid iron alloy within the core (Anderson et al., 2011). The specific dynamo mechanisms which drive Mercury's magnetic field, as well as the interior structure and thermal state of the core remain debated (Davies et al., 2024; Genova et al., 2019; Knibbe et al., 2025; Kolhey et al., 2025; Steinbrügge et al., 2021; Takahashi et al., 2019). The field strength, geometry and strong non-dipolar features of Mercury's magnetic field are related to the processes which drive convection, as well as the thermal evolution of the core over geologic time. Mercury's dynamo is thought to be a consequence of convection due to inner core crystallization, however there are several possible explanations for the observed characteristics of the magnetic field. The observed magnetic field requires an intrinsically weak dynamo, thermal or chemical stratification of the core (Kolhey et al., 2025; Takahashi et al., 2019). The degree to which thermal stratification can occur depends on the thermal conductivity of the core-forming iron alloy (Davies et al., 2024; Knibbe et al., 2025). If the thermal conductivity of the core-forming material is too high, thermal evolution models indicate that a thermally stratified layer expands to the whole liquid core, while if the thermal conductivity is too low no thermal stratification can occur and other mechanisms are needed to explain observations (Davies et al., 2024; Knibbe et al., 2025). Chemical stratification has also been proposed to explain key characteristics of Mercury's magnetic field, where either a solid metallic layer or a chemically distinct liquid layer is present at the core-mantle boundary (CMB) (Chabot et al., 2014; Dumberry & Rivoldini, 2015; Hauck et al., 2013). Chemical stratification of a liquid layer requires liquid-liquid immiscibility or the formation of iron-rich snow, which may require bulk core compositions which are unrealistically enriched in elements lighter than iron (Davies et al., 2024; Dumberry & Rivoldini, 2015; Edmund et al., 2022; Genova et al., 2019; Steinbrügge et al., 2021). The formation of a solid, electrically conductive layer at the CMB would require the accumulation of iron sulfides which may not occur for the highly reducing conditions expected in Mercury's interior (Namur

et al., 2016; Pommier et al., 2025). A dynamo driven by weak compositional convection in the presence of a thermally stratified layer provides perhaps the simplest explanation for Mercury's core dynamics and the resulting magnetic field (Knibbe et al., 2025), but more investigations into the physical properties of core-forming materials are needed to address uncertainties surrounding the thermal state and thermal evolution of the core.

The thermal conductivity of iron and iron alloys are uncertain at the conditions relevant to small planetary cores such as Mercury ($P = 5\text{--}40$ GPa (Rivoldini & Van Hoolst, 2013)) and Mars ($P = 18\text{--}40$ GPa (Stähler et al., 2021)). For iron, reported experimental thermal conductivity values at about 25 GPa vary from 25 W/m/K to 87 W/m/K (Andrault et al., 2025; Ezenwa & Yoshino, 2021; Hasegawa et al., 2024; Konôpková et al., 2016; Ohta et al., 2016; Yong et al., 2019). This large spectrum of possible values results in divergent descriptions of the present-day thermal state and thermal evolution of Mercury, as well as reflecting the technical challenges inherent in measuring this quantity at extreme conditions (Text S1 in Supporting Information S1). With such large differences between reported thermal conductivity values for iron, it is similarly challenging to assess the effect of composition, as Mercury's core could contain significant quantities of silicon, as well as smaller amounts of sulfur and/or carbon (Nittler et al., 2018).

The thermal conductivity of a metal is either measured directly, or inferred from measurements of electrical conductivity. Thermal conductivity values derived from electrical experiments are determined using the Wiedemann-Franz Law:

$$k_{el} = L \cdot T \cdot \sigma \quad (1)$$

Where k_{el} is the electronic contribution to the material thermal conductivity in W/m/K, L is the Lorenz number of the material ($\text{W}\Omega/\text{K}^2$), T is temperature (K) and σ is the electrical conductivity ($1/\Omega/\text{m}$). Both σ and k are independently important for understanding dynamo generation in planetary cores; the electrical currents generated in a convecting iron-alloy core are dependent on the electrical conductivity of the core-forming material, which in turn will influence the characteristics and field strength of the resulting magnetic field (Driscoll & Du, 2019; Johnson et al., 2016). In parallel, the thermal conductivity of the material controls the amount of energy available to drive convective motion in the core relative to the cooling rate of the core (Driscoll & Du, 2019). Core processes which produce a viable dynamo must provide enough heat energy to exceed what is lost due to conduction, or provide sufficient chemical buoyancy to drive convection without thermal contributions. Understanding the thermal conductivity of iron and core-forming alloys, as well as the connection between the thermal and electrical conductivity in these alloys is fundamental to our understanding of telluric planetary dynamos.

Here we applied laser-flash heating to make direct measurements of the thermal conductivity of fcc-Fe at pressures between 15 and 60 GPa, and at temperatures above 1700 K (McWilliams et al., 2015), employing new technical developments to refine the thermal conductivity of fcc iron at the conditions of small planetary cores. Such data is necessary to better understand existing discrepancies between studies of the thermal and electrical conductivity of iron, described further in Text S1 of Supporting Information S1. Furthermore, we apply these values to explore the effect of composition on the thermal conductivity and thermal evolution of Mercury's core.

2. Methods

2.1. Sample Preparation and Characterization

Fe disks were initially cut from a wire (Sigma Aldrich, 99.995%) and pressed into the desired thickness between two diamond culets. These disks were then sandwiched between plates of KCl and loaded into diamond anvil cells. The diamond anvils had culet diameters ranging between 200 μm to 300 μm , and were used with 250 μm Re gaskets. The gaskets were preindented to initial thicknesses ranging from 30 μm to 60 μm , after which a cavity was drilled in the gasket to roughly half the diameter of the diamond culet. Pressure was determined using ruby fluorescence, the Raman shift of the T_{2g} diamond phonon of the stressed anvil, or the equation of state of KCl (Akahama & Kawamura, 2006; Dewaele et al., 2012; Shen et al., 2020) depending on the experimental run. Sample geometry was determined in situ using the white light interference technique (WL) (Konôpková et al., 2016; McWilliams et al., 2015) to measure the culet-culet and sample-culet distances employing a LEUKOS supercontinuum (SC) laser and the known refractive index of KCl (Johannsen et al., 1997), recently extended to Mbar pressure (Lobanov & Geballe, 2022). KCl refractive indices are reported in Data set S1 of

Supporting Information S1. All in situ measurements of sample geometry are compared with ambient pressure measurements made on sample cross-sections using a focused ion beam (FIB). KCl thickness was observed to differ significantly from that measured at high pressures (Lobanov & Geballe, 2022), and differences in Fe thickness determined with different techniques are shown in Figure S1 in Supporting Information S1. Chemical integrity of the Fe samples was verified with synchrotron X-ray diffraction at GeoSoilEnviroCARS at the Advanced Photon Source (APS), Sector 13 ID-D, or by using FIB and scanning electron microscopy (SEM). Details of specific runs are noted in Data set S1 of Supporting Information S1.

Synchrotron X-ray diffraction at APS (Sector 13 ID-D) was used in select experimental runs to measure the density of KCl and Fe at high pressures. Specific runs where XRD was collected are noted in Data set S1 of Supporting Information S1. The X-ray beam was focused to a 3 μm by 3 μm (horizontal by vertical) spot size, and diffraction images were collected using a Pilatus 1M detector. Diffraction images were reduced to 1D patterns using Dioptas (Prescher et al., 2015), and unit cell volumes of observed phases were determined using Le Bail refinements of the diffraction patterns as implemented in JANA2006 (Václav et al., 2014). Diffraction data shows that pressure gradients between spots on the same sample are smaller than stated uncertainties. Figure S2 in Supporting Information S1 shows the absence of carbide formation after high temperature experiments.

2.2. Thermal Conductivity Measurements

The details of laser flash heating (FH) at the Earth & Planets Laboratory (EPL) have been described previously (Geballe et al., 2020; Konôpková et al., 2016; McWilliams et al., 2015). Infrared light from a YLF continuous-wave (CW) laser (IPG Photonics, 1064 nm) is split into two optical paths and impinges on opposing, parallel surfaces of a flat iron plate loaded in a diamond anvil cell. Intensity of the laser is increased until thermal emissions are sufficiently intense to result in measurable temperatures on the Sydor streak camera. The time-dependent heat wave which propagates through the sample is generated using a pulsed infrared laser, focused onto one side of the sample (hereafter the downstream side, or DS), which is aligned coaxially with the DS CW laser beam. Separate measurements are made on both the DS and upstream (US) sides of the sample to measure the time-dependent diffusion of heat through the sample. The light from the pulsed laser (IPG Photonics) of wavelength 1,064 nm, is directed through a Pockels cell (FastPulse Technology Inc.) which acts to modulate the polarization of the laser pulse as a function of time. This allows for fast modulation of the laser pulse envelope, resulting in a shorter heating event, such that the entirety of the DS heating event resolves over the timescale of the 10 μs streak camera window. Fast modulation of the pulsed laser with a Pockels cell greatly simplifies the shape of the laser pulse for modeling, as the laser pulse delivered to the sample is close to a square wave. The pulsed laser is triggered synchronously with the streak camera, with a selected pulse duration of 1.5 μs . In this study, the streak camera is set to collect thermal emissions from the sample in a 10 μs time window, with a repetition rate of 1 kHz. During a measurement the CW laser is on for 5 s, the pulsed laser and streak camera are triggered at 1 kHz and this results in a total accumulation time of 50 ms on the streak camera. Over the 5 s measurement duration, a repetitive, time-dependent change in temperature is generated at the downstream KCl/sample interface with magnitude on the order of 100–200 K. The experimental design of the system requires that the upstream and downstream sides of the sample are measured separately, and so multiple measurements are made for each side to verify reproducibility of the temperature histories. In all measurements made in this way, total heating times remained below 30 s for a given heated spot to minimize the possibility of chemical contamination and thinning of the sample during laser heating (Figures S1 and S2 in Supporting Information S1). Pressure gradients between different measurement positions are smaller than reported uncertainties in pressure.

Additional thermal conductivity measurements of Fe have been performed at GeoSoilEnviroCARS at the Advanced Photon Source, Sector 13 ID-D. These experiments, hereafter referred to as mixed mode (MM) heating, are performed by heating a sample with a CW laser from the US side, and employing a pulsed laser on the DS side to generate a time-dependent temperature perturbation as in standard flash-heating experiments. Time-resolved temperature measurements are made relative to the arrival time of the pulsed laser using a time-gated ICCD spectrometer. Each temperature measurement at a given time window corresponds to roughly 10,000 repetitions of the pulsed laser. Varying the ICCD collection window relative to the arrival time of the laser pulse allows for the collection of temperature histories from a transiently heated sample at microsecond timescales. Pulse profile and ICCD timing is determined with an oscilloscope. Example experiments and their corresponding models are shown in Figure S3 of Supporting Information S1. Details of streak camera timing and calibration are described in Text S2 of Supporting Information S1. Time-dependent heat transfer is investigated using finite

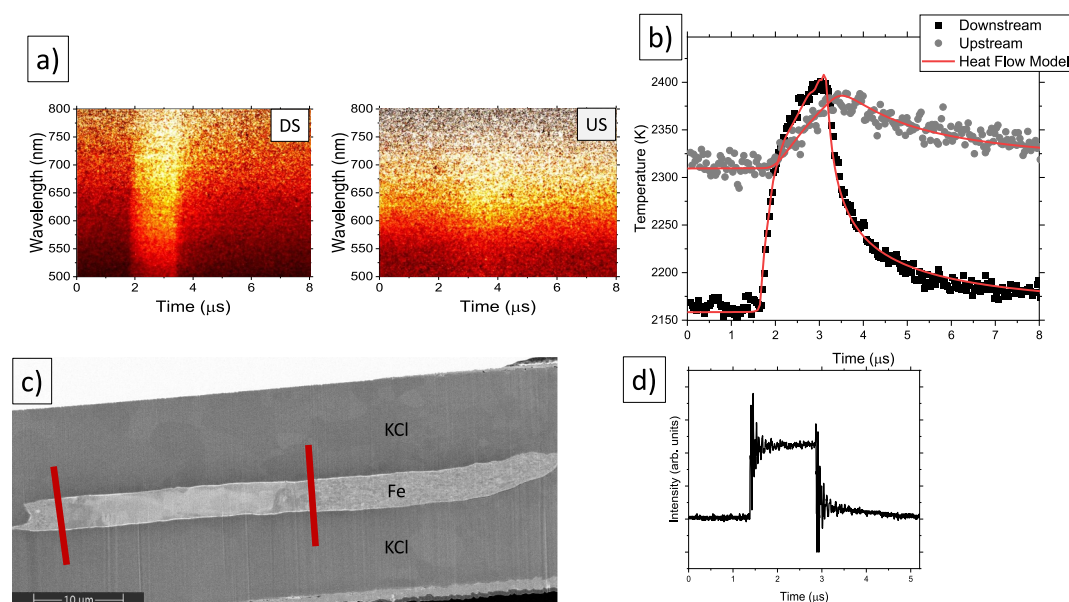


Figure 1. (a) Geometry-corrected streak camera images which record variations in thermal emissions with time from the downstream (DS) and upstream (US) sides of the sample over an 8 μ s window. (b) Measured temperature histories are shown as black (DS) and gray (US) squares, and lines show the output of finite-element models of heat flow through the sample. Model thermal conductivity values for this experiment (Run 9 P2) for iron and KCl are $k_{Fe} = 44$ W/m/K and $k_{KCl} = 10$ W/m/K. (c) Focused Ion Beam cross-section of the Fe foil after recovery from high pressures and temperatures in Run 9 P2. Red lines highlight the area of the sample which has been recrystallized at high temperatures during laser heating. (d) Recorded pulse shape of the pulsed laser probe used to generate the temperature perturbation. Oscillations in intensity at the beginning and end of the pulse are due to electrical interference in the oscilloscope.

element models of diffusive heat transfer as well as the measured temperature histories, sample geometry and either measured or inferred thermophysical parameters. All data and model inputs required to determine thermal conductivity, as well as an example FlexPDE file are reported in Data set S1 of Supporting Information S1 and an open-source repository (<https://doi.org/10.5281/zenodo.22212574>). Modeling of data is performed using standard methods (Andrault et al., 2025; Edmund, Bi, et al., 2024; Edmund, Chuvashova, et al., 2024; Konôpková et al., 2016; Montoya & Goncharov, 2012), described in Text S3 of Supporting Information S1. Text S3, Figures S1, and S4–S7 of Supporting Information S1 provide additional details on the uncertainties in determining thermal conductivity with flash heating at high pressures.

3. Results

The results of flash heating measurements at EPL are shown in Figure 1. Figures 1a and 1b show experimental output from the flash heating experiments - the geometrically corrected streak camera image and the fitted temperature histories, respectively. In Figure 1c, it is observed that laser-induced thinning of the sample is negligible for this set of measurements, and Figure 1d shows an example output of the pulsed laser waveform. The waveform is used to determine the temporal evolution of the pulsed laser intensity for input into FE models to match experimental temperature histories. The resulting thermal conductivity values from these measurements, as well as the P-T conditions investigated are shown in Table 1. Data set S1 of Supporting Information S1 provides more details on model inputs for the experiments.

Recent work suggests that thickness measurements, which are vital to the accuracy of thermal and electrical transport measurements, may be in error by 20% or more when using *ex situ* methods to determine sample thickness at high pressures (Lobanov & Geballe, 2022). In the present study, both *ex situ* and *in situ* methods have been used to investigate the link between thicknesses of recovered samples, and those measured at high pressure. For *ex situ* measurements, thickness is measured after milling with a Focused Ion Beam (FIB), and the ambient pressure thickness is extrapolated to high pressures using an equation of state. In Figure S1 of Supporting Information S1 we find that thicknesses extrapolated to high pressures after recovery can differ by up to 30% from

Table 1

Table of Iron Thermal Conductivity Values From the Present Study, As Well As the P-T Conditions of Measurement

Experiment run	P (GPa)	T_{base} (K)	ΔT_{base} (K)	ΔT_{pls} (K)	k_{Fe} (W/m/K)
1 EPL	33(3)	2105	35	190	45(10)
4 EPL	57(5)	2010	200	50	60(20)
6 EPL	46(4)	1720	50	150	65(25)
6 EPL	46(4)	2020	20	100	65(25)
8 P1 EPL	19(2)	1730	20	130	42(8)
8 P1 EPL	19(2)	1905	25	160	42(8)
8 P1 EPL	19(2)	1930	30	170	42(8)
8 P2 EPL	23(2)	1930	100	150	45(10)
8 P2 EPL	23(2)	2150	50	140	45(10)
9 P1 APS	16(2)	1185	115	730	40(15)
9 P2 EPL	28(3)	2150	50	180	44(7)
9 P2 EPL	28(3)	2225	75	230	44(7)
9 P2 APS	28(3)	1250	225	970	40(10)

Note. Parentheses indicate experimental uncertainties. For discussion of uncertainties in temperature see Text S3 of Supporting Information S1. Finite-element model parameters are reported in Data set S1 of Supporting Information S1.

what is measured directly at high pressures with WL interference, and depend on the compression path. This supports the use of in situ techniques to characterize sample geometry over the use of extrapolations derived from recovered samples.

In the present study, we employ literature thermodynamic models to calculate sample heat capacity as heat capacity values derived from experimental thermal equations of state are usually not constrained by experimental calorimetric data. At 1500 K and ambient pressure, the specific heat for fcc iron is ~ 650 J/kg/K (Desai, 1986), while literature P-V-T thermal equations of state in the Mie-Grüneisen-Debye formalism give values of 573 and 555 J/kg/K respectively (Campbell et al., 2009; Tsujino et al., 2013). We adopt heat capacity values derived from the thermodynamic model of Komabayashi (2014), which gives $C_p = 647$ J/kg/K at 1500 K and 1 bar. Another commonly used thermodynamic model yields $C_p = 697$ J/kg/K for fcc iron at the same conditions (Dorogokupets et al., 2017).

Thermal conductivity measurements have been carried out in the P-T range shown in Figure 2. Circles indicate measurements carried out at EPL, where double-sided CW heating is coupled to single-sided pulsed heating, while triangles indicate measurements carried out at APS, where single-sided CW heating is coupled with single-sided pulsed heating. The larger temperature range measured in experiments at APS is due to technical limitations imposed by the available heating geometry. All thermal conductivity measurements performed at APS were carried out on the same samples as those studied with flash heating at EPL, reported in Table 1. For Runs 4 and 9, KCl and Fe densities were measured directly at high temperatures. Examples of the data and modeling are shown in Figure S3 in Supporting Information S1.

Where data overlaps the two methods yield consistent results, albeit with larger uncertainties for those carried out at APS, and experiments at APS allowed for the extension of thermal conductivity measurements down to about 15 GPa, a pressure where several attempts at EPL were unsuccessful. This is due to the ability of MM heating to measure lower temperatures, alongside the low melting point of iron at pressures below 20 GPa (Ezenwa & Fei, 2023). Uncertainties in thermal conductivity values determined with this method are larger due to greater uncertainties in baseline temperature and the large variation in temperature required for the experiment, however this situation would be improved for experiments at higher pressures (e.g., >40 GPa) where samples are thinner and melting points of the various materials in the DAC are higher.

Several recent studies have investigated the thermal conductivity of fcc iron at the conditions of Mercury's core using electrical methods (Ezenwa & Yoshino, 2021; Ohta et al., 2016; Yong et al., 2019), as well as direct thermal conductivity measurements (Andrault et al., 2025). However, there is no consensus on the inferred k values, or its

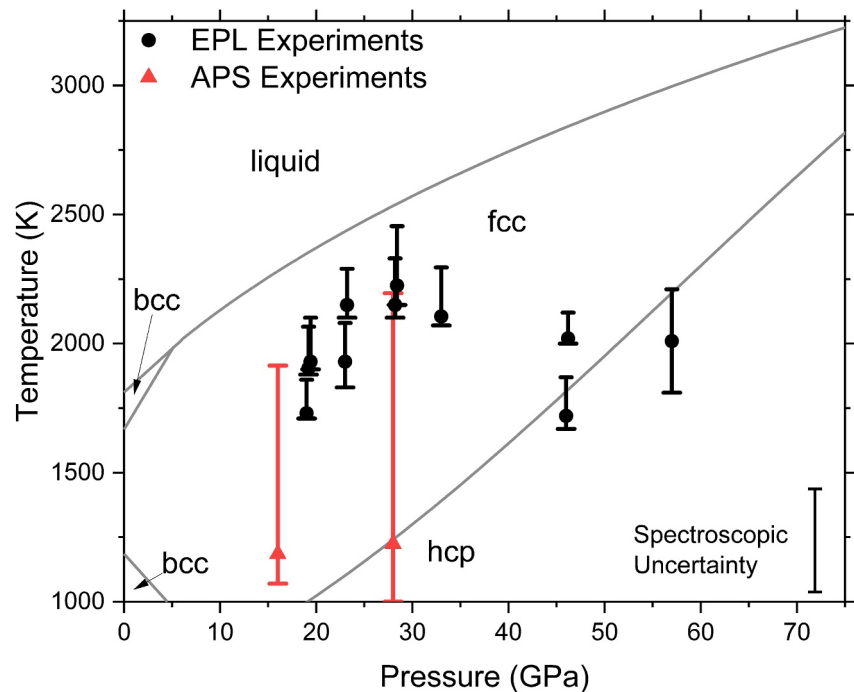


Figure 2. Pressure-Temperature phase diagram of iron, alongside P-T conditions for thermal conductivity measurements reported here. Black circles indicate EPL experiments, while red triangles indicate APS experiments as reported in Table 1. Phase lines are based on recent DAC literature (Dewaele & Garbarino, 2017; Morard, Boccato, et al., 2018).

temperature dependence. Figure 3 shows the results of the present study as a function of temperature at pressures around 25 GPa along with k and k_{el} inferred from resistivity experiments. Our experiments yield k values that are consistent with the lower end of reported k_{el} values at similar P-T conditions (Ohta et al., 2016; Yong et al., 2019), and higher than recent k measurements for fcc iron at similar P-T conditions (Andraut et al., 2025). Agreement with Ohta et al. (2016) at the P-T conditions of our measurements may be due to cancellation of errors, as the temperature dependence of k_{el} in that study is very different from what is reported in multi-anvil k_{el} studies at similar pressures (Ezenwa & Yoshino, 2021; Yong et al., 2019). We report higher k values than other direct measurements, which is likely due to the use of white light interference to measure in situ sample thickness instead of *ex situ* measurements of recovered samples, as well as differences in model parametrization. Our fitted k_{KCl} values are 4–14 times higher than the fixed value employed in Andraut et al. (2025), we adopt heat capacity values for fcc-Fe that are 30% higher and Andraut et al. (2025) assumes the ambient pressure density of iron for fcc measurements, which is 9% lower than the density of fcc iron at 25 GPa and 2000 K (Komabayashi, 2014). Most of the discrepancy likely comes from the parametrization of iron density and heat capacity; k_{KCl} strongly trades off against the pulsed laser spot size, which is fixed to its measured value in the present study and varied as a free parameter in Andraut et al. (2025). Changes in k with temperature are small relative to experimental uncertainties ($\sim 18\%$ – 35%) at the conditions investigated here, which is consistent with ambient pressure literature where k varies by $\sim 10\%$ between 1200 and 1800 K (Ho et al., 1972).

Thermal conductivity values from the present study are plotted as a function of pressure in Figure 4. While k values in the present study are systematically higher than Konôpková et al. (2016) at comparable pressures, they are within mutual uncertainties. Discrepancies likely relate to improvements in modeling the shape of the heat pulse, the use of literature thermodynamic models to determine heat capacity at high P-T conditions and employing direct determination of Fe and KCl density at or near measured P-T conditions. Data on hcp iron from Konôpková et al. (2016) has been omitted for clarity, as the primary focus of the present study is iron in the fcc structure. At the studied P-T conditions, changes in thermal conductivity with temperature are not resolvable and so the data has been fit with a linear relation $k(P)_{HT} = P(dk/dP)_{HT} + k_{0,HT}$. Fitted parameters to this equation are $(dk/dP)_{HT} = 0.51(5)$ and $k_{0,HT} = 32(3)$. Including or omitting mixed phase data at 40–60 GPa does not significantly change the fit. Available thermal conductivity and electrical conductivity data show that the temperature variation of k for fcc-Fe

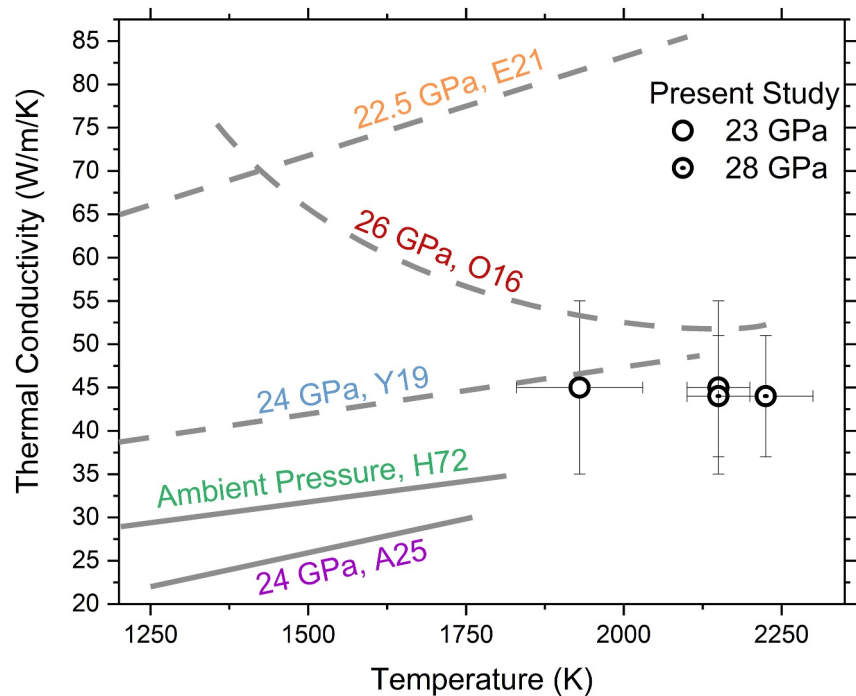


Figure 3. Thermal conductivity of fcc-Fe versus temperature. Dashed lines indicate k_{el} values inferred from electrical studies at pressures near 25 GPa (E21 (Ezenwa & Yoshino, 2021), O16 (Ohta et al., 2016), Y19 (Yong et al., 2019)). These values are inferred from Equation 1 and a Lorenz number of $2.445 \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$. Solid lines indicate k at ambient pressure for solid fcc and bcc iron (H72 (Ho et al., 1972)) and k at 24 GPa for fcc iron (A25 (Andrault et al., 2025)). Direct measurements of k from the present study are reported as open circles (23 GPa) and dotted circles (28 GPa).

is only a few W/m/K over its stability field (Ho et al., 1972; Yong et al., 2019), which is not resolvable with the present technique, and there is no 300 K reference data for fcc-Fe as it is not thermodynamically stable at this temperature. The present results indicate that the thermal conductivity of fcc iron is weakly dependent on pressure at the thermodynamic conditions of Mercury's core, 5–40 GPa and 1600–2400 K (e.g., Edmund et al., 2022) in contrast with earlier studies where the thermal conductivity of iron was reported to increase from 32 W/m/K at ambient pressure to 65 W/m/K at 5 GPa and high temperature (Deng et al., 2013; Pommier, 2018).

4. Discussion

4.1. Comparison to Previous Literature and the Lorenz Number of Iron

The thermal conductivity values obtained in the present study are in good agreement with thermal conductivity values inferred from electrical resistivity experiments on fcc-Fe at comparable pressures and temperatures (Ohta et al., 2016; Yong et al., 2019). This finding suggests that the Wiedemann-Franz law provides a reasonable relationship between the thermal and electrical conductivity of fcc-Fe at high temperatures and pressures. It is also consistent with a relatively small phonon contribution, although the comparison alone does not uniquely constrain the magnitude of the lattice contribution. At around 18–20 GPa and 1800–2000 K, comparison with a recent multi-anvil study gives an apparent Lorenz number of about $2.3(3) \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$ (Yong et al., 2019). Within uncertainties, this is consistent with the ambient pressure behavior of fcc-Fe where it is proposed that $L = 2.1 \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$ and k_{ph} is about 10% of the total thermal conductivity (Secco, 2017). This value of the Lorenz number is also close to the ideal value ($2.445 \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$) and appears to describe well the behavior of fcc-Fe over the entire pressure range relevant to the cores of Mercury and Mars. Ab initio calculations of the thermal conductivity of liquid Fe have proposed that a large change in transport properties occurs between 20 and 50 GPa which coincides with the transition from a paramagnetic to a diamagnetic state (Korell et al., 2019). Such a transition was proposed to occur experimentally in solid and liquid iron at around 18 GPa based on electrical resistivity measurements (Ezenwa & Yoshino, 2021), however such an effect is not observed in the present data

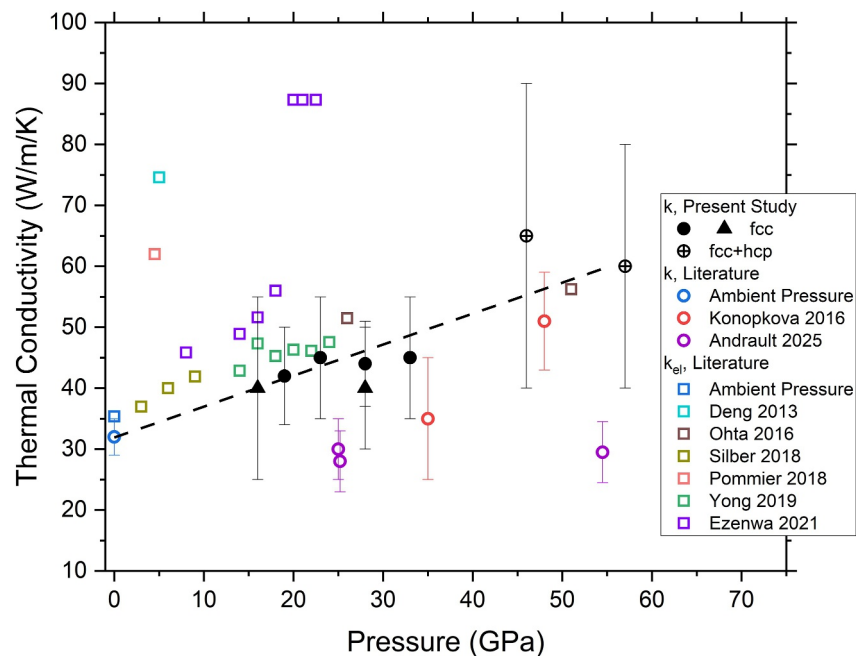


Figure 4. Thermal conductivity of fcc-Fe as a function of pressure is presented in this study and compared with previous thermal conductivity measurements (Andraut et al., 2025; Konopková et al., 2016), as well as with estimates of electronic thermal conductivity (k_{el}) from the literature (Deng et al., 2013; Ezenwa & Yoshino, 2021; Ohta et al., 2016; Pommier, 2018; Silber et al., 2018; Yong et al., 2019). All values shown are from data on solid iron between 1700 and 2300 K, the temperature range studied in the present work. Black filled and crossed circles indicate EPL experiments, black filled triangles indicate APS experiments. Open squares indicate literature k_{el} studies, open circles indicate literature k studies. Black dashed line is a linear fit to our data and ambient pressure literature. Electrical resistivity data has been converted to k_{el} values using the reported temperature and the Sommerfeld value for the Lorenz number ($L = 2.445 \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$).

set within the uncertainty of the measurements. If the magnetic transition in fcc-Fe has a strong influence on its thermal transport properties, it would have to occur beyond the central pressures of Mercury and Mars (>40 GPa).

It is striking that an agreement can be found across the present study and some electrical literature for fcc-Fe (Ohta et al., 2016; Yong et al., 2019), given the large spread of values at higher pressures. At Earth's core pressures (>135 GPa), different reports vary by nearly an order of magnitude (Konopková et al., 2016; Ohta et al., 2016). Different mechanisms have been suggested to explain the discrepancy; a significant factor given the agreement at low pressures (<60 GPa) is likely the precision of available techniques for the measurement of sample thickness. Estimated electronic thermal conductivity values vary proportionally to sample thickness (l) with l^{-1} in electrical resistivity experiments (Gomi et al., 2013), while it varies between $l^{1.5}$ to l^2 in thermal conductivity experiments (Geballe et al., 2020; Yagi et al., 2010) such that a systematic offset in thickness will cause both measurements to diverge in reported thermal conductivity values. On both compression and decompression the sample chamber is observed to thin along the axis of compression due to the deformation of the sample chamber, which in turn would influence the shape of the sample depending on the experimental design (Konopková et al., 2016; Lobanov & Geballe, 2022). This results in a systematic bias of thickness measurements using *ex situ* methods on recovered samples toward a thinner sample. Being a direct measurement of thickness between different interfaces in the DAC, the WL technique can overestimate or underestimate foil thickness depending on the configuration of the sample chamber and the variation of cavity thickness between measurement points. At pressures exceeding ~80 GPa where diamond anvils themselves deform plastically (Li et al., 2018), the simple assumptions employed to determine sample thickness are likely violated; direct measurements of sample thickness at extreme pressures are needed in order to experimentally address the thermal conductivity of iron at the conditions of Earth's core and beyond.

4.2. Thermal Conductivity of Mercury's Core

In order to assess how the present results on iron influence the thermal conductivity of Mercury's core, we provide simple compositional models for the core based on the iron-silicon-sulfur system. These compositions are

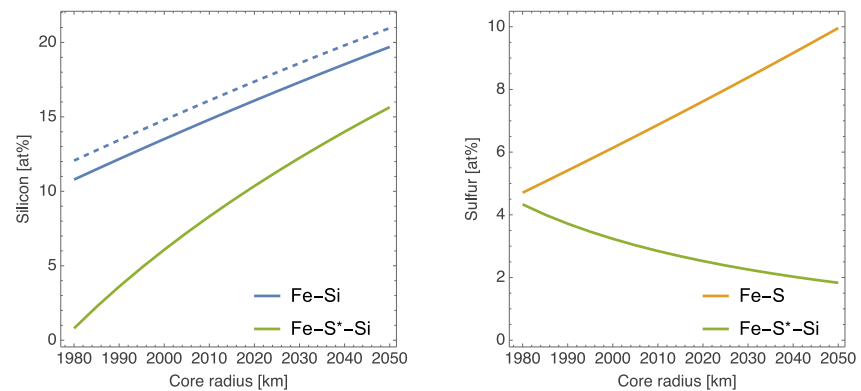


Figure 5. Left: Silicon content (at%) of Mercury's core versus core radius. Right: Sulfur content (at%) of Mercury's core versus core radius. In all panels, blue, green and gold lines indicate different compositional models for Mercury's core. Blue, green and gold lines indicate iron-silicon, iron-sulfur-silicon and iron-sulfur compositional models respectively. For ternary compositions, silicon and sulfur contents from both plots are combined to give the bulk core composition. Solid lines indicate core compositions calculated when the core-side core-mantle boundary (CMB) temperature is 1750 K, while the dashed line indicates core compositions where CMB temperature is 1850 K.

computed from an estimate of core temperature, core mass and an equation of state for the alloy of interest. These calculations assume an adiabatic temperature gradient within the core and a Core-Mantle Boundary (CMB) temperature of 1750 K or 1850 K in Figure 5. The thermodynamic properties of Mercury's core are computed using thermal equations of state for liquid iron-silicon-sulfur alloys (Terasaki et al., 2019). Core mass is determined from the bulk density of Mercury, after subtracting the mass and volume of the mantle. The mantle is assumed to be composed of 80 vol% enstatite and 10 vol% forsterite and 10 vol% diopside, with elastic properties computed following Stixrude and Lithgow-Bertelloni (2024). The temperature change across the CMB is assumed to be 250 K, below which the mantle adopts an adiabatic temperature gradient, and then decreases linearly to the surface temperature (350 K) near the surface (Figure S8 in Supporting Information S1). The radius of the core is presently constrained to be 2010(40) km (cf. Steinbrügge et al., 2021). Figure 5 shows the range of bulk core compositions in the iron-silicon-sulfur system which satisfy constraints on the core mass of Mercury, as a function of core radius. As there are degenerate solutions in the iron-silicon-sulfur ternary system when relating composition to core density and temperature, coupled sulfur and silicon abundances are additionally constrained using geochemical considerations (Boujibar et al., 2019). Bulk core compositions calculated here are comparable with the results of studies using geodetic data to constrain the interior structure (Genova et al., 2019; Steinbrügge et al., 2021).

The thermal conductivity of fcc iron provides an upper bound on the thermal conductivity of the core-forming alloys found in Mercury. While the outermost core of Mercury is liquid, it is expected that the thermal conductivity of liquid iron is only 5%–10% lower than that of the solid in this pressure range (Pommier, 2018; Yong et al., 2019). The effect of impurity elements is more significant; recent results suggest a 30% reduction in thermal conductivity for silicon-bearing alloys compared to our values (Berrada et al., 2021; Hsieh et al., 2020), and 50%–70% for sulfur-bearing alloys (Edmund, Bi, et al., 2024; Hsieh, Deschamps, et al., 2024). While carbon can also be present in the core of Mercury (Byrne et al., 2014; Vander Kaaden & McCubbin, 2015), the effect of carbon on the thermal conductivity of solid or liquid iron alloys at high pressures and temperatures is not known.

In the absence of a more detailed microscopic description of the alloying mechanisms of core-forming alloys, a first-order estimate of the thermal conductivity of the alloy can be made by adding the thermal conductivity of the end-members in series or in parallel (Kingery, 1959). Adding conductivities in series provides an upper bound, $k_{\text{alloy}} = x \cdot k_{\text{LE}} + (1 - x) \cdot k_{\text{Fe}}$ where alloy denotes the alloy composition of interest, LE denotes the light-element composition chosen as the end-member, and x is the mol fraction of the light element. A lower bound can be provided by adding thermal conductivities in parallel as $r_{\text{alloy}} = x \cdot r_{\text{LE}} + (1 - x) \cdot r_{\text{Fe}}$ where $r = 1/k$ is the thermal resistance of the material, which is equivalent to Matthiessen's rule. The thermal conductivity of Mercury-relevant alloys are shown in Figure 6. The choice of averaging scheme does not significantly influence interpolated thermal conductivity values. The variation of the bulk core conductivity is bracketed between about 23 W/m/K and 38 W/m/K at 25 GPa, and most strongly influenced by sulfur alloying. As shown in Figure 4 the thermal

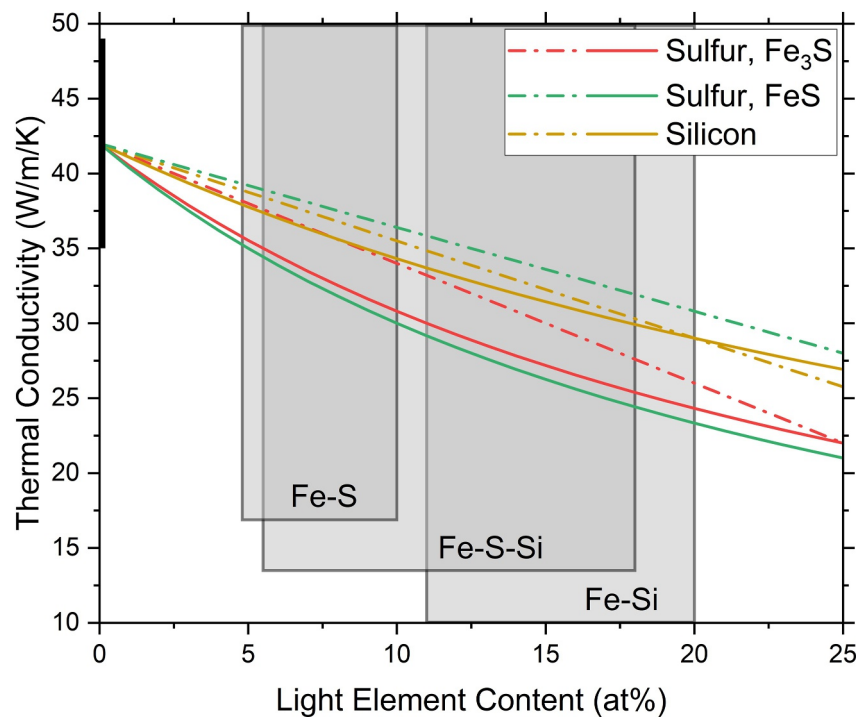


Figure 6. Modeled thermal conductivity values of iron alloys versus light element content. Values are based on temperatures of ~ 2000 K and pressures between 20 and 30 GPa. Dash-dotted lines indicate models where k values are interpolated in series, solid lines are interpolated in parallel. Red and Green lines indicate models using either Fe_3S (red) or FeS (green) as end-member compositions. Gray shaded areas indicate the range of bulk core compositions which satisfy constraints on core density and core radius. For model details see text.

conductivity of iron varies by about 40% between the pressure of the core-mantle boundary and the central pressure of the core. However, the alloying of light elements to iron tends to reduce both the pressure and temperature dependence of thermal conductivity so the pressure effect on iron can be used as an upper bound for the amount by which k varies within Mercury's core (Hsieh, Deschamps, et al., 2024; Hsieh et al., 2020; Zhang et al., 2022).

In Figure 6, the effect of sulfur on core conductivity is stronger than the effect of silicon. However, smaller concentrations of sulfur are needed to match the density of the core; sulfur has a more significant effect on the compressibility of iron alloys when compared to silicon (Morard, Bouchet, et al., 2018; Terasaki et al., 2019). Silicon and sulfur abundances are anti-correlated which favors ternary compositions where silicon is the dominant light element (Figure 5). A silicon-rich core is more compatible with geochemical constraints on core-mantle equilibration at reducing conditions than a purely iron-sulfur core (Cartier et al., 2020; Namur et al., 2016), favoring silicon-rich and sulfur-depleted core compositions. The geochemical coupling between sulfur and silicon abundances is a consequence of the lithophile nature of sulfur at highly reducing conditions relevant to Mercury's interior (Namur et al., 2016). The effect of both silicon and sulfur on core conductivity for realistic core compositions is similar due to tradeoffs in composition and conductivity. Iron-silicon-sulfur liquids which are sufficiently rich in silicon and sulfur can form immiscible sulfur-rich and silicon-rich liquids (Morard & Katsura, 2010), which would substantially alter the thermal conductivity of the core near the core mantle boundary due to the large change in composition. However, bulk sulfur concentrations calculated here are likely too low for chemical stratification to occur. Sulfur-rich compositions could additionally alter the transport properties of the core due to changes in the ionic structure of the liquid (Alfè & Gillan, 1998; Wagle et al., 2019), but this process requires compositions which are disfavored by current geodetic and geochemical constraints.

4.3. Thermal Evolution of Mercury's Core

In this section we illustrate how thermal conductivity and core compositions resulting from core radius variations affect necessary conditions for a dynamo to occur, as well as how thermal conductivity and core composition

determine the present-day thickness of the stable thermal layer and inner core radius. To model the evolution of the stable layer above the convecting core we use the approach and computer code detailed in Greenwood et al. (2021). The core model is coupled to the parametrized mantle thermal evolution code TEMPURA (Bau-meister et al., 2023). Mantle mineralogy and initial thermal state are the same as in Section 4.2. The abundances of heat producing elements are taken from Peplowski et al. (2012) and the mantle solidus and liquidus are from Namur et al. (2016). We set the mantle reference viscosity to 10^{22} Pa s and we compute the thermal conductivity of the mantle aggregate from its composition (Guo et al., 2024; Zhang et al., 2019). For the crust we assume a thermal conductivity of 3 W/m/K and a crustal enrichment factor of 20 (Tosi et al., 2013). Since there is little data on the phase equilibria for ternary iron alloy compositions shown previously in Figure 5 we restrict our investigations to the binary iron-silicon system. To parametrize the thermal evolution models, constraints on the solidus and liquidus temperatures of iron-silicon alloys are needed, which are taken from literature studies on the melting relations and eutectic composition of iron-silicon alloys (Edmund et al., 2022). The results of Edmund et al. (2022) agree with recent data from large volume press experiments as shown in Figure S9 of Supporting Information S1 (Ezenwa et al., 2024). Solidus temperatures are assumed to vary linearly between iron and the iron-silicon solidus curve of Edmund et al. (2022), at an eutectic silicon concentration of 12.5 wt%. In our models, we assume that solidus and liquidus temperatures are the same, as melting loops in the iron-silicon system are on the order of 10–20 K over a wide compositional range (Meco & Napolitano, 2005). Following the previous section, all required thermodynamic properties of the liquid iron-silicon alloy are computed from Terasaki et al. (2019). For the solid inner core we use the bcc equation of state from Edmund et al. (2022). The melting entropy of the core alloy is calculated from 1 bar data (Cui & Jung, 2017) and scaled to high pressure using the equations of state of bcc Fe and liquid Fe (Dorogokupets et al., 2017). As in the previous section, we compute an interior structure model for each core radius value. The resulting thermodynamic properties are then averaged over the mantle and core volume and used for the subsequent thermal evolution simulation. We vary the thermal conductivity of the core between 25 W/m/K and 40 W/m/K to represent the compositional variation shown in Figure 6. While the thermal conductivity of iron increases with pressure, silicon-rich iron alloys show more complex behavior which results in a smaller pressure dependence than observed for iron over most of the core (Berrada et al., 2021; Hsieh et al., 2020). This weak pressure dependence justifies the use of a constant thermal conductivity value to describe the core, as shown in Figure S10 of Supporting Information S1.

The primary energy sources to drive dynamo action in Mercury's iron-silicon core are the secular cooling of the core and the latent heat released due to inner core nucleation. Some degree of either compositional stratification or thermal stratification is likely required to explain the unique characteristics of Mercury's magnetic field (e.g., Kolhey et al., 2025). The ability for thermal evolution models to produce viable core thermal structures for dynamo action is strongly linked to the thermal conductivity of Mercury's core. Following Davies et al. (2024), we require a dynamo entropy above 1 MW/K at present-day for a model to be viable. If core thermal conductivity is increased, it results in faster cooling of the core, an earlier solidification of the inner core and a larger present-day inner core radius (Figure 7a). The models shown here have compositions which are consistent with Mercury's core density by construction, and models with viable dynamos at the present-day are found for core conductivity values consistent with existing experimental constraints (Figure 6 and S10). Figure 7 shows the inner core radius and thermally stratified layer thickness at present day from thermal evolution models run for 4.5 billion years. Successful models are defined as those which exhibit both thermally stratified and convecting liquid layers in the outer core. While these models show viable thermal structures for dynamo generation in the core in the present day, crustal remanence studies point to early periods where Mercury may have had a magnetic field, around 3.8 billion years ago (Johnson et al., 2015). Our thermal evolution models do not provide features which may explain an early but long-lived stage of interior cooling as suggested by crustal magnetic remanence studies (Johnson et al., 2015). Viable dynamo models reported here exhibit timescales for inner core growth of at most 2.1 billion years from the present day. An early thermally driven dynamo would last at most for 250 million years after formation of the planet in our models. Our results are highly sensitive to the liquidus phase relations and thermal conductivity of the core-forming material, and likely the answer to simultaneously addressing past and present constraints on Mercury's dynamo requires stronger constraints on both phase relations and thermal conductivity values for a wide range of compositions relevant to Mercury's core. Additionally, further constraints are needed on the properties that govern the thermal evolution of Mercury's mantle. In terms of phase relations, both carbon and sulfur exhibit complex partitioning behavior in iron which can change the energetics of long-term core thermochemical evolution (Tao & Fei, 2021; Vander Kaaden et al., 2020), and also potentially lead to changing

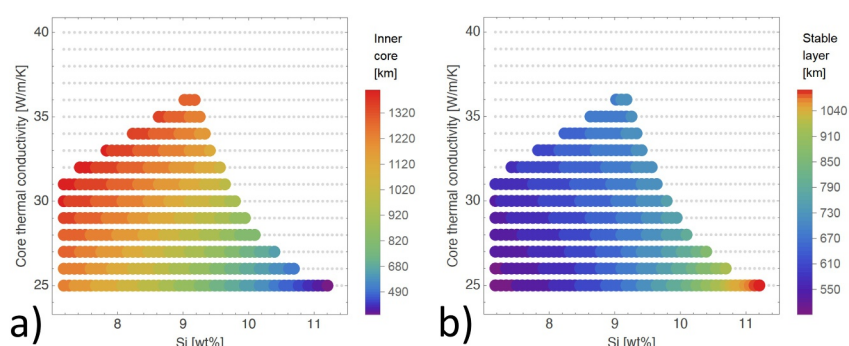


Figure 7. (a) Present day inner core size from thermal evolution models as a function of core silicon content and thermal conductivity. (b) Stable layer thickness from thermal evolution models as a function of core silicon content and thermal conductivity. For both figures, gray points indicate combinations of parameters which did not result in a core thermal structure where the outer core was still convecting after 4.5 billion years.

compositions of the inner and outer core during inner core growth. Interestingly, non-linear interactions between sulfur and silicon have been shown to strongly fractionate them both between solid and liquid (Tao & Fei, 2021). While the similar atomic mass of silicon and sulfur leads to only a minor change in chemical buoyancy during inner core growth when considering solid-liquid partitioning of both elements, progressive enrichment of sulfur in the liquid outer core would have a strong effect on the thermal conductivity of the core, and progressively reduce the growth rate of the inner core by a reduction in core conductivity. Nevertheless, further work is needed to investigate the thermal conductivity and phase relations in Mercury-relevant iron-alloy systems.

5. Conclusions

In conclusion, the total thermal conductivity of fcc-Fe has been measured between 16 and 60 GPa, and at temperatures between 1000 and 2500 K. Comparisons with literature electrical studies of fcc-Fe suggest that the Wiedemann-Franz law provides a reasonable description of the relationship between the thermal and electrical conductivity of fcc-Fe at high pressures and temperatures, and that the Lorenz number of fcc-Fe is likely close to the Sommerfeld value ($L = 2.445 \cdot 10^{-8} \text{ W} \cdot \Omega \cdot \text{K}^{-2}$) at Mercury's core conditions. At the central pressures of Mercury (35–40 GPa), we obtain a thermal conductivity value of 52(7) W/m/K for fcc-Fe. Models which couple the bulk core composition of Mercury and the effect of alloying on core conductivity show that the core conductivity is, at most, 40% lower than the conductivity values for pure iron reported here. Our thermal evolution models show that this range of thermal conductivity values are consistent with an active dynamo in the presence of thermally stratified layer.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

A description of experiments and model parameters is found in found in Data set S1 of Supporting Information S1. Data used in this study as well as an example flexPDE model file is deposited in a Zenodo repository (Edmund et al., 2026).

References

- Akahama, Y., & Kawamura, H. (2006). Pressure calibration of diamond anvil raman gauge to 310GPa. *Journal of Applied Physics*, 100(4), 043516. <https://doi.org/10.1063/1.2335683>
- Alfè, D., & Gillan, M. J. (1998). First-principles simulations of liquid Fe-S under Earth's core conditions. *Physical Review B: Condensed Matter*, 58(13), 8248–8256. <https://doi.org/10.1103/PhysRevB.58.8248>
- Anderson, B. J., Johnson, C. L., Korth, H., Purucker, M. E., Winslow, R. M., Slavin, J. A., et al. (2011). The global magnetic field of Mercury from MESSENGER orbital observations. *Science*, 333(6051), 1859–1862. <https://doi.org/10.1126/science.1211001>
- Andraut, D., Pison Pacynski, L., Monteux, J., Gardès, E., & Mathieu, A. (2025). Long-lived magnetic field in Earth-like terrestrial planets. *Physics of the Earth and Planetary Interiors*, 360, 107315. <https://doi.org/10.1016/j.pepi.2025.107315>

Acknowledgments

I. Ezenwa is acknowledged for fruitful discussions. Many thanks to S. Vitale for technical assistance with Focused Ion Beam milling. Portions of this work were performed at GeoSoilEnviroCARS (The University of Chicago, Sector 13), Advanced Photon Source (APS), Argonne National Laboratory. GeoSoilEnviroCARS is supported by the National Science Foundation—Earth Sciences (EAR-1634415). This research used resources of the Advanced Photon Source, a U.S. Department of Energy (DOE) Office of Science User Facility operated for the DOE Office of Science by Argonne National Laboratory under Contract No. DE-AC02-06CH11357. This work was supported by Grant NSF-EAR 2049127 to A. F. G. The work of AR was financially supported by the Belgian PRODEX program managed by the European Space Agency in collaboration with the Belgian Federal Science Policy Office, contract numbers PEA4000140326 Planet Interior of the European Space Agency. Open Access funding enabled and organized by Projekt DEAL.

- Baumeister, P., Tosi, N., Brachmann, C., Grenfell, J. L., & Noack, L. (2023). Redox state and interior structure control on the long-term habitability of stagnant-lid planets. *AA*, 675, A122. <https://doi.org/10.1051/0004-6361/202245791>
- Berrada, M., Secco, R. A., & Yong, W. (2021). Adiabatic heat flow in Mercury's core from electrical resistivity measurements of liquid Fe-8.5 wt%Si to 24 GPa. *Earth and Planetary Science Letters*, 568, 117053. <https://doi.org/10.1016/j.epsl.2021.117053>
- Boujibar, A., Habermann, M., Richter, K., Ross, D. K., Pando, K., Richter, M., et al. (2019). U, Th, and K partitioning between metal, silicate, and sulfide and implications for Mercury's structure, volatile content, and radioactive heat production. *American Mineralogist*, 104(9), 1221–1237. <https://doi.org/10.2138/am-2019-7000>
- Byrne, P. K., Klimczak, C., Şengör, A. M. C., Solomon, S. C., Watters, T. R., & Hauck, S. A. (2014). Mercury's global contraction much greater than earlier estimates. *Nature Geoscience*, 7(4), 301–307. <https://doi.org/10.1038/ngeo2097>
- Campbell, A. J., Danielson, L., Richter, K., Seagle, C. T., Wang, Y., & Prakapenka, V. B. (2009). High pressure effects on the iron-iron oxide and nickel-nickel oxide oxygen fugacity buffers. *Earth and Planetary Science Letters*, 286(3), 556–564. <https://doi.org/10.1016/j.epsl.2009.07.022>
- Cartier, C., Namur, O., Nittler, L., Weider, S., Crapster-Pregont, E., Vorburger, A., et al. (2020). No FeS layer in Mercury? Evidence from Ti/Al measured by MESSENGER. *Earth and Planetary Science Letters*, 534, 116108. <https://doi.org/10.1016/j.epsl.2020.116108>
- Chabot, N. L., Wollack, E. A., Klima, R. L., & Minitti, M. E. (2014). Experimental constraints on Mercury's core composition. *Earth and Planetary Science Letters*, 390, 199–208. <https://doi.org/10.1016/j.epsl.2014.01.004>
- Cui, S., & Jung, I.-H. (2017). Critical reassessment of the Fe-Si system. *Calphad*, 56, 108–125. <https://doi.org/10.1016/j.calphad.2016.11.003>
- Davies, C. J., Pommier, A., Greenwood, S., & Wilson, A. (2024). Thermal and magnetic evolution of Mercury with a layered Fe-Si(-S) core. *Earth and Planetary Science Letters*, 641, 118812. <https://doi.org/10.1016/j.epsl.2024.118812>
- Deng, L., Seagle, C., Fei, Y., & Shahar, A. (2013). High pressure and temperature electrical resistivity of iron and implications for planetary cores. *Geophysical Research Letters*, 40(1), 33–37. <https://doi.org/10.1029/2012GL054347>
- Desai, P. D. (1986). Thermodynamic properties of iron and silicon. *Journal of Physical and Chemical Reference Data*, 15(3), 967–983. <https://doi.org/10.1063/1.555761>
- Dewaele, A., Belonoshko, A. B., Garbarino, G., Occelli, F., Bouvier, P., Hanfland, M., & Mezouar, M. (2012). High-pressure–high-temperature equation of state of KCl and KBr. *Physical Review B: Condensed Matter*, 85(21), 214105. <https://doi.org/10.1103/PhysRevB.85.214105>
- Dewaele, A., & Garbarino, G. (2017). Low temperature equation of state of iron. *Applied Physics Letters*, 111(2), 021903. <https://doi.org/10.1063/1.4989688>
- Dorogokupets, P. I., Dymshits, A. M., Litasov, K. D., & Sokolova, T. S. (2017). Thermodynamics and equations of state of iron to 350 GPa and 6000 K. *Scientific Reports*, 7(1), 41863. <https://doi.org/10.1038/srep41863>
- Driscoll, P. E., & Du, Z. (2019). Geodynamo conductivity limits. *Geophysical Research Letters*, 46(14), 7982–7989. <https://doi.org/10.1029/2019gl082915>
- Dumberry, M., & Rivoldini, A. (2015). Mercury's inner core size and core-crystallization regime. *Icarus*, 248, 254–268. <https://doi.org/10.1016/j.icarus.2014.10.038>
- Edmund, E., Bi, T., Geballe, Z., Brugman, K., Lin, J.-F., Chariton, S., et al. (2024). Structure and transport properties of FeS at planetary core conditions. *Earth and Planetary Science Letters*, 646, 118959. <https://doi.org/10.1016/j.epsl.2024.118959>
- Edmund, E., Chuvashova, I., Konopkova, Z., Husband, R., Strohm, C., Appel, K., et al. (2024). The thermal conductivity of bridgmanite at lower mantle conditions using a multi-technique approach. *Journal of Geophysical Research: Solid Earth*, 129(6), e2024JB028823. <https://doi.org/10.1029/2024jb028823>
- Edmund, E., Geballe, Z. M., Rivoldini, A., Tosi, N., Chariton, S., Prakapenka, V. B., & Goncharov, A. F. (2026). Temperature histories and model file [dataset]. *Zenodo*. <https://doi.org/10.5281/zenodo.22212574>
- Edmund, E., Morard, G., Baron, M. A., Rivoldini, A., Yokoo, S., Boccato, S., et al. (2022). The Fe-FeSi phase diagram at Mercury's core conditions. *Nature Communications*, 13(1), 387. <https://doi.org/10.1038/s41467-022-27991-9>
- Ezenwa, I. C., & Fei, Y. (2023). High pressure melting curve of Fe determined by inter-metallic fast diffusion technique. *Geophysical Research Letters*, 50(6), e2022GL102006. <https://doi.org/10.1029/2022GL102006>
- Ezenwa, I. C., Fei, Y., Hrubak, R., & Kenney-Benson, C. (2024). High pressure melting curve of Fe-Si: Implication for the thermal properties in Mercury's core. *Journal of Geophysical Research: Planets*, 129(12), e2024JE008353. <https://doi.org/10.1029/2024JE008353>
- Ezenwa, I. C., & Yoshino, T. (2021). Martian core heat flux: Electrical resistivity and thermal conductivity of liquid Fe at Martian core P-T conditions. *Icarus*, 360, 114367. <https://doi.org/10.1016/j.icarus.2021.114367>
- Geballe, Z. M., Sime, N., Badro, J., van Keken, P. E., & Goncharov, A. F. (2020). Thermal conductivity near the bottom of the earth's lower mantle: Measurements of pyrolite up to 120 GPa and 2500 K. *Earth and Planetary Science Letters*, 536, 116161. <https://doi.org/10.1016/j.epsl.2020.116161>
- Genova, A., Goossens, S., Mazarico, E., Lemoine, F. G., Neumann, G. A., Kuang, W., et al. (2019). Geodetic evidence that Mercury has a solid inner core. *Geophysical Research Letters*, 46(7), 3625–3633. <https://doi.org/10.1029/2018GL081135>
- Gomi, H., Ohta, K., Hirose, K., Labrosse, S., Caracas, R., Verstraete, M. J., & Hernlund, J. W. (2013). The high conductivity of iron and thermal evolution of the earth's core. *Physics of the Earth and Planetary Interiors*, 224, 88–103. <https://doi.org/10.1016/j.pepi.2013.07.010>
- Greenwood, S., Davies, C. J., & Pommier, A. (2021). Influence of thermal stratification on the structure and evolution of the martian core. *Geophysical Research Letters*, 48(22), e2021GL095198. <https://doi.org/10.1029/2021GL095198>
- Guo, X., Feng, B., Zhang, B., Zhai, S., Xue, W., Song, Y., & Yan, X. (2024). Effect of iron content on the thermal conductivity and thermal diffusivity of orthopyroxene. *Geochemistry, Geophysics, Geosystems*, 25(6), e2023GC011419. <https://doi.org/10.1029/2023GC011419>
- Hasegawa, A., Ohta, K., Yagi, T., Hirose, K., & Yamashita, Y. (2024). Inversion of the temperature dependence of thermal conductivity of hcp iron under high pressure. *Scientific Reports*, 14(1), 23582. <https://doi.org/10.1038/s41598-024-74110-3>
- Hauck, S. A., Margot, J.-L., Solomon, S. C., Phillips, R. J., Johnson, C. L., Lemoine, F. G., et al. (2013). The curious case of Mercury's internal structure. *Journal of Geophysical Research*, 118(6), 1204–1220. <https://doi.org/10.1002/jgre.20091>
- Ho, C. Y., Powell, R. W., & Liley, P. E. (1972). Thermal conductivity of the elements. *Journal of Physical and Chemical Reference Data*, 1(2), 279–421. <https://doi.org/10.1063/1.3253100>
- Hsieh, W.-P., Deschamps, F., Tsao, Y.-C., Yoshino, T., & Lin, J.-F. (2024). A thermally conductive martian core and implications for its dynamo cessation. *Science Advances*, 10(12), eadk1087. <https://doi.org/10.1126/sciadv.adk1087>
- Hsieh, W.-P., Goncharov, A. F., Labrosse, S., Holtgrewe, N., Lobanov, S. S., Chuvashova, I., et al. (2020). Low thermal conductivity of iron-silicon alloys at earth's core conditions with implications for the geodynamo. *Nature Communications*, 11(1), 3332. <https://doi.org/10.1038/s41467-020-17106-7>
- Johannsen, P. G., Reib, G., Bohle, U., Magiera, J., Müller, R., Spiekermann, H., & Holzapfel, W. B. (1997). Refractive index of the alkali halides. ii. effect of pressure on the refractive index of 11 alkali halides. *Physical Review B: Condensed Matter*, 55, 6865–6870. <https://doi.org/10.1103/PhysRevB.55.6865>

- Johnson, C. L., Phillips, R. J., Purucker, M. E., Anderson, B. J., Byrne, P. K., Denevi, B. W., et al. (2015). Low-altitude magnetic field measurements by MESSENGER reveal Mercury's ancient crustal field. *Science*, 348(6237), 892–895. <https://doi.org/10.1126/science.aaa8720>
- Johnson, C. L., Philpott, L. C., Anderson, B. J., Korth, H., Hauck II, S. A., Heyner, D., et al. (2016). MESSENGER observations of induced magnetic fields in Mercury's core. *Geophysical Research Letters*, 43(6), 2436–2444. <https://doi.org/10.1002/2015GL067370>
- Kingery, W. D. (1959). Thermal conductivity: XIV, conductivity of multicomponent systems. *Journal of the American Ceramic Society*, 42(12), 617–627. <https://doi.org/10.1111/j.1151-2916.1959.tb13583.x>
- Knibbe, J. S., Rivoldini, A., Zhao, Y., & Van Hoolst, T. (2025). On the thermal evolution and magnetic field generation of planet Mercury. *Physics of the Earth and Planetary Interiors*, 363, 107348. <https://doi.org/10.1016/j.pepi.2025.107348>
- Kolhey, P., Heyner, D., Wicht, J., Gastine, T., Glassmeier, K.-H., & Plaschke, F. (2025). Dynamo models with a Mercury-like magnetic offset dipole. *Journal of Geophysical Research: Planets*, 130(3), e2024JE008660. <https://doi.org/10.1029/2024JE008660>
- Komabayashi, T. (2014). Thermodynamics of melting relations in the system Fe-FeO at high pressure: Implications for oxygen in the Earth's core. *Journal of Geophysical Research: Solid Earth*, 119(5), 4164–4177. <https://doi.org/10.1002/2014jb010980>
- Konôpková, Z., McWilliams, R. S., Gómez-Pérez, N., & Goncharov, A. F. (2016). Direct measurement of thermal conductivity in solid iron at planetary core conditions. *Nature*, 534(7605), 99–101. <https://doi.org/10.1038/nature18009>
- Korell, J.-A., French, M., Steinle-Neumann, G., & Redmer, R. (2019). Paramagnetic-to-diamagnetic transition in dense liquid iron and its influence on electronic transport properties. *Physical Review Letters*, 122(8), 086601. <https://doi.org/10.1103/PhysRevLett.122.086601>
- Li, Y., Vočadlo, L., & Brodholt, J. P. (2018). The elastic properties of hcp-Fe alloys under the conditions of the Earth's inner core. *Earth and Planetary Science Letters*, 493, 118–127. <https://doi.org/10.1016/j.epsl.2018.04.013>
- Lobanov, S. S., & Geballe, Z. M. (2022). Non-isotropic contraction and expansion of samples in diamond anvil cells: Implications for thermal conductivity at the core-mantle boundary. *Geophysical Research Letters*, 49(19), e2022GL100379. <https://doi.org/10.1029/2022GL100379>
- McWilliams, R. S., Konôpková, Z., & Goncharov, A. F. (2015). A flash heating method for measuring thermal conductivity at high pressure and temperature: Application to Pt. *Physics of the Earth and Planetary Interiors*, 247, 17–26. <https://doi.org/10.1016/j.pepi.2015.06.002>
- Meco, H., & Napolitano, R. (2005). Liquidus and solidus boundaries in the vicinity of order-disorder transitions in the Fe-Si system. *Scripta Materialia*, 52(3), 221–226. <https://doi.org/10.1016/j.scriptamat.2004.09.026>
- Montoya, J. A., & Goncharov, A. F. (2012). 06. Finite element calculations of the time dependent thermal fluxes in the laser-heated diamond anvil cell. *Journal of Applied Physics*, 111(11), 112617. <https://doi.org/10.1063/1.4726231>
- Morard, G., Boccato, S., Rosa, A. D., Anzellini, S., Miozzi, F., Henry, L., et al. (2018). Solving controversies on the iron phase diagram under high pressure. *Geophysical Research Letters*, 45(20), 11074–11082. <https://doi.org/10.1029/2018gl079950>
- Morard, G., Bouchet, J., Rivoldini, A., Antonangeli, D., Roberge, M., Boulard, E., et al. (2018). Liquid properties in the Fe-FeS system under moderate pressure: Tool box to model small planetary cores. *American Mineralogist*, 103(11), 1770–1779. <https://doi.org/10.2138/am-2018-6405>
- Morard, G., & Katsura, T. (2010). Pressure–temperature cartography of Fe–S–Si immiscible system. *Geochimica et Cosmochimica Acta*, 74(12), 3659–3667. <https://doi.org/10.1016/j.gca.2010.03.025>
- Namur, O., Charlier, B., Holtz, F., Cartier, C., & McCammon, C. (2016). Sulfur solubility in reduced mafic silicate melts: Implications for the speciation and distribution of sulfur on Mercury. *Earth and Planetary Science Letters*, 448, 102–114. <https://doi.org/10.1016/j.epsl.2016.05.024>
- Nittler, L. R., Chabot, N. L., Grove, T. L., & Peplowski, P. P. (2018). *Mercury: The view after messenger*. In S. C. Solomon, L. R. Nittler, & B. J. Anderson (pp. 30–51). Cambridge University Press.
- Ohta, K., Kuwayama, Y., Hirose, K., Shimizu, K., & Ohishi, Y. (2016). Experimental determination of the electrical resistivity of iron at earth's core conditions. *Nature*, 534(7605), 95–98. <https://doi.org/10.1038/nature17957>
- Peplowski, P. N., Lawrence, D. J., Rhodes, E. A., Sprague, A. L., McCoy, T. J., Denevi, B. W., et al. (2012). Variations in the abundances of potassium and thorium on the surface of Mercury: Results from the messenger gamma-ray spectrometer. *Journal of Geophysical Research*, 117(E12). <https://doi.org/10.1029/2012je004141>
- Pommier, A. (2018). Influence of sulfur on the electrical resistivity of a crystallizing core in small terrestrial bodies. *Earth and Planetary Science Letters*, 496, 37–46. <https://doi.org/10.1016/j.epsl.2018.05.032>
- Pommier, A., Tauber, M. J., Renggli, C., Davies, C., & Wilson, A. (2025). Electrical properties of alkaline earth sulfides and implications for the interior of mercury. *Journal of Geophysical Research: Planets*, 130(2), e2024JE008651. <https://doi.org/10.1029/2024JE008651>
- Prescher, C., Dubrovinsky, L., Bykova, E., Kuzenko, I., Glazyrin, K., Kantor, A., et al. (2015). High poisson's ratio of earth's inner core explained by carbon alloying. *Nature Geoscience*, 8(3), 220–223. <https://doi.org/10.1038/ngeo2370>
- Rivoldini, A., & Van Hoolst, T. (2013). The interior structure of Mercury constrained by the low-degree gravity field and the rotation of mercury. *Earth and Planetary Science Letters*, 377–378, 62–72. <https://doi.org/10.1016/j.epsl.2013.07.021>
- Secco, R. A. (2017). Thermal conductivity and seebeck coefficient of Fe and Fe-Si alloys: Implications for variable Lorenz number. *Physics of the Earth and Planetary Interiors*, 265, 23–34. <https://doi.org/10.1016/j.pepi.2017.01.005>
- Shen, G., Wang, Y., Dewaele, A., Wu, C., Fratauduno, D. E., Eggert, J., et al. (2020). Toward an international practical pressure scale: A proposal for an ipps ruby gauge (Ipps-ruby2020). *High Pressure Research*, 40(3), 299–314. <https://doi.org/10.1080/08957959.2020.1791107>
- Silber, R. E., Secco, R. A., Yong, W., & Littleton, J. A. H. (2018). Electrical resistivity of liquid Fe to 12 GPa: Implications for heat flow in cores of terrestrial bodies. *Scientific Reports*, 8(1), 1–9. <https://doi.org/10.1038/s41598-018-28921-w>
- Stähler, S. C., Khan, A., Banerdt, W. B., Lognonné, P., Giardini, D., Ceylan, S., et al. (2021). Seismic detection of the martian core. *Science*, 373(6553), 443–448. <https://doi.org/10.1126/science.abi7730>
- Steinbrügge, G., Dumberry, M., Rivoldini, A., Schubert, G., Cao, H., Schroeder, D. M., & Soderlund, K. M. (2021). Challenges on Mercury's interior structure posed by the new measurements of its obliquity and tides. *Geophysical Research Letters*, 48(3), e2020GL089895. <https://doi.org/10.1029/2020GL089895>
- Stixrude, L., & Lithgow-Bertelloni, C. (2024). Thermodynamics of mantle minerals – III: The role of iron. *Geophysical Journal International*, 237(3), 1699–1733. <https://doi.org/10.1093/gji/ggae126>
- Takahashi, F., Shimizu, H., & Tsunakawa, H. (2019). Mercury's anomalous magnetic field caused by a symmetry-breaking self-regulating dynamo. *Nature Communications*, 10(1), 208. <https://doi.org/10.1038/s41467-018-08213-7>
- Tao, R., & Fei, Y. (2021). High-pressure experimental constraints of partitioning behavior of Si and S at the Mercury's inner core boundary. *Earth and Planetary Science Letters*, 562, 116849. <https://doi.org/10.1016/j.epsl.2021.116849>
- Terasaki, H., Rivoldini, A., Shimoyama, Y., Nishida, K., Urakawa, S., Maki, M., et al. (2019). Pressure and composition effects on sound velocity and density of core-forming liquids: Implication to core compositions of terrestrial planets. *Journal of Geophysical Research: Planets*, 124(8), 2272–2293. <https://doi.org/10.1029/2019JE005936>

- Tosi, N., Grott, M., Plesa, A.-C., & Breuer, D. (2013). Thermochemical evolution of Mercury's interior. *Journal of Geophysical Research: Planets*, 118(12), 2474–2487. <https://doi.org/10.1002/jgre.20168>
- Tsujino, N., Nishihara, Y., Nakajima, Y., Takahashi, E., ichi Funakoshi, K., & Higo, Y. (2013). Equation of state of γ -Fe: Reference density for planetary cores. *Earth and Planetary Science Letters*, 375, 244–253. <https://doi.org/10.1016/j.epsl.2013.05.040>
- Václav, P., Michal, D., & Lukáš, P. (2014). Crystallographic computing system jana2006. *General features*, 229.
- Vander Kaaden, K. E., & McCubbin, F. M. (2015). Exotic crust formation on Mercury: Consequences of a shallow, FeO-poor mantle. *Journal of Geophysical Research: Planets*, 120(2), 195–209. <https://doi.org/10.1002/2014je004733>
- Vander Kaaden, K. E., McCubbin, F. M., Turner, A. A., & Ross, D. K. (2020). Constraints on the abundances of carbon and silicon in Mercury's core from experiments in the Fe-Si-C system. *Journal of Geophysical Research: Planets*, 125(5), e2019JE006239. <https://doi.org/10.1029/2019JE006239>
- Wagle, F., Steinle-Neumann, G., & de Koker, N. (2019). Resistivity saturation in liquid iron–light-element alloys at conditions of planetary cores from first principles computations. *Comptes Rendus Geoscience*, 351(2–3), 154–162. <https://doi.org/10.1016/j.crte.2018.05.002>
- Yagi, T., Ohta, K., Kobayashi, K., Taketoshi, N., Hirose, K., & Baba, T. (2010). Thermal diffusivity measurement in a diamond anvil cell using a light pulse thermoreflectance technique. *Measurement Science and Technology*, 22(2), 024011. <https://doi.org/10.1088/0957-0233/22/2/024011>
- Yong, W., Secco, R. A., Littleton, J. A. H., & Silber, R. E. (2019). The iron invariance: Implications for thermal convection in Earth's core. *Geophysical Research Letters*, 46(20), 11065–11070. <https://doi.org/10.1029/2019gl084485>
- Zhang, Y., Luo, K., Hou, M., Driscoll, P., Salke, N. P., Minar, J., et al. (2022). Thermal conductivity of Fe-Si alloys and thermal stratification in Earth's core. *Proceedings of the National Academy of Sciences*, 119(1), e2119001119. <https://doi.org/10.1073/pnas.2119001119>
- Zhang, Y., Yoshino, T., Yoneda, A., & Osako, M. (2019). Effect of iron content on thermal conductivity of olivine with implications for cooling history of rocky planets. *Earth and Planetary Science Letters*, 519, 109–119. <https://doi.org/10.1016/j.epsl.2019.04.048>

References From the Supporting Information

- Andraut, D., Pison, L., Morard, G., Garbarino, G., Mezouar, M., Bouhifd, M. A., & Kawamoto, T. (2022). Comment on: Melting behavior of SiO₂ up to 120 GPa (Andraut et al. 2020). *Physics and Chemistry of Minerals*, 49(2), 3. <https://doi.org/10.1007/s00269-021-01174-2>
- Beck, P., Goncharov, A. F., Struzhkin, V. V., Militzer, B., Mao, H.-k., & Hemley, R. J. (2007). Measurement of thermal diffusivity at high pressure using a transient heating technique. *Applied Physics Letters*, 91(18), 181914. <https://doi.org/10.1063/1.2799243>
- Berrada, M., Secco, R. A., Yong, W., & Littleton, J. A. H. (2020). Electrical resistivity measurements of Fe-Si with implications for the early lunar dynamo. *Journal of Geophysical Research: Planets*, 125(7), e2020JE006380. <https://doi.org/10.1029/2020JE006380>
- Dewaele, A., Loubeyre, P., Occelli, F., Mezouar, M., Dorogokupets, P. I., & Torrent, M. (2006). Quasihydrostatic equation of state of iron above 2 Mbar. *Physical Review Letters*, 97(21), 215504. <https://doi.org/10.1103/PhysRevLett.97.215504>
- Giampaoli, R., Kantor, I., Mezouar, M., Boccato, S., Rosa, A. D., Torchio, R., et al. (2018). Measurement of temperature in the laser heated diamond anvil cell: Comparison between reflective and refractive optics. *High Pressure Research*, 38(3), 250–269. <https://doi.org/10.1080/08957959.2018.1480017>
- Hsieh, W.-P., Lin, C.-H., Chen, C.-C., & Chang, J.-W. (2024). Testing the validity of the Wiedemann–Franz law for metals and alloys at high pressures. *Applied Physics Letters*, 125(25), 252201. <https://doi.org/10.1063/5.0234126>
- McGuire, C., Sawchuk, K., & Kavner, A. (2018). Measurements of thermal conductivity across the B1–B2 phase transition in NaCl. *Journal of Applied Physics*, 124(11), 115902. <https://doi.org/10.1063/1.5042407>
- Pourovskii, L. V., Mravlje, J., Georges, A., Simak, S. I., & Abrikosov, I. A. (2017). Electron–electron scattering and thermal conductivity of e-iron at Earth's core conditions. *New Journal of Physics*, 19(7), 073022. <https://doi.org/10.1088/1367-2630/aa76c9>
- Pozzo, M., Davies, C. J., & Alfè, D. (2022). Towards reconciling experimental and computational determinations of Earth's core thermal conductivity. *Earth and Planetary Science Letters*, 584, 117466. <https://doi.org/10.1016/j.epsl.2022.117466>
- Silber, R. E., Secco, R. A., Yong, W., & Littleton, J. A. H. (2019). Heat flow in earth's core from invariant electrical resistivity of Fe-Si on the melting boundary to 9 GPa: Do light elements matter? *Journal of Geophysical Research*, 124(6), 5521–5543. <https://doi.org/10.1029/2019JB017375>
- Tateno, S., Komabayashi, T., Hirose, K., Hirao, N., & Ohishi, Y. (2019). Static compression of B2 KCl to 230 GPa and its P–V–T equation of state. *American Mineralogist*, 104(5), 718–723. <https://doi.org/10.2138/am-2019-6779>
- Walker, D., Cranswick, L. M., Verma, P. K., Clark, S. M., & Buhre, S. (2002). Thermal equations of state for B1 and B2 KCl. *American Mineralogist*, 87(7), 805–812. <https://doi.org/10.2138/am-2002-0701>
- Yin, Y., Liu, C., Zhai, S., Liu, Y., & Fei, Y. (2025). Lattice thermal conductivity of hcp Fe-Si alloys determined by machine learning potentials. *Geophysical Research Letters*, 52(17), e2024GL111953. <https://doi.org/10.1029/2024gl111953>